UNITED STATES DEPARTMENT OF THE INTERIOR

GEOLOGICAL SURVEY

GEOLOGY OF THE DEEP CREEK AREA, WASHINGTON, AND ITS REGIONAL SIGNIFICANCE

By

Robert G. Yates

Open-file report --76-537

1976

This report is preliminary and has not been edited or reviewed for conformity with Geological Survey standards.

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1. The readers attention is called to the fact that the table of contents does not show proper rank of many headings.

2. Page 13, line 18 and 19: "the connotation of" should be followed by "a clay content refers to the original sediment. Collectively rocks" of this group ....

3. Page 27, line 13: Delete sentence "The ..... appendix."

4. Page 43, heading: "Rocks below the thrust fault" and "Gypsy quartzite" are main and subheadings rather than a single heading.

5. Page 45, heading: "Rocks above the thrust fault" and "Maitlen Phyllite" are main and subheadings rather than a single heading.

6. Page 117, line 17: "bent" omitted before "albite lamellae."

7. Page 160: Delete 1st paragraph as a repeat of paragraph on previous page.

8. Page 239, line 17: change ", and those" to "were."

9. Page 295: Add heading above 3rd paragraph and in table of contents: "The Lime Creek Mountain block and the decollément problem."

10. Page 302, Next to last sentence: Should read <u>un</u>reasonable rather than reasonable.

11. Page 402, line 6: add "can be assumed to have been" continuous.

12. Page 404, line 8: Should read "p. 83), of Kootenay Lake (Crosby, 1968, p. 72), and of the panhandle of Idaho. It requires no"

13. Page 422, last line: change "its" to "ifs."

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Foreword

This report is divided into three parts. The first part describes the stratigraphy and structure of the Deep Creek area of northern Stevens County, Washington, the second part describes the regional setting, and the third part interprets the events and processes that produced the rocks and structures described in the first two parts. The first part is straight forward description, the second is both description and correlation, but includes as well, appreciable --and unavoidable -- interpretation. The third part, being strictly interprative, introduces no new descriptions except those necessary to develop concepts.

The description of the stratigraphy and structure of the Deep Creek area is arranged in an unorthodox manner. Because the area consists of several tectonic blocks of contrasting stratigraphy and structure, each block is described as a geographic unit -- much as quadrangles in a block that are published in sequence. As a result, descriptions of stratigraphic units common to two or more blocks are scattered under several headings. The reader only interested in a broad knowledge of a particular formation accordingly is inconvenienced, but not deprived.

The second part, the regional setting, includes previously unpublished descriptions of the Northport quadrangle and incorporates knowledge gained by reconnaissance study in the Colville and Metaline quadrangles. Most information, however, comes from published reports and maps, particularly that on the geology of British Columbia, which comes solely from this source. As a result, widely scattered data is assembled and integrated to a condensed description of a poorly understood region. Whenever brevity permits, original descriptions are quoted and the development of ideas is traced.

The third part of the report, in summarizing the historical geology of the region, digresses into the problems of genesis that obsess all areas of complex geology. The problems are explored and explanations suggested. To come to grips with many problems, assumptions had to be made and speculations advanced, but nowhere should the reader confuse fact with speculation.

Abstract

This report, although primarily concerned with the stratigraphy and structure of a lead-zinc mining district in northern Stevens County, Washington, discusses and integrates the geology of the region about the Deep Creek area. Although the study centers in an area of about 200 square miles immediately south of the International Boundary, the regional background comes from: (1) the previously undescribed Northport quadrangle to the west, (2) published reports and reconnaissance of the Metaline quadrangle to the east, and (3) from published reports and maps of a 16 mile wide area that lies to the north adjacent to these three quadrangles in British Columbia. The report is divided into three parts: (1) descriptions of rocks and structures of the Deep Creek area, (2) descriptions of the regional setting of the Deep Creek area, and (3) an analysis and interpretation of the depositional and tectonic events that produced the geologic features exposed today.

In the Deep Creek area surficial deposits of sand and gravel of glacial origin cover much of the consolidated rocks, which range in age from greenschist of the late Precambrian to albite granite of the Eocene. Three broad divisions of depositional history are represented: (1) Precambrian, (2) lower Paleozoic and (3) upper Paleozoic; the record of the Mesozoic and Eocene is fragmentary. The lower Paleozoic division is the only fossil-controlled sequence; the age of the other two divisions were established by less direct methods. Both Precambrian and upper Paleozoic sequences are dominated by fine-grained detrital sediments, the Precambrian tending towards the alumina-rich and the upper Paleozoic tending towards the black shale facies with high silica. Neither sequence has more than trivial amounts of coarse clastics. Both include limestones, but in minor abundance.

The lower Paleozoic sequence, on the other hand, represents a progressive change in deposition. The sequence began during the very late Precambrian with the deposition of clean quartz sand. This was followed by the accumulation of a comparatively thin limestone unit succeeded by a thick shale. The shale grades into a thick carbonate unit which in turn is overlain by black graptolitic slates (Ordovician). This general order of deposition holds for the Cambro-Ordovician throughout the area.

Precambrian rocks indigenous to the Deep Creek area, have undergone at least six tectonic events of greatly different intensities. The first three of these events are epeirogentic, the fourth involves intense folding, the fifth, crossfolding, and the sixth, block faulting without folding. These events are dated with varying degrees of precision. The two epeirogentic events of the Precambrian, one gentle folding at the beginning of Windermere time and the other high angle faulting and volcanism in mid-Windermere time, did little to deform or metamorphose the rocks. The third event consists of uplift of northern Idaho and adjacent Montana and westward décollement thrusting of essentially unfolded lower Paleozoic rocks. The décollement faulting is inferred to explain anomalous rock distribution and cannot be accurately dated. It occurred sometime after the Devonian and before the Jurassic. A late Paleozoic age is favored.

The fourth event, the principal period of folding also cannot be dated with precision. The northeasterly trending folds produced by this event dominate the structure of the Deep Creek area; only locally are these folds destroyed by later deformations. To the northeast in British Columbia, rocks as young as Middle Jurassic are folded along northeast axes; however, the folding recorded in Jurassic rocks may not necessarily represent the total length of time that northeast trending folds were developing in the underlying Paleozoic rocks. In the Deep Creek area, the 100 million year old Spirit pluton clearly crosscuts the folds and thus puts an upper limit on their age. A mid- to late Jurassic age is believed most probable but an earlier Late Triassic folding along similar axes is conceivable. It would be convenient if the folds could be called Jurassic or Nevadan, but until their age is more precisely established the term "Arc folds," is used in reference to their importance as the defining structure of the Kootenay arc, the northeasterly trending segment of the Cordilleran fold-belt, in which the Deep Creek area is located.

The deformation in the Kootenay arc is polyphase; the fifth deformation, the crossfold event is superimposed on the northeast folds. The northeast (arc) folds, when subjected to northeast-southwest compression, were buckled into kinkfolds and broken by south and north dipping thrust faults, both of which are of near east-west trend. These buckles and thrusts were contemporaneously torn by "tears" that broke across the beds and through the thrust faults allowing some segments to move further northward than other segments. The tear faults, most of N 10° E trend, may in part be faults that extend into the crystalline basement rocks, because the last event, that of block faulting, was dominated by high angle faults of similar trend, but without measurable horizontal movement. The last event high angle, block faulting, which occurred during the Eocene was accompanied by, and followed by, the extrusion of volcanic rocks.

The second part of the report, which describes the regional geology, emphasizes the contrasting geologic provinces that neighbor the Deep Creek area and the enclosing Kootenay arc. Most of the area described in this part is in the Kootenay arc, but rocks of the Belt-Purcell anticlinorium, which lies east of the arc, and those in the eastern fringe of the Pacific Borderland, which lies west of the arc, are included. The tectonic behavior of the three provinces created environments that favored the accumulation of distinctive rock assemblages. The Belt-Purcell rocks of the anticlinorium, dominantly fine- to medium-grained clastics, accumulated on the craton that consisted of a crystalline basement complex whose internal tectonic activity was limited to downwarping and block faulting during the 600 million years it took the Belt-Purcell sediments to accumulate. The Kootenay arc, basically a foldbelt, represents a zone of change from this stable, continental basement environment to the unstable environment at the oceanic border. It lies along the hinge zone where the sediments of the craton thicken to a miogeosynclinal prism. The Pacific Borderland consists of the Cordilleran eugeosyncline and its volcanic-bearing deposits. It represents the crustal unrest typical of this environment. Its thick covering of sedimentary and volcanic rocks are believed to rest upon an oceanic basement.

The Belt-Purcell anticlinorium consists of from 14 to 23 kilometres of Precambrian rocks of the Belt Supergroup. These rocks extend into the Kootenay arc where they are associated with a stratigraphic variant, the Priest River Group, which is unconformably overlain by 4.5 kilometres of rocks of the Windermere Series of Canada, which includes the Shedroof Conglomerate, Leola volcanics, and Monk Formation. The Windermere rocks contrast with the Belt rocks by representing an environment of tectonism rather than one dominated by crustal stability. The Monk Formation is overlain by the thick quartz sand facies of the Lower Cambrian, followed by a mud facies, and a carbonate facies of the middle and upper Cambrian, all typical miogeosynclinal deposits. Abruptly overlying the carbonate deposits are black slates of the lower and middle Ordovician, which extend upward in time through the Silurian and Devonian to be lost in the time scale in thousands of metres of predominantly gray to black unfossiliferous pelites. The post-Devonian part of this black shale facies, the transitional assemblage, gradually changes in character both upward and westward through the addition of mafic volcanic rocks. The youngest post-Devonian rocks, the Pennsylvanian Mount Roberts Formation, is typically eugeosynclinal with mafic volcanic rocks and graywackes. Although the evidence is not positive, the rocks are believed to change across the arc, in time and space, from miogeosynclinal to eugeosynclinal.

The area defined by the regional map is divided into four northeasterly trending structural belts, which from southeast to northwest are: The homocline belt, fold belt, thrust belt, and Jurassic volcanic belt. The homocline belt bridges the Belt-Purcell anticlinorium and the Kootenay arc; and northeast trending fold and thrust belts fall within the Kootenay arc; and the Jurassic volcanic belt, which has arc structures in its eastern part, corresponds to the eugeosynclinal province of the Pacific Borderland. Although the belts are bounded by faults, the characterizing features of the belts overlap. The intensity of deformation increases from southeast to northwest, reaching a peak in the thrust belt.

The homocline belt is a northeasterly striking northwestward dipping sequence of Precambrian and Cambrian rocks. The regularity of the homocline is interrupted by the high angle Flume Creek fault, which segments it, by northwest trending cross faults, and by the Brodie-Sullivan kink fold, a N 80° W cross buckle, and by a northwest dipping strike fault. The fold belt, which is the heart of the Kootenay arc, is divided by transverse faults into four tectonic blocks. In British Columbia the fold belt consists of multiple folds that coalesce at the International Boundary into a single anticline and flanking syncline. The anticline dominates the structure of three blocks; the fourth block is the overturned limb of a fold that lies northwest of the anticline. Prefold thrust faults confuse the internal structure of the blocks, as well as relations between blocks.

The thrust belt is a modified part of the fold belt: it consists of an anticlinal fold, the Columbia anticline that is badly broken along numerous thrust faults. The thrust faults strike nearly east-west at a low angle to the fold, and are the product of a later deformation. The core of the anticline is composed of the Cambrian miogeosynclinal rocks and the outer envelope, the transitional assemblage, is largely black pelite.

The Jurassic volcanic belt, structurally the least understood belt, is dominated by the Jurassic volcanic rocks of the Rossland Formation. Folds in this belt are generally open and trend northwesterly except near the edge of the arc where they bend into arc trends. The high angle faults are of diverse trends.

The third section of the report, "Sequence of events and interpretation of regional geology" reviews the geologic events of the region and interprets the processes that produced them. In order to establish the location and form of the continental margin during the Precambrian and Paleozoic, the Belt-Purcell terrane is analyzed by assuming that the depositional characteristics of the rocks, along with the thickness, deformation, and metamorphism are indicators of a stable basement. Unconformities and provenances of the Windermere are discussed, as is the possibility of a Windermere rifting, which is regarded as untenable. Evidence is presented that faults similar in trend and probably movement pattern to that of the Osburn and Hope faults of Idaho, are late Precambrian in age.

The miogeosynclinal deposits of the Cambrian are interpreted as forming on the continental slope. Provenance, environment of accumulation, and diachrony of the sand facies are discussed. Regional correlations of both sand and mud facies are attempted and the case for transition between the miogeosynclinal and eugeosynclinal suites of the late Paleozoic is presented. The conglomerates and problems of a Mississippian orogeny are discussed and evaluated. No evidence was found to support a mid-Paleozoic orogeny, although an epeirogenic event was considered likely. No great facies-dividing faults are evident, but infrafacies gravity slides from an eastern high are postulated.

The late Paleozoic is recorded in eugeosynclinal rocks deposited on an oceanic crust. The evolution of the high grade metamorphic rocks of the Shuswap complex, which is surrounded by the eugeosynclinal rocks, creates special problems that are outlined and speculatively interpreted. A stable join between continental and oceanic crusts during this period is envisioned. In contrast, the first orogeny that of the Mesozoic, represents the breaking of the join and the crumpling of the continental margin. Multiphase folding occurred between Late Triassic and mid-Cretaceous, in two episodes. As a result of this folding, the ancestral Kootenay arc, a depositional basin, became a fold belt.

An attempt is made to fit the Mesozoic history to a plate tectonics model. Conventional models that relate granitic rocks to plate subduction do not apply because the plutons are not systemically distributed by age or composition. The model proposed relates the high grade metamorphic rock of the Shuswap complex and associated granites to a high heat rise that extended at least to the western part of the Belt-Purcell anticlinorium. This created a zone of partial melting upon which rested the basaltic rocks of the oceanic crust and the eugeosynclinal rocks deposited thereon, and the miogeosynclinal and crystalline rocks of the continent. It is speculated that the anomalous heat rise was the result of the subduction of a western, oceanic plate, which partitioned a heat cell thus increasing the thermal gradient beneath the continental margin. As the two plates moved toward each other, the sedimentary prism that had existed across the formerly stable join between the two plates was detached along décollement faults and crumpled into folds and thrusts.

Because of the widespread distribution of Eocene granitic rocks across eastern Washington, Idaho, and western Montana, it is assumed they also are related to a heat rise that produced a zone of flow within the continental crust, and it is further assumed that arching of the crust created a slope down which the rigid crust above the zone of flow could move eastward, to be ultimately expressed at the surface as Eocene movement on the thrusts of the Rocky Mountain front. The tectonic gap created by this down slope movement is believed to be compensated by extension on listric faults.

Stratigraphy and structure of the Deep Creek area

Introduction

The Deep Creek area, which lies adjacent to the International Boundary in Stevens County, Washington (fig. 1), includes much of what is commonly considered the Northport Mining district, a moderate producer of lead-zinc ores. The name, Deep Creek area, is a title of convenience, a name selected to avoid the awkwardness of repeated reference to the "area of Boundary, Leadpoint, Spirit and Deep Lake quadrangles," which although precise, is undeniably cumbersome. Although these four 7 1/2-minute quadrangles include a somewhat larger area than the drainage basin of Deep Creek, the convenience far outweighs the inaccuracy. The northwestern of the four quadrangles (Boundary quadrangle) is crossed by the broad terraced valley of the Columbia River, which contrasts with the steep sloped, heavily forested topography of the Deep Creek drainage basin.

Interest by the mining industry in the lead-zinc deposits is responsible for selecting the area for an intensive field investigation, which began in midsummer 1955. During the progress of the work the project was shifted from the Mineral Deposits Branch of the U.S. Geological Survey to a newly created Branch of Regional Geology. Accordingly, the emphasis shifted from a study of the mineralization to a study of the geologic events that created the environment for mineralization. It is believed that the redirection has yielded a product of value to the mining industry.

Figure 1.--Index map showing location of the Deep Creek area, area covered by map of regional geology (~~Plate~~ Fig 14), and quadrangle and map areas referred to in text.

U. S. GOVERNMENT PRINTING OFFICE : 1959 O - 511171

This is by no means the first geologic study of the area. Shortly after the discovery and early development of the lead-zinc deposits, about the turn of the century, technical papers appeared in the mining journals and in publications of State and Federal agencies. These accounts, from 1897 to 1921, were primarily on the ores; regional geology and the problems of stratigraphy and structure were incidental. The early surveys of the geology along the International Boundary likewise contributed little to a knowledge of the geology of the Deep Creek area. No United States survey ever crossed the area, the Canadian Survey of 1859-61 (Bauerman, H., 1885) detoured far to the south to avoid "this most inaccessible part," and the Survey of R. A. Daly (1912), because of its reconnaissance nature, was forced to lump the rocks into gargatuan units, far too ponderous to reveal the complexities of structure.

It was not until 1920 that the area ~~s~~ was represented on a geologic map. At this time the Washington Geological Survey published the report of Charles E. Weaver (1920) on the "Mineral Resources and Geology of Stevens County." Although Weaver's report emphasized the mineral deposits, he described the rocks, arranged them in what he believed to be a sequential manner, and plotted their distribution as formational units on a highly inadequate base map. Although Weaver's map and rock correlations have not--as he anticipated--held up under the scrutiny of time and more detailed mapping, his contribution should not be minimized; it is indeed an accomplishment to traverse 2400 square miles of highly diverse terrane on foot and by horse and wagon in the span of three short months and come up with a map that shows the variety of lithology that his does--est modus in rebus.

This map by Weaver was the only one portraying the geology in and about the vicinity of the Deep Creek area until the U.S. Geological Survey published the map by C. F. Park, Jr. and R. S. Cannon, Jr. (1943) of the Metaline quadrangle, which lies immediately to the east. In the interim, however, work was going on to the north in the Salmo Map area of British Columbia by J. F. Walker (1934) of the Geological Survey of Canada. The stratigraphic succession mapped by Walker, which was also that of Daly, was to their knowledge devoid of fossils, consequently it is not surprising that they considered all units Precambrian in age. Although fossils had been found far to the south in the Chewelah quadrangle as early as 1928 (Branson, 1931), these were from rocks neither structurally nor stratigraphically linked to anything along or north of the International Boundary. The importance of Park and Cannon's work in the Metaline quadrangle is that they described and mapped a sequence of rocks that contained fossils representing Cambrian, Ordovician, and Devonian time. For the first time, the stratigraphy of the region was on a solid foundation.

The remapping of the Salmo Map area by H. W. Little (1950) brought to light other fossils and a consequent reinterpretation of both stratigraphy and structure accompanied by a greater appreciation of the complexity of the geology of the region. Part of this area, the belt known as the Salmo lead-zinc area, was restudied by J. T. Fyles and C. G. Hewlett (1959) of the British Columbia Department of Mines and Petroleum Resources. Mapping at a larger scale, which permitted considerable detail, led them to introduce the concept of isoclinal folding and thrust faulting as a process of fundamental importance in the structural evolution of the area.

About the same time, the Metalinemining district was being restudied by McClelland G. Dings and Donald H. Whitebread of the U.S. Geological Survey (Dings and Whitebread, 1965). They studied that part of the Metaline quadrangle that contains the Cambrian carbonate rocks, the host to the lead-zinc deposits. They subdivided both Cambrian and Ordovician rocks and separated diagenetic dolomite from hydrothermal dolomite

The above sketch of past geological investigations in and about the Deep Creek area is far from complete, the contributions of some go unmentioned, the contributions of others go unemphasized. As for example, H. W. Little's report and map of the "Nelson Map-area, west half, British Columbia (1960), furnished the writer a broad understanding of the geology of the area immediately north. This report and others, as well as those mentioned above, are freely drawn from in the pages that follow. The contributions of two geologists, W. A. G. Bennett and Charles D. Campbell, are more properly listed under the acknowledgements.

Field work and acknowledgements

The field study of the Deep Creek area began in August of 1955 and was completed in 1960. To better understand the geology of the Deep Creek area, the project was expanded into adjacent quadrangles. As a result the Northport quadrangle to the west was mapped (Yates, 1971) and the geology of parts of the Metaline quadrangle was reexamined. Reconnaissance to the south in the Colville and Chewelah quadrangles added breadth to the knowledge of the stratigraphy and structures.

The geologic mapping and fact gathering for this report was participated in by Jaques F. Robertson, David Haines, Arthur B. Ford, Howard Albee, Philip M. Blacet, Gilbert W. Franz, David W. Baker, Fred K. Miller, B. Vaughn Marshall, and James B. Pinkerton. The contributions of all of these geologists is appreciated.

Information and wholehearted cooperation was given by Eskil Anderson of the Grandview Mining Company and T. S. Higginbotham and Albert V. Quine of the Goldfield Consolidated Company. Many others engaged in mining as well as numerous local residents did much to aid the work. The U.S. National Park Service, represented at Kettle Falls by Jerry D. Lee, kindly supplied accommodations for a trailer during some of the field work.

During the war years between 1942 and 1944, Charles D. Campbell spent considerable time in the area making examinations of the lead-zinc mines as part of the U.S. Geological Survey's Strategic Mineral Program. Campbell's work was not all routine mine mapping; somehow he found time to study and map the areas between mines, with the result that he was the first to obtain a broad picture of the geology of the district. He recognized stratigraphic units as counterparts of those in the Metaline district and recognized that the northeast trending folds had southeast dipping axial planes and associated reverse faults (Campbell, C. D., 1947). His mine descriptions, published as Open File reports, add pertinent facts to critical areas of interpretation.

W. A. G. Bennett of the Washington Division of Mines and Geology also requires special recognition. Bennett's knowledge extends over a wide area in northeastern Washington and over a time span of more than thirty years. Although most of his work has been south of the Deep Creek area (Bennett, 1944), his broad knowledge of regional problems has on many occasions been generously shared with the writer.

I cannot forget the warm and friendly interchange of information with my colleagues in Canada. I am indebted to Heward W. Little of the Geological Survey of Canada and to James T. Fyles of the British Columbia Department of Mines and Petroleum Resources who made it possible through

Figure 2. Geologic map and sections of the Deep Creek area, Stevens and Pend Oreille Counties, Washington.

their publications and discussions to relate the geology of the Deep Creek area to the geology of adjacent British Columbia.

My colleagues in the Geological Survey also deserve recognition, particularly McClelland G. Dings, who introduced me to the stratigraphy of the lower Paleozoic of Pend Oreille County, Washington and all those who took part in the field work, H. F. Albee, D. W. Baker, P. M. Blacet, A. B. Ford, David Haines, B. V. Marshall, F. K. Miller, P. B. Pinkerton and J. E. Robertson. Preston E. Hotz and Fred K. Miller, who struggled through the manuscript as technical reviewers are in my debt.

General relations

The geology of the Deep Creek area could be summarized by representing it as a thick accumulation of marine sedimentary rocks ranging from Precambrian to Jurassic in age that are folded, faulted, and intruded and metamorphosed by granitic rocks of Cretaceous and Eocene age. Such a summary adheres to the observable facts but unfortunately oversimplifies the geologic history so that it appears incompatible with the incongruities of rock distribution shown on the geologic map (fig. 2). These incongruities are evidence of a complex history, a history interpreted as requiring the westward thrusting of segments of the stratified crust into juxtaposition against segments of unlike crust, a history whose interpretation is confused by later folding of the thrust plates. The *a priori* introduction of these postulates is perhaps justified, but their development must be postponed until the rocks and structures are described and the problems become self-evident. Before proceeding to the descriptive phase of this report, a brief discussion of what lies behind the uncommon grouping of the stratigraphic units in the Explanation for figure 2 is in order.

The geologic map of the Deep Creek area is divided into seven sub-areas (see index map on figure 2). One sub-area is occupied by the Spirit pluton, a granodiorite mass that intrudes the pre-Tertiary rocks of adjacent areas. The sub-areas north of the Spirit pluton are designated "blocks" and identified by the names of prominent geographic features. The term, block, was used by Park and Cannon (1943, p. 28) as an aid in describing the geology of areas in the Metaline quadrangle that have either structural or stratigraphic unity--or both. Although the boundaries between blocks are faults, the blocks are not fault blocks in the sense of basin and range structure, where the blocks are expressed topographically and are bounded by high angle normal faults; instead bounding faults include both low angle thrust or reverse faults and high angle faults. There is almost no topographic distinction between blocks. The area south of the Spirit pluton is composite in character and includes part of three blocks that extended far south of the Deep Creek area.

The stratigraphic sequence in one block may represent a different span of time than that in an adjacent block or it may represent the same or an overlapping time span, with age equivalent units being lithologically different. Although these differences in lithology are not extreme, they are consistent; lithology and lithologic variation within a block is constant throughout the block; gradations towards lithologies of neighboring blocks was not detected.

The blocks as structural units have gross major structures that contrast with those in neighboring blocks. As for example, beds in one block may be right side up, whereas those in the adjacent block may be overturned. A block may be internally complicated, by including two thrust plates.

Considerable effort has been made to keep interpretation on the geologic map (fig. 2) to a minimum; in contrast, the geologic cross-sections are highly interpretative. The geologic relations shown more than a few hundred feet below and above the surface are diagrammatic representations of structural concepts that come solely from surface observations.

The geologic map has certain built-in limitations. It represents the geology of an area not only heavily forested but more than half covered by glacial drift. Many stratigraphic units are gross, measured in hundreds of metres; others are monotonous sequences of argillite with almost no marker beds; and only a few have sedimentary features that indicate the top side of beds. Fossils are rare. Gradational boundaries are common. The background metamorphism is low grade (muscovite-chlorite) in most of the area but increases to medium high grade (pyroxene hornfels) in the innermost part of the aureole about the Spirit pluton.

Stratigraphy

Introductory statement

The stratigraphic and structural heterogeneity became evident during early stages of the mapping. As the work progressed, it became increasingly obvious that all correlations should be made with great caution and that a mappable lithologic unit should not be forced into a formational pigeon hole until all observations were made. The earliest product of the project, the preliminary map of the Leadpoint quadrangle (Yates and Robertson, 1958) followed this concept by excluding numerous rock units from formal sequential arrangement. The present product (fig. 2), which includes the area of the Leadpoint quadrangle, considerably reduces the number of rock units shown on the preliminary map through correlations and common map symbols, but it retains much of my early reluctance to rigidly formalize the stratigraphy. It was decided that formal stratigraphic names should not be introduced until nearby areas with common stratigraphic problems are mapped.

Although new formational names are not introduced in this report, no major rock unit is without a "handle." Established formational names are extended wherever there is no question of the correlation, from type localities in British Columbia and from the Metaline quadrangle. C. D. Campbell's suggestion (Campbell, 1947, p. 600) to not use the names introduced by Weaver (1920) is followed. As Campbell points out, the names Weaver applied to rocks in the Deep Creek area should not be perpetuated because they lack chronologic significance, because the same rock units are included under different names, and because unrelated rocks are included under the same name.

A major rock unit that has no formally named counterpart in a nearby area is designated a "sequence" and is given a geographic prefix after a prominent feature of the area in which it occurs, as for example, the slates, argillites, and argillaceous limestone on Grass Mountain (fig. 2) are designated the Grass Mountain sequence (Cg) and divided into five subunits that have no names but numerical subscript symbols, such as Cg_3, which are used in the text in lieu of names. These rocks are not given formal formation status because both upper and lower limits are faults, and because their age is largely inferred. The above procedure has all the advantages of formalized naming and the additional advantage of keeping usage flexible.

Some common rock names that have no unanimity of usage also need definition. Many rocks were, before metamorphism, the fine-grained detrital sedimentary rocks called shales, mudstones, and siltstones. These are recrystallized into rocks termed slate, argillite, phyllite, schist, and hornfels, whose mineral composition is basically quartz, sericite, and minor chlorite. The slate, argillite, and phyllite after metamorphism remain fine grained but have developed a preferred orientation of the sheet silicates. The slate and argillite are too fine grained to allow megascopic identification of individual grains; the phyllite, a little coarser, has the sheen of incipient recrystallization. The slates are devoid, or almost devoid, of visible bedding; the argillites and the phyllites commonly are bedded, but rarely well-bedded. The slates have excellent planar parting (slaty cleavage), either parallel to or at an angle to the bedding; the phyllites have a fair parting but one inferior to that of the slates; the argillites have no slaty cleavage, but may have a fracture cleavage (not parallel to planar fabric). All three represent the same grade of metamorphism. The term "argillaceous" is applied to rocks containing sericite or muscovite; the connotation of this group, slate, argillite, phyllite, are referred to as lutite or as pelite. The terms, schist and hornfels are used here to refer to rocks of higher metamorphic grade whose component grains are megascopically determinable. The schist has a well developed preferred orientation of the sheet silicates; the hornfels has no preferred orientation of sheet silicates. Both schist and hornfels formed only in the metamorphic aureoles adjacent to the plutonic rocks.

Other rock names follow established usage, quartzites are quartz sandstones recrystallized to a decussate texture, limestone and dolomites are relatively pure end members, commonly recrystallized. One seldom used term, "siltite," refers to the fine-grained equivalent of quartzite.

The rock distribution on the geologic map (fig. 2) when unscrambled into the arrangement of the map explanation indicates a relatively simple depositional history, albeit a puzzling one. This apparent simplicity can be easily misinterpreted by assuming the rock sequences are indigenous. The map, explanation not withstanding, represents three broad divisions of depositional history: Precambrian, lower Paleozoic and upper Paleozoic and also includes a fragmentary record of Mesozoic and Eocene time. The lower Paleozoic division is the only fossil-controlled sequence; the other two divisions were established by less direct methods. Both Precambrian and upper Paleozoic sequences are dominated by fine-grained detrital sediments, the Precambrian tends towards the alumina-rich and the upper Paleozoic tends towards the black shale facies with high silica. Neither sequence has more than trivial amounts of coarse clastics. Both include limestones, but in minor amounts.

The lower Paleozoic sequence, on the other hand, represents a progressive change in deposition. The sequence began during the very late Precambrian with the deposition of clean quartz sand which continued well into the early Cambrian. This was followed by the accumulation of a comparatively thin limestone unit (Reeves limestone), which was succeeded by a thick shale. The shale grades into a thick carbonate unit which in turn is overlain by black graptolitic slates (Ordovician). This general order of deposition holds for the Cambro-Ordovician throughout the area; it is recorded in the Northport, Lime Creek Mountain, Hooknose-Baldy, and Red Top Mountain blocks and is also present south of the Spirit pluton. Variations from block to block in lithologic types and thickness of equivalent units in the lower Paleozoic sequence give clues to the depositional and tectonic history.

The stratigraphy is described block by block. Within a block it proceeds from the oldest to the youngest unit. Because a unit is fully described when first introduced, descriptions are not duplicated if the unit is present in blocks described thereafter; only variations are mentioned. The reader's knowledge of a stratigraphic unit consequently, comes piecemeal; but the alternate approach, jumping from block to block, would make it more difficult to grasp the overall assemblages within a block and differences between blocks.

Hooknose-Baldy block

The Hooknose-Baldy block lies in the eastern part of the area (see index map on figure 2). It is the western extension of Park and Cannon's (1943, p. 28-30) "Flume Creek-Russian Creek block," a name that is not retained because it bounds only a part of the larger strato-structural unit here described. The Hooknose-Baldy block is bounded on the west by the Leadpoint fault, which extends from the northeast corner of the map S. 32° W. to the Spirit pluton. Rocks included in this block range in age from lower Cambrian to Ordovician.

This block is described first because it brings the well established stratigraphy of the Metaline district into the Deep Creek area. The Gypsy Quartzite and the Maitlen Phyllite are shown along the west boundary of Park and Cannon's map. The Metaline Limestone (Metaline Formation in this report) extends from the Deep Creek area into the northwest corner of the Metaline quadrangle where it was not recognized as Metaline Limestone but was correlated with Devonian rocks that had been identified to the south. The Ledbetter Slate, although not continuous with exposures in the Metaline quadrangle, is easily correlated from its stratigraphic position, lithology, and graptolites.

Gypsy Quartzite

Only the upper part of the Gypsy Quartzite, the oldest formation in the Hooknose-Baldy block, is present in the part of the Hooknose-Baldy block in the Deep Creek area, but it crops out in its entirety in the eastern part of the block in the Metaline quadrangle (Park and Cannon, 1943, p. 13-15), where it was named and identified as Early Cambrian. It occurs in the core of the Hooknose anticline (see fig. 2), a badly broken structure and a poor place to establish the details of lithologic variations. About 600 to 750 metres of beds are shown in the composite section (fig. 3) where the section of Sherlock Park is compared with the Metaline section. This partial section is divided into two units; an upper unit of mixed quartzite and schist, which is probably equivalent to the Reno Formation of the West Nelson Map area (Little, 1960, p. 27-30), and a "lower" unit of bedded, relatively pure quartzite (the Quartzite Range Formation of West Nelson map area, p. 24-27). To the north in British Columbia in the Salmo Lead-zinc area (Fyles and Hewlett, 1959, p. 19, Table 3) these two units were divided into subunits that are not recognized with certainty in the Deep Creek area. The top of the formation is defined as the base of the Reeves Limestone Member of the Maitlen Phyllite, an easily recognized horizon, and an abrupt change from the interbedded phyllite and quartzite of the Gypsy to the limestone beds of the Reeves.

Figure 3.--Composite stratigraphic sections of the Cambrian rocks in northeastern Washington and adjacent British Columbia. Sections are plotted on 125,000 scale base of figure 14. Base of Reeves Limestone Member of the Laib-Maitlen Phyllite is located at the approximate center of the composite section. The upper ends of the open ended columns are terminated by either erosion or a fault, the open ended lower ends, by lack of exposure.

From the Reeves Limestone downward, the Gypsy becomes progressively, but irregularly, more siliceous. The uppermost 90 to 210 metres of beds are about equally quartzite and phyllite; the next 150 to 180 metres are roughly three-fourths quartzite and one-fourth phyllite; below this horizon the formation grades rapidly into a rock with a quartz content that will average more than 95 percent. The downward increase in quartz content results not only from a decrease in the number and thickness of phyllite beds, but also an increase in the quartz content of individual quartzite beds[1/]. The thickness of the individual quartzite beds also increases downward; those in the uppermost part of the formation are commonly thinner than 2 inches, those in the lower part of the section range from 4 inches to 4 feet, although some beds have micro layers of phyllite less than one one-hundredth of an inch in thickness.

[1/] Quartzite with a quartz content of over 95 percent are referred to as pure quartzites, those having from 75 to 95 percent quartz as fairly pure quartzites, and those having from 50 to 75 percent quartz as impure quartzites.

The division between the upper and lower units is not along a rigidly defined bedding plane and doubtless may vary as much as 30 metres from exposure to exposure. On the map, the boundary was drawn where phyllite interbeds are insignificant and the rock is essentially a pure quartzite.

The quartzites in the lower unit are light gray to white, medium- to thick-bedded rocks. Grain size and shape of the detrital grains is largely obliterated by recrystallization, however, wherever original grain shapes can be recognized, they are subrounded to rounded and range from coarse sand, 1 mm in diameter, down to very fine angular sand. Most beds in the lower unit of the formation probably consist of medium to coarse sand. Where original shapes of sand grains are determinable, many grains are composite, having the sutured texture of a quartzite, indicating the quartzite is a second generation. The sands are moderately well sorted; in the purer quartzites the larger grains are matrixed with fine quartz and muscovite. Pebble beds and cross bedding are rare. Heavy minerals are not abundant, and are represented by rare small rounded grains of zircon, apatite, and brown tourmaline.

Quartzite beds of the upper unit of the Gypsy Quartzite are darker in color, ranging from dark gray to dark greenish gray, and more variable in composition than those of the lower division. This unit contains, in addition to many phyllite layers, several thin brown-weathering impure dolomite beds 15 cm to 30 cm thick.

The upper part of the Gypsy Quartzite is characterized over wide areas by beds containing euhedral magnetite crystals. These magnetite-bearing beds occur near the base of the upper unit through a stratigraphic interval of 150 metres. Only a small percentage of the beds in this interval are magnetite-bearing and the total iron content of the interval is probably less than 1 percent. The zone, however, is widespread, having been recognized wherever the Gypsy Quartzite is found. It is also exposed about 6 miles southwest of the area on Joe Creek.

Individual beds, 3 to 4 centimetres thick, contain from traces, to as much as 50 percent magnetite. Concentration of magnetite within beds is very erratic. Magnetite grains range seriately from the very minute to those 1 mm in diameter. Crystal form is well developed and appears to be the result of growth later than the foliation. The magnetite occurs not only in quartzite beds but also in phyllite interbeds an occurrence which suggests it is not recrystallized detrital grains but the metamorphic product of a sedimentary iron mineral such as siderite or ankeritic carbonate. In addition to magnetite, most thin sections examined contained dolomite or iron-bearing carbonate.

Maitlen Phyllite

The Maitlen Phyllite is a name given by Park and Cannon (1943, p. 15) to the stratigraphic unit, predominantly phyllite, that over lies the Gypsy Quartzite and underlies the Middle Cambrian Metaline Limestone (Metaline Formation in this report). In the Deep Creek area this phyllite unit occurs in four of the six structural blocks and is abundant in two, especially in the Hooknose-Baldy block.

The Maitlen Phyllite in the Hooknose-Baldy block is divided into two units. The lower unit, the Reeves Limestone Member, consists of about 120 metres of limestone; the upper unit, hereafter referred to as the phyllite unit, consists of more than one thousand metres of phyllite, plus a few thin limestone and quartzite beds. If an unfaulted section of the phyllite unit was available, it might be possible to divide the phyllite unit into two or more members on the position of the limestone, but because of its faulted character it is uncertain whether there is one or several widely separated beds of limestone.

Reeves Limestone Member is the name given by Fyles and Hewlett (1959, p. 25) to a limestone exposed in and near the Reeves MacDonald mine, which is in British Columbia a few miles northeast of the Deep Creek area. The limestone at the Reeves MacDonald mine is correlated by Fyles and Hewlett (1959, p. 25) with the limestone that occurs at the base of the Laib Formation (essentially a synonym for Maitlen Phyllite in Canadian usage) in the Sheep Creek anticline, which in turn is correlative with the limestone unit which crosses the International Boundary near Monument 193 (see Little, 1960, p. 34) and Okulitch, 1948, p. 340-349) and continues southward as a part of the Maitlen Phyllite of Park and Cannon (1943, p. 15). Park and Cannon did not separate it in mapping, but spoke of it as follows: "the base of the phyllite (Maitlen) on Sullivan Mountain and to the northeast is placed about 100 feet below a gray-white limestone band about 200 feet thick." Park and Cannon (1943, p. 15-16) considered this band equivalent to one in a similar stratigraphic position in the Hooknose-Baldy block. Although archaeocyathids of Early Cambrian age have been reported from the limestone that appears on Sullivan Mountain and to the northeast (Howell, 1944 and Okulitch, 1948), none has been found in the limestone band in the Hooknose-Baldy block, however, the writer has little doubt that the correlation indicated by Park and Cannon is correct.

The Reeves Limestone Member is a creamy white, medium- to coarse-grained rock from 90 to 120 metres thick that has fair to poor bedding and a pronounced shear cleavage. It is banded white and gray, but the banding is continuous for only short distances. In places it is faintly mottled with pinkish buff mottles in a white background. The mottles consist of light tan dolomite crystals set in a calcite matrix. Quartz, as rounded grains and in clusters of irregularly shaped grains, is the other common impurity. Although no analyses are available, the silica content is estimated to range from less than 1 percent to about 10 percent. Muscovite, which ranges from a trace to about 1 percent, is concentrated along shear planes.

The relation between the limestone and the overlying phyllite is cyclic. Thin layers of limestone alternate with phyllite through a stratigraphic interval of more than 30 metres.

In a general sense, the phyllite unit is the Maitlen. The Reeves Limestone Member on the basis of its wide areal extent and mapability could readily be removed from the Maitlen and elevated to formation rank. If this were done, the remaining 95 percent of the formation would be in complete agreement with the lithology implied by the name.

The bulk of the phyllite unit is a gray to green-gray, fine-grained rock with the typical sheen of incipient recrystallization that defines a phyllite. Thin beds of fine-grained quartzite are interlayered near the base and one or more thin layered limestones occur in an unmeasured position in the upper half of the unit. Near the Spirit pluton the phyllite is thermally metamorphosed to a fine-grained schist.

The phyllite is composed of quartz, muscovite, plagioclase, and pyrite and limonite in variable proportions. The chlorite probably never exceeds 10 percent of the volume. Both quartz and muscovite range from 10 to 90 percent in the microlayers. The estimated aggregate composition of the phyllite unit is however, muscovite 50 percent, quartz 35 percent, chlorite 5 percent, plagioclase 5 percent, pyrite and limonite 5 percent, plus the inevitable trace amounts of zircon, garnet, sphene, apatite, and tourmaline, which occur as tiny well-rounded grains. Most of the phyllite unit agrees with the above mineral assemblage; however, a narrow band of phyllite extending southwestward from the saddle south of Sherlock Peak is spotted by dark green-gray porphyroblasts of chloritoid. These spotted rocks appear to lie within a stratigraphic zone that has a composition favorable for the formation of chloritoid.

Bedding, although commonly determinable, is not obvious in much of the unit. Bedding is defined by variations in quartz content; where bedding is absent the quartz content is constant. Individual beds are rarely more than 3 cm thick, commonly less than 1 mm.

The phyllite unit is structurally imcompetent. In many places it is crumpled, stretched, and sheared; microfolds are common. Cleavages of at least two generations are well developed. The earliest and most common cleavage lies parallel to, or nearly parallel to, the bedding; later cleavages cross the earlier cleavage at both high and low angles.

The phyllite is thoroughly recrystallized, original boundaries of quartz grains are not determinable. Mica folia developed along each generation of shears. Pyrite cubes, locally common in the upper part of the unit, are distorted and quartz and calcite have grown flamboyantly in the strain shadows of the cubes. Where shearing accompanied recrystallization, the fabric is that of tightly packed quartz lenses bounded by foils of muscovite (sericite) that outline coalescing shears.

Maitlen Phyllite in the southern part of the Hooknose-Baldy block within two miles of the Spirit pluton, is coarser in texture, slightly darker in color, and different in mineralogy. Quartz and muscovite are present, in place of chlorite biotite appears and in the more aluminous phases tabular porphyroblasts of andalusite spot the rock. In the zone a few hundred feet from the pluton, both sillimanite and cordierite are common. Accompanying these changes in mineralogy is a coarsening of grain size and a strengthening of schistosiy through the growth of micas.

The upper contact of the Maitlen Phyllite of the Hooknose-Baldy Block is exposed on the ridge between Republican Creek and the north fork of Currant Creek (Deep Lake quadrangle) as well as on Windy Ridge (Leadpoint Quadrangle). At both places the contact is placed an abrupt change from phyllite to the impure limestone of the basal member of the Metaline Formation. North of Silver Creek (Leadpoint Quadrangle), however, the contact is gradational: beds of phyllite appear in the basal Metaline Formation. These interlayered phyllites are limey in contrast to the almost calcite-free phyllite of the Maitlen. The boundary between Maitlen Phyllite and Metaline Formation was mapped at the base of the lowermost mottled limestone characteristic of the lower member of the Metaline Formation.

Metaline Formation

The Metaline Formation is the Metaline Limestone of Park and Cannon (1943, p. 17-19); "Formation" is substituted for "limestone" to avoid the confusion that comes when speaking of "dolomites of the Metaline Limestone." In the Hooknose-Baldy Block the Metaline Formation is divided into a basal limestone unit, a middle dolomite unit, an intraformational dolomite breccia unit, and an upper unit containing the lithologies of the other units plus black calcareous argillite.

A major thrust fault divides the rocks of the Hooknose-Baldy Block into two plates that juxtapose sections of the Metaline Formation presumably deposited miles apart. The upper plate contains all four units of the formation; the lower plate contains only the lower unit.

At no place in the Hooknose-Baldy Block is there a complete, unfaulted section of the Metaline Formation. Gladstone Mountain has all four units, but only the middle dolomite unit is completely represented. The composite Leadpoint section (fig. 3) represents the general lithologic sequence and probably the aggregate thickness of the formation that occurs in the upper plate of the thrust, but measurements of individual units are subject to an unknown correction for displacements on faults--both known and inferred. Because of the chopped-up nature of the rocks, it was not feasible to measure thickness variations from one end of the block to the other.

The lower limestone unit, the most extensive and thickest unit and the only unit present in both thrust plates is complete in neither plate, therefore its maximum thickness is unknown. The lower part of the lower limestone unit, which is only present in the lower plate, grades into the underlying Maitlen Phyllite and is estimated to be between 450 to 900 metres thick. The upper part of the lower unit is only present in the upper plate where it is overlain by the middle dolomite unit. In both upper and lower plates the lower limestone unit is dominated by a fine-grained dark gray limestone mottled with dolomitic limestone. That in the lower part has greater uniformity; the only variant is a zone from 30 to 90 metres thick crowded with Girvanella, an algae of spherical form. The upper part of the lower unit, representing a thickness of more than 450 metres, contains the same limestone beds that characterize the lower part, and in addition contain thin-layered dark gray limy shale that is increasingly more abundant upward in the section. It also contains a subunit of fine-grained black dolomite from 30 to 90 metres thick and one or more similar but much thinner subunits. Fossils representing a trilobite fauna of Middle Cambrian age were collected from several localities, all from limy shales adjacent to the black dolomite. The collections are described in the appendix. The upper part of the lower unit of the Metaline Formation is best developed on the hill east of Leadpoint, but even here the uppermost beds are not exposed, being cut off by the Leadpoint fault. The uppermost beds are thin-bedded limy shale alternating with dense black limestone, rocks very similar to the "Thin-bedded member of the lower limestone unit of the Metaline Formation" measured by Dings and Whitebread (1965, p. 11) at the Lehigh Quarry Hill, Metaline district.

The contact of the lower limestone unit of the Metaline Formation with the overlying middle dolomite unit is only exposed--and poorly--in the valley of the West Fork of Silver Creek (Leadpoint quadrangle) where the uppermost beds of the lower unit are dense mottled limestone with interlayered limy shale. The change from limestone to dolomite is abrupt.

What is regarded here as the middle dolomite unit of the Metaline Formation includes under a single unit the three subunits that were mapped as composing the middle member of the Metaline Formation in the Leadpoint quadrangle (U.S. Geol. Survey MF-137) and Deep Lake quadrangle (U.S. Geol. Survey MF-237) by Yates and Robertson (1958) and Yates and Ford (1960) respectively. The correlation chart (fig. 4) shows equivalent symbols used on the three maps. The map boundaries for upper and lower limits of the unit are the same as those of the middle member shown on the Deep Lake and Leadpoint quadrangle maps. The three subdivisions shown on the two earlier maps separate a well-bedded middle part from the massive dolomite of a lower and an upper part. The boundaries that define these distinctions, although mappable in some places, are in other places tenuous and do not everywhere represent the same stratigraphic planes. An additional complication to using these subdivisions is in the dolomite near the Spirit pluton where metamorphism largely destroyed the bedding.

The middle part of the middle dolomite unit is typically a fine- to medium-grained banded dolomite consisting of medium gray to dark gray layers alternating with light gray to white layers. The color bands are locally absent or indistinct and here the rock is uniformly light gray or dark to medium gray. The color bands, generally 5 to 45 cms thick, represent beds whose outlines are locally discontinuous and irregular because of migration of the carbonaceous pigmenting material during recrystallization of the rock. Many dark layers contain small blebs and thin streaks of light gray or white dolomite; in other dark layers the distribution of white dolomite gives the rock a worm-eaten appearance. About 300 metres of this banded dolomite is present on Gladstone Mountain. The dolomite both below and above the banded dolomite is a light gray to white, massive medium- to coarse-grained dolomite.

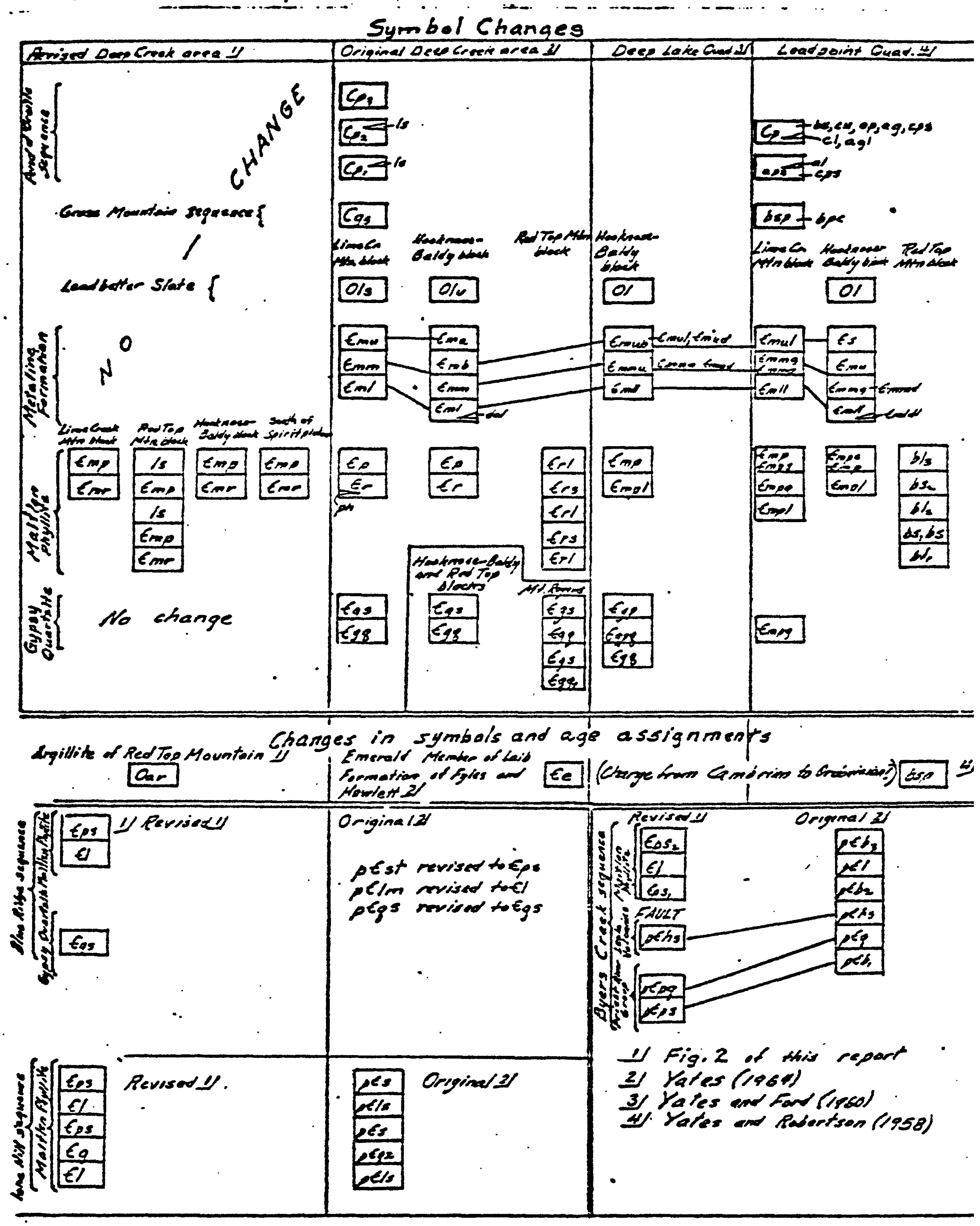

Figure 4. -- Equivalence of stratigraphic symbols shown on the Leadpoint and Deep Lake quadrangles and the Deep Creek area.

The intraformational dolomite breccia unit of the Metaline Formation is, as the name implies, entirely dolomite, but differs principally from the underlying unit by being about one-third an intraformational breccia of dolomite fragments in a dolomite matrix. The remaining two thirds of the unit commonly is a dark gray to light gray, medium-grained to fine-grained, thin bedded dolomite. In places the bedding is vague, poorly defined, or absent; where absent the rock is medium gray in an aggregate color, which on closer inspection is seen to be a compromise between a mixture of white and black dolomite grains. A field name for this variety was "salt and pepper" dolomite.

The two varieties of dolomite, breccia and bedded, alternate in the section; the alternations are from 8 metres to 100 metres thick, which in the aggregate total more than 400 metres thick. Boundaries between the two are sharp or gradational. Lateral changes from breccia to nonbreccia could not be measured in the field, but small scale gradations suggest that large scale gradations probably exist. The gross layering is overall parallel, but in detail the contacts between breccia and bedded dolomite are highly irregular.

The breccia fragments are mostly thin slabs broken parallel to bedding; they range from less than one centrimetre to 1 1/2 metres in length, are subangular to angular, and occur unsorted in a matrix of fine-grained darker dolomite (fig. 5a). In some places fragments are preferentially oriented parallel to the bedding planes of nearby bedded dolomite; more commonly the fragments are rotated; the relatively few fragments that are bent, probably are the result of differential compaction during their accumulation. Locally recrystallization has destroyed or obscured outlines of breccia fragments or modified bedding planes to form banded to irregular segregations of darker dolomite. On the other hand, the outlines of some breccia fragments are preserved and their shapes enhanced by replacements of quartz along their borders. Quartz also occurs as irregular nodules, lenses, sheets, and peculiar button-shaped nodules (fig. 5b). Some of the nodules probably were deposited as chert, which later recrystallized to quartz. Silicified mud cracks were found at one place on the west slope of Gladstone Mountain (fig. 5c). Fossils (phosphatic brachiopods) were found in three places: two localities are near the base of the unit on Gladstone Mountain and the other, near the top of the unit, on the old Frisco-Standard road. One of these localities on the north side of Gladstone Mountain is that of C. D. Campbell (1947, p. 606) who found Acrotreta sp. and Nisusia sp., recognized by Josiah Bridge as Early Middle Cambrian.

The contact of the intraformational breccia unit with the underlying dolomite is gradational; that with the overlying, uppermost, unit is everywhere faulted, but the presence of intraformational breccia in the uppermost Metaline unit suggests that this contact is also gradational.

The intraformational breccia unit is continuous with rocks in the northwest corner of the Metaline quadrangle, which was mapped as Devonian by Park and Cannon (1943, p. 22-23), who considered it the same unit as a limestone bearing Devonian fossils that outcrops 6 1/2 kilometres to the southeast. The correlation was based on a similarity to a "limetsone breccia or conglomerate" at the base of the fossil bearing unit. They did not consider it part of the Metaline Formation because it did not resemble any of the rocks that composed that formation in the Metaline quadrangle. The tracing of this intraformational dolomite breccia from the northwest corner of the Metaline quadrangle into the Deep Creek area, where it is demonstrately a part of the Metaline Formation and where it contains Cambrian fossils firmly establishes its Cambrian age and its lack of affinity with the Devonian rocks.

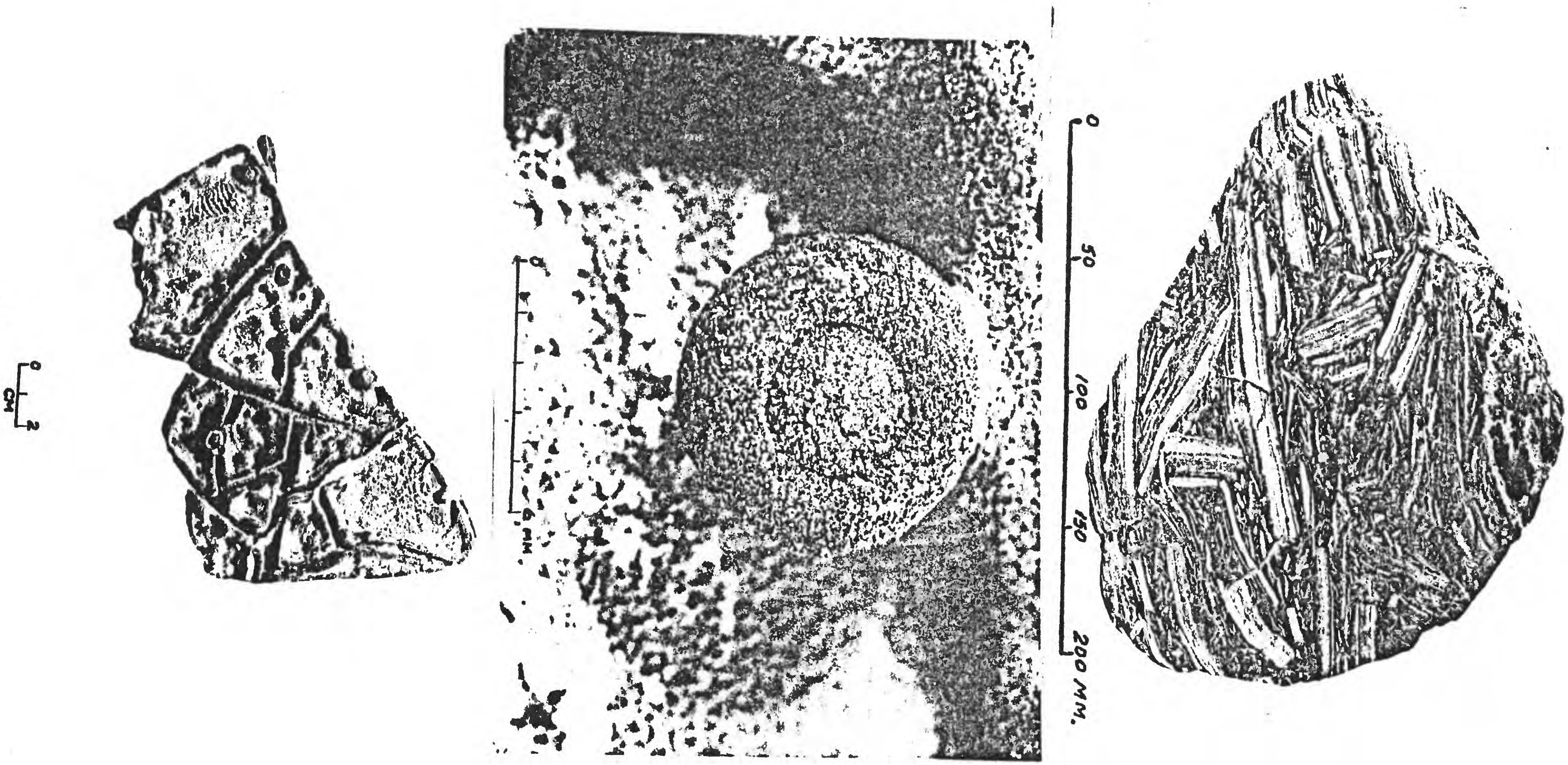

Figure 5. — Photographs of rocks in intraformational breccia unit, Metaline Formation.

A, typical breccia; B, nodules; C, mud cracks.

The upper unit, which includes both clastic and carbonate rocks, has beenthe most difficult unit to fit into the stratigraphic sequence of Cambrian rocks. Campbell (1947, pl. 1) included some exposures of the unit with the Metaline Formation and other exposures with the Maitlen Phyllite. Yates and Robertson (1958) show the unit on the preliminary map of the Leadpoint quadrangle as a unit of questionable Cambrian age lying between the Metaline Formation and the Ordovician Ledbetter Slate. Fossils were later found which confirmed the inferred Cambrian age and stratigraphic position, but until these fossils were found, the possibility that the unit was younger than the Ledbetter Slate could not be eliminated. The contact with the overlying Ledbetter is exposed southeast of Deep Lake and southeast of Leadpoint (see fig. 2), where it is both conformable and abrupt. Nowhere has it been seen in depositional contact with the underlying unit. This upper unit of the Metaline was not identified in the northeast corner of the map area (fig. 2), where exposures are so poor that the depositional contact interpreted between the dolomite breccia unit and the Ledbetter Slate may be in actuality a fault that has eliminated the upper unit.

The estimated thickness of about 500 metres (see fig. 3) for the upper unit is conjectual; at no place is an unfaulted section exposed, in fact, the restored section is largely an interpretation of the better exposures of the unit that are in widely separated localities, the lower part west of Deep Lake and the upper part southeast of Leadpoint. Although no key beds were found that are common to both exposures, this does not necessarily indicate that a middle part of the unit is missing because rapid lateral lithologic changes are more probable in this, than in any other Cambrian unit.

The lower part of the unit contains medium-bedded, black, medium-grained dolomite, black limy shale, and intraformational dolomite breccia. Above this is a dark, fine-grained limestone with a faint buff mottling, a rock that can be easily confused with the mottled limestone of the lower unit. The mottled limestone contains two or more discontinuous layers of dolomite breccia. The upper part of the unit as seen southeast of Leadpoint consists of about 300 metres of medium-grained gray dolomite, in part a breccia. Midway in the dolomite is about 45 metres of shaly dolomite, gray brown in color. At the very top of the dolomite is a metre or more of limestone on which rests the Ledbetter slate.

Age

Cambrian fossils occur in the Metaline Formation in the lower limestone unit, intraformational dolomite breccia unit, and upper unit. Collections of trilobites from six localities in the lower limestone that were studied by Allison R. Palmer were assigned to the Bathyuriscus-Elrathina zone, which is Middle Middle Cambrian. Faults of unknown displacement, both known and inferred prevent locating the collections stratigraphically with any accuracy: an intermediate position in the lower limestone is interpreted. Collections from three localities in the intraformational breccia unit yielded phosphatic brachiopods that Palmer considered Middle Cambrian. From the top of the upper unit, a collection of phosphatic brachiopods that seemed referable to Conotreta, "appeared to be Late Cambrian in age."

The following fossil collections from the Metaline Formation in Leadpoint quadrangle were studied and identified by Allison R. Palmer, whose opinions on probable age are quoted below.

Coll. F-4 Sec. 14, T. 39 N., R. 41 E., Elev. 2,480 feet; 915 metres from SE cor. of sec. In thin bedded, light tan limy dolomite in upper unit of Metaline Formation. "This collection (by Yates) contains fragments of a phosphatic brachiopod with strong concentric ridges. Brachiopods of this type are generally referred to Paterina. They are known from many levels in the Cambrian and in addition, from the Middle Ordovician of eastern United States. The collection cannot be more closely dated than Middle Ordovician or older." A. R. Palmer, 11-29-55

Coll. F-7. Sec. 14, T. 39 N., R 41 E.; elev. 2720; S. 20° W. 700 metres from NE corner of section. Collected from dolomite breccia of upper unit of Metaline Formation within 30 metres of base of Ledbetter Slate.

After solution in formic acid, the residue was found to contain "linguloid brachiopods, small phosphatic brachiopods that seemed referable to Conotreta, acrotretid brachiopods that lack a distinct intertrough in the conical pedicle, valve, and spinose scraps of the appendages of an undetermined arthropod.

Conotreta is a late Cambrian brachiopod known from Alaska and central Nevada. The lack of a distinct intertrough in the acrotretid brachiopods is generally a characteristic of Upper Cambrian forms. The linguloids and asthropod appendages have no present stratigraphic significance. This collection appears to be of Late Cambrian age on the basis of present knowledge." A. R. Palmer 11-29-55

The above locality of F-7, was collected by Palmer in July, 1956 (LPW- 4-56). The new collection contained graptolites and fragments of an agnostid trilobite. "Two pygidia of an agnostid...seem most referable to Lotagnostus, a genus of Late Cambrian age. Agnostids become quite rare in Lower Ordovician rocks and it is possible that the forms here referred to Lotagnostus could be as young as Lower Ordovician. They are certainly no older than middle Late Cambrian." A. R. Palmer, 12-13-56

The graptolites in collection LPW-4-56 (USGS No. 2176€0) were studied by R. J. Ross, Jr. who reported, "Dictyonema sp. (and an) unidentifiable fragment possibly belonging to Diolymograptus. Probably Ordovician but no positive proof here. Could be younger." 1-9-57

Another collection by Palmer (LPW-5-56) 3 metres higher stratigraphically than LPW-4-56 is reported on as follows, "This collection contains a specimen of the phosphatic brachiopod Schizambon and a single nearly smooth agnostid. The association of these forms indicates either a late Cambrian or early Ordovician age for the collection." A. R. Palmer, 12-13-56

Three collections, F9, F13, and F14 were made from and immediately below the intraformational dolomite breccia unit and these are also reported on by Palmer (11-29-55).

Coll. F9. Sec. 19, T.39N., R.42E.; elev. 3320; S.26°E., 655 metres from NW corner of section. Collected from light gray dolomite.

"This collection contains fragments of an indeterminate acrotetid brachiopod and of a brachiopod with a quincunxial pattern of ridges on its outer surface. Forms with this pattern are generally referred to Dictyonia.

Although Dicyonina-like brachiopods are known from rocks as young as Middle Ordovician, the ones from this collection have more of the aspect of Middle Cambrian forms than of those presently known from younger rocks. This collection is probably Middle Cambrian in age." A. R. Palmer, 11-29-55

Coll. F-13. Sec. 7, T.39N., R.42E.; elev. 3680 ft; N.33½°E. 137 metres from SW cor. of sec. Collected from gray dolomite unit near base of dolomite breccia unit. "This collection contains acrotretid brachiopods and a specimen referable to Paterina. Its age is Middle Ordovician or older." A. R. Palmer, 11-29-55

Coll. F-14. Sec. 7, T.39N., R.42E.; elev. 3760 ft; 90 metres in N. 45° E. from SW cor. of sec. About 30 metres above locality, F-13.

This collection contains a Middle Cambrian assemblage of phosphatic brachiopods referable to Acrothele, Homotreta, and Paterina. This locality is probable the same as the collected by C. D. Campbell (1947, p. 606), whose collection contained, according to Josiah Bridge, two brachiopods, Acrotreta sp. and Nisusia sp. considered early Middle Cambrian in age.

The lower limestone unit of the Metaline Formation supplied six fossil localities in the Leadpoint quadrangle that represented the Bathyuriscus-Elrathina fauna of the Middle Middle Cambrian. Most of the identifiable fossils have been described from the Stevens Formation of the Canadian Rockies. The six collections described below were identified by A. R. Palmer.

Coll. F-6. Sec. 14, T.39N., R.41E.; elev. 2320 ft; S.61°W., 730 metres from NE corner of section. Collected from thin bed dark gray limestone.

cf. Alokistocare paranotatum Rasetti
Elrathina cordillerae (Rominger)
Ogygopsis Klotzi (Rominger)
Oryclocephalus reynoldsi Reed
Pagetia sp.
Parkaspis sp.
Peronopsis columbiensis Rasetti
Peronopsis montis (Mathew)
Zacanthoides divergens Rasetti

Coll. 55-R-7. Sec. 6, T.39N., R.42E.; elev. 3440 ft; S.39½°E., 685 metres from NW cor. sec. Collected from dark gray, thin bedded argillaceous limestone.

Alokistocare cf. A. sinuatum Rasetti	Kootenia burgessensis Resser
Alokistocare sp.	Nisusia sp.
Bathyuriscus adaeus Walcott	Oelandia sp.
Elrathia permulta (Walcott)	Olenoides serratus (Rominger)
Elrathina cf. E. cordillerae (Rominger)	Oryctocare sp.
Glysphaspis sp.	Oryctocephalus burgessensis Resser
Helcionella sp.	Pagetia cf. P. bootes Walcott
Homotreta sp.	Peronopsis sp.
Hyolithes sp.	Pegmatreta sp.
Klotzeilla? sp.	Zacanthoides divergens Rasetti
Kootenia cf. K. spencei Resser	Zacanthoides romingeri Resser

A. R. Palmer, 11-29-55

Coll. 55-R-8. Sec. 6, T.39N., R.42E.; elev. 3300 ft; N.32½°W, 850 metres from SE cor. sec.

Bathyuriscus? sp.

Elrathina sp.

Pagetia sp.

Peronopsis sp.

Pegmatreta sp.

Coll. 55-R-13. Sec. 31, T.40N., R.42E.; elev. 3660 ft; 670 metres N.88°E. from SW cor. of sec. Collected from dark gray limestone and probably stratigraphically equivalent to locality 55-R-7, but 600 metres north along strike.

Bathyuriscus sp.

Elrathina sp.

Olenoides? sp.

Pagetia sp.

Peronopsis sp.

Pegmatreta sp.

Coll. 55-R-18. Sec. 31, T.40N., R.42E.; elev. 3700 ft; N.84E., 840 metres from SW cor. sec. Collected from dark gray limestone.

Kootenia sp.

Nisusia sp.

Coll. 61-1 (3798-CO) Sec. 31, T.40N., R.42E., N. 170 metres from SE cor. of sec. Dark gray limestone.

Peronopsis sp.	Oryctocephalus sp.
Ptychagnostus sp.	Bathyuriscus sp.
Pagetia sp.	Elrathina sp.

2 undet. ptychoparioids

A. R. Palmer, 4-24-62

Ledbetter Slate

The Ledbetter slate, the youngest sedimentary rock in the Hooknose Baldy block, was named by Park and Cannon (1943, p. 20) from exposures containing Ordovician graptolites on the slopes west of Ledbetter Lake in the Metaline quadrangle. Because rocks mapped as Ledbetter Slate by Park and Cannon were later found to contain Silurian and Devonian fossils, Dings and Whitebread (1964, p. 23) redefined the formation to include only rocks of Ordovician age. It is a poorly bedded, extremely fine-grained metamorphosed black shale, which in places is graptolite-bearing and has well developed, but erratically distributed slaty cleavage. In the Metaline quadrangle it rests with sharp but conformable contact on the uppermost limestone beds of the Metaline Formation and continues from here upward through 670 metres of section with no change in lithology to slates that bear Silurian graptolites. In the Hooknose-Baldy block the lower contact is similar to that in the Metaline quadrangle, but its upward limit is everywhere fixed by faults. Graptolites (see fig. 19A) collected from near the base of the formation and from several undetermined stratigraphic levels within the formation, although poorly preserved, can be assigned to the Ordovician. The thickest and most continuous section, southeast of Deep Lake, has a minimum thickness of about 750 metres. Since this section is not internally controlled by fossils, it is conceivable that it could include Silurian as well as Ordovician rocks.

Unweathered Ledbetter Slate is a dark gray to black, extremely fine-grained rock, which rarely has discernible bedding but commonly has one or more cleavages. The bedding where present is always faint and is represented by slight color changes unaccompanied by changes in grain size. The dominant cleavage parallels bedding or makes a small angle with it in the few places where the two features could be observed together. Commonly one or more sets of cleavage, or closely spaced joints, intersect the dominant cleavage at large angles. Graptolites are only preserved where the strong cleavage parallels the bedding.

Examination of thin sections and a chemical analysis of a sample collected from the south slope of Black Canyon (southern part of Boundary quadrangle) indicate that it is composed principally of quartz and muscovite and a lesser amount of chlorite. The color comes from finely disseminated carbonaceous material, obviously not graphite as it can be bleached by moderate amounts of heat. In places the rock contains visible crystals of pyrite, a mineral that in fine disseminations may be partly responsible for the color. Most rock does not contain calcite, but some gives a slight efferescence when tested with dilute hyrodchloric acid.

The formation is remarkably uniform in lithology. The thin quartzite beds present in the Ledbetter in the Metaline quadrangle (Park and Cannon, 1943, p. 20 and Dings and Whitebread, 1965, p. 2-5) are absent. A thin, discontinuous layer of fine-grained gray limestone from 30 to 90 centimetres thick, which is locally replaced by barite, occurs 213 metres above the base of the section southeast of Deep Lake. Thin argillaceous limestones, from 5 to 30 centimetres thick occur in rocks mapped as Ledbetter in the northeast corner of the Deep Creek area, but exposures are so poor here that the correlation of these rocks is uncertain.

Lime Creek Mountain block

The Lime Creek Mountain block is a triangular area separated on the east from the Hooknose-Baldy block by the Leadpoint fault and on the northwest from the Red Top-Grass Mountain block by the Black Canyon and O'Hare Creek faults (see index map for fig. 2). The south side of the block is intruded by the Spirit pluton. The Lime Creek Mountain block contains the same general sequence of lower Paleozoic rocks as that in the Hooknose-Baldy block, as well as a unit containing probable Devonian fossils. The Gypsy quartzite and the Reeves Limestone Member of the Maitlen Phyllite are separated from the phyllite member of the Maitlen by an overturned thrust fault. Descriptions of the various units emphasize differences between stratigraphically equivalent rocks described in the preceding section.

Rocks below the thrust fault Gypsy quartzite

The Gypsy quartzite, the oldest rock in the Lime Creek Mountain block, occurs in structurally isolated fault segments along the southern part of the block. The upper part of the Gypsy is best represented; only a few hundred feet of the lower part crops out at a single locality, about 1.6 kilometres north od Deep Lake, where it is isolated by faults and alluvium. Relations between the upper and lower parts, therefore are unknown, as is the amount of section that has been removed by faulting. The aggregate thickness of the exposed parts of the formation is roughly estimated to be about 450 metres.

The lower part of the Gypsy quartzite is a white to light gray, medium-grained quartzite similar to that in the Hooknose-Baldy block. If anything, it is more massive; bedding is rarely discernible. The exposed upper part contains much less quartzite than the upper unit in the Hooknose-Baldy block, as it is about one half interbedded phyllite. The quartzite beds are gray to green-gray, are commonly stained brown by limonite, and are never pure; some contain as much as 50 percent mica and variable amounts of calcite. The upper part also contains a few thin lenses and layers of medium-grained, buff to gray sandy limestone. Magnetite-bearing beds, similar to those in the Hooknose-Baldy block, occur at what may be the same stratigraphic horizon.

Maitlen Phyllite

The Reeves Limestone Member is shown in the Lime Creek Mountain and Red Top-Grass Mountain blocks in the geologic map of the Deep Creek area published as Map I-412 (Yates, 1964) as a multiple unit intercalated with phyllite or schist. At the time this correlation was made, the relations between the carbonate and phyllite units of the Maitlen were not understood. After mapping the Northport quadrangle to the west (Yates, 1971) where the Maitlen consists of an alternation of limestone and phyllite units, with phyllite in the minority, it was realized that the Reeves Limestone Member is a name that should apply only to the limestone unit immediately above the Gypsy quartzite and that the other limestones should go unnamed, at least for the time being. This correction has been made on the revised geologic map of the Deep Creek area (fig. 2).

The Reeves Limestone Member in this block is much thicker than that in the Hooknose-Baldy block, but it is so structurally confused that only generalizations can be made about its thickness, lithologic variation, and stratigraphic relations. It overlies the Gypsy quartzite with a sharp but conformable contact. At the one place (south center of sec. 21, T.39N., R.40E) where this contact can be clearly seen, the beds immediately above the quartzite are mottled like those of the lower limestone unit of the Metaline Formation. About 100 metres to the northeast the contact is less sharp and the transition is a few feet of buff sandy limestone. Two or three thin layers of similar buff sandy limestone, or limy quartzite, occur further up within the mottled limestone. These layers cannot be traced laterally more than 90 metres and were not found elsewhere in what appears to be comparable parts of the section. The mottled limestone beds appear to be confined to the lower part of the Reeves Member, up section the rock is a light gray well-bedded limestone; together they have a maximum thickness of 300 metres.

The phyllite is so similar to that in the upper plate it needs no further description. That immediately above the Reeves Limestone is 60 metres thick and is overlain by 90 metres of bedded limestone similar to that forming the upper part of the Reeves. Some of the limestones in the faulted segments probably are other units that lie stratigraphically above this unnamed limestone.

Rocks above the thrust fault Maitlen Phyllite

The general mineralogy of the phyllite unit is very similar to that of the phyllite of equivalent metamorphic grade in the Hooknose-Baldy block. It is typically a fine-grained, gray to green-gray, faintly thin-bedded quartz-muscovite-chlorite rock having well developed cleavage. Dark gray sheared varieties and calcareous varieties are present. Rocks spotted with porphyroblasts of chlorite and chloritoid represent compositional variations particularly susceptible to metamorphism Near the Spirit pluton the rocks are schistose and the porphyroblasts are biotite and andalusite.

The thickness of the phyllite unit exceeds 1070 metres, of which at least 90 metres is limestone in beds from 15 metres to 60 metres thick. These limestones appear to occur in the middle or lower middle part of the Maitlen. They may be at approximately the same stratigraphic position as the thinner finer grained limestones in the phyllite member in the Hooknose-Baldy block. They differ from these by being less pure; some contains almost 50 percent quartz, muscovite, and plagioclase in varying proportions with quartz always dominant.

Metaline Formation

The Metaline Formation in the Lime Creek Mountain block has the lower limestone unit and the middle dolomite unit of the Hooknose-Baldy block and has in addition a basal phyllitic limestone unit and an upper limestone unit which are equivalent to units that occur in the Metaline quadrangle (Park and Cannon, 1943) but do not occur in the Hooknose-Baldy block. The intraformational dolomite breccia unit and the overlying unit of mixed carbonate lithology, characteristic of the Metaline Formation of the Hooknose-Baldy block are both absent in the Lime Creek Mountain block (fig. 3). Total thickness is about 3,000 metres.

The phyllitic limestone unit, about 300 metres thick, consists of dark gray fine-grained limestone and thin interlayers of black phyllite. This combination of competent and highly incompetent layers yielded to folding by flowage of the phyllite and fracture and boudinage of the limestone thus producing an intimate mixture of the two phases, which as a result, seldom exhibits bedding. The phyllite is very similar to the highly sheared phase of the Maitlen, and the limestone resembles those of the overlying lower limestone unit. This phyllitic limestone unit represents a transition from the phyllites of the Maitlen to the limestone of the Metaline formation; however, it does not represent a gradual change from phyllite to limestone, because the proportions of these rocks vary but little from top to bottom of the unit.

The lower limestone unit of the Metaline Formation, which has an estimated thickness of about 460 metres, consists of limestone very similar to that in the Hooknose-Baldy block and a phyllite subunit, 60 to 90 metres thick, whose base is about 120 metres above the base of the unit. The phyllite subunit was not recognized in the western part of the block. Missing from the section are the fine-grained black dolomites and shaley limestones that occur in the lower limestone unit in the upper plate in the Hooknose-Baldy block.

The contact with the overlying middle dolomite unit is abrupt. The upper part of the unit in the lower reaches of Sherlock Creek has been hydrothermally dolomitized without any change in the general appearance of the rock.

The middle dolomite unit is the counterpart of the middle dolomite unit of the Hooknose-Baldy block and is the youngest stratigraphic unit common to both blocks. Near the mouth of McKinnon Creek, is the only apparently unfaulted section, 1430 metres thick. The unit is conformably overlain, apparently without transitional beds, by the upper limestone unit. Above the Deep Creek mine this boundary is difficult to define, but this may be due in part to hydrothermal dolomitization. The only nondolomite in the unit is a 15 metre limestone similar to the lower limestone; it occurs only in the western end of the block.

The upper limestone unit, 760 metres thick, is a light gray to white medium to coarse grained limestone or marble. Bedding is poor and is usually defined by streaks and irregular bedding-controlled replacements of gray medium-grained dolomite. Quartz occurs in the limestone as single grains, clusters of grains, and small nodules, which are probably recrystallized chert; muscovite occurs as thin flakes along bedding and cleavage planes. The contact of the unit with the younger Ledbetter Slates is sharp.

The Ledbetter Slate

Rocks in the Lime Creek Mountain block mapped as Ledbetter Slate consist of three units: (1) a lower unit of 150 metres of black, graptolitic slate, (2) a middle unit of 150 metres of soft, black, "sooty" argillaceous limestone, and (3) an upper unit of 180 metres of fine-grained quartzite or "siltite." The black slate unit is typical of that in the Hooknose-Baldy block and needs no further description, but the other two units, although restricted to this block, are distinctive enough to be of stratigraphic value. Because neither of these two units appear in the type section of Ledbetter Slate in the Metaline quadrangle, which Dings and Whitebread (1964, p. 23) redefined by restricting it to rocks of Ordovician age, there is some question as to whether they should be included here, under the heading of the Ledbetter Slate.

The sooty limestone is well exposed in cuts along the road to the Advance mine. From the mine it can be traced for about 3 kilometres to the northeast and less than two kilometres to the southwest. Whether the unit pinches or is eliminated by faults was not determined. It grades downward into the slate unit, and is overlain abruptly by the siltite unit. It is a thin- to medium-bedded rock that ranges considerably in composition. Under the microscope it appears to contain more carbonate minerals than the field acid test indicated, which suggests that it is dolomitic. Besides the carbonate mineral, it is composed of silt size grains of quartz, shreds of sericite and chlorite and flakes of black slate.

The upper unit of the Ledbetter Slate, the siltite, is a dark gray, massive to micro-bedded, and fine-grained. It is in fault contact with all younger rocks and its distribution is similar to that of the sooty limestone. Most of the siltite is poorly bedded; thin-bedded and micro-bedded varieties occur in the upper 100 feet. The micro-bedded rock has structures indicative of depositional slumping. The massive variety is mottled by brownish, olive-green blotches. The composition is about 75 percent subrounded quartz, with the remaining 25 percent about equally divided between shreds of muscovite, detrital grains of albite and microcline, and limonite, pseudomorphing pyrite, calcite, and dolomite. A few well rounded grains of zircon and blue-gray tourmaline were seen.

Devonian argillite unit

Above the siltite lies a black argillite, which was separated only on the southeastern side of Black Canyon, in the south central part of the Boundary quadrangle. Although the argillite of the unit is indistinguisable from that of the several black argillite units that occur north of the Black Canyon fault in the Red Top Grass Mountain block, the unit is distinctive mainly because it contains several quartzite lenses and beds, as well as, a 20 centimetre thick bed of clastic limestone composed largely of crinoid debris and finely comminuted calcite of indeterminate origin. Among this material was found one favositoid coral, which, according to J. Thomas Dutro, Jr. (personal communication) is "probably the genus, Alveolites." The specimen "is not well enough preserved to enable specific identification. Indeed, the generic assignment must be somewhat doubtful. The range of this genus is Silurian through Devonian." Because of its stratigraphic position a Devonian age for this unit is considered more probable than a Silurian age.

The unit is tentatively correlated with the unnamed reef limestones in the Metaline quadrangle that are described by Park and Cannon (1943, p. 22-23) and Dings and Whitebread (1965, p. 27-30). These lenses and reefs of fossiliferous clastic limestone occur in slates that lie above the graptolitic beds of the Ordovician Ledbetter Slate. The limestones contain middle Devonian brachipods and corals and the slate below the limestones contain Silurian graptolites. On fossil collection from the limestone contained Alveolites sp., and another contained a coral considered Favosites limitaris and a form possibly Alveolites sp.? (p. 29-30). The lithic association of the Devonian limestones of the Metaline quadrangle is very similar to that of the Black Canyon area.

Sparse exposures and probable structural complexities prevented resolving the internal stratigraphy of the unit. The argillite of the unit is the typical fine-grained, black rock that locally exhibits good slaty cleavage and as such is indistinguishable from other slates in the map areas, from which it is separated by the presence of the above mentioned clastic limestone bed and sparse quartzite lenses and beds. The number and exact stratigraphic position of these lithic variants is uncertain but the best exposures (central part of sec. 13, T.39N., R.40E.) suggest that the clastic limestones are stratigraphically below the quartzites. The thin fossiliferous clastic limestone bed appears to be about 30 metres above the siltite unit of the Ledbetter; about 75 metres above the siltite, one outcrop contained a lense 3 metres thick, of conglomerate composed of clasts of limestone and black chert in a limestone matrix. The largest quartzite lense, which is 6 metres thick, and composed of massive dark gray, medium fine-grained dolomitic quartzite, lies about 150 metres from the contact. Further to the northeast the black argillite is interlayered with 15-centimetre layers of dark gray quartzite, which appear to lie at the same stratigraphic position as the quartzite lense.

Red Top-Grass Mountain block

The Red Top-Grass Mountain block is a triangular area, bounded on the south by the Black Canyon and O'Hare Creek faults and on the north by the Columbia fault. Rocks in the eastern end of the block are referred to as the Red Top Mountain sequence and are equivalent to Cambrian units described above; those in the western part of the block, the Grass Mountain sequence, are unique to the block and are believed to be upper Paleozoic. The Grass Mountain sequence yielded no fossils and conceivably--but unlikely--is a facies variant of the post-Cambrian, lower Paleozoic rocks. The O'Hare Creek fault separates the Grass Mountain rocks from those in eastern part of the block.

Red Top Mountain assemblage

Rocks in the eastern part of the block include units that are correlated with the Gypsy Quartzite and the Maitlen Phyllite. Included are six, easily separated, light-colored rock units of quartzite, limestone, and phyllitic schist and one contrasting unit of black argillite that suggests a tie to the predominant black argillite of the Grass Mountain sequence. The units, from oldest to youngest, are quartzite, limestone, schist, limestone, schist, limestone, and argillite.

Correlations of some units within the sequence can be made confidently, those of other units are tenuous. The quartzite is correlated with the Gypsy Quartzite, the lower limestone with the Reeves Limestone Member, and the middle limestone is an uncorrelated and unnamed limestone in the phyllite of the Maitlen. The two phyllitic schist units belong to the Maitlen Phyllite. The upper limestone unit, which is conformably overlain by the argillite unit, is interpreted as equivalent to the upper limestone of the Metaline Formation that occurs in the Columbia anticline of the Northport quadrangle (Yates, 1971).

The argillite unit is an enigma. If the upper limestone unit is an age equivalent of the upper limestone of the Metaline Formation, the overlying argillite should be equivalent to the Ledbetter Slate. Although the argillite, which is unfossiliferous, belongs to the black shale facies, it is too dissimilar to the Ledbetter and the correlation too tenuous to extend the name Ledbetter to the Red Top-Grass Mountain block.

On the other hand, the correlation of the argillite on Red-Top Mountain with the Emerald Member of the Laib Formation (Maitlen Phyllite) of the Salmo District of British Columbia (Fyles and Hewlett, 1959, p. 26) that was made on the geologic map of the Deep Creek area (Yates, 1964) is wrong. At the time this correlation was made the geology of the Northport quadrangle was unmapped and it was not realized that the uppermost limestone on Red-Top Mountain was equivalent to the Metaline Formation, and that therefore neither Emerald nor Upper Laib could be above this unit in the Red-Top Mountain block. Because there is no positive evidence that links the argillite to the Ledbetter Slate and good evidence that the argillite is not equivalent to the Emerald Member, or to any part of the Laib Formation, it is in need of a name of distinguish it from the several other argillites in the Red Top-Grass Mountain block. It is therefore informally named argillite of Red Top Mountain on the geologic map (fig. 2) a name shortened in the text to the Red Top argillite--strictly as an informal means of identification.

Gypsy Quartzite

The Gypsy Quartzite is subdivided into two units: (1) a lower, massive, white to light gray quartzite and (2) an upper, thin-bedded quartzite interlayered with quartz muscovite schist. The lower unit is a gray to white, medium- to coarse-grained rock composed of elongate, vitreous quartz grains. Bedding is very poorly developed and, where observed, was mainly indicated by thin partings of mica shist. Cross-bedding was not seen, but strong recrystallization under stress may have obliterated this, as well as other sedimentary features. The base of the unit is not exposed.

Above the massive quartzite lies about 250 metres[1/] of thin-bedded quartzite intercalated with quartz sericite schist. This unit is a medium-grained, gray to brownish gray rock with well developed bedding. Many beds contain small amounts of muscovite, and some beds, biotite concentrated in micro-layers. The quartz is typically leaf-like, which gives the rock a pronounced planar structure parallel to the bedding. The interlayers of schist are mineralogically similar to the quartzite layers, only a mica is dominant and quartz subordinate. Within 30 metres of the top of the unit, are beds of magnetite-bearing quartzite similar to those in the Hooknose-Baldy block. Towards the top of the unit, near the overlying limestone, the quartzite contains as much as 20 percent of calcite cement.

Maitlen Phyllite and Metaline Formation

The quartzite is succeeded by three mappable units of limestone separated by phyllitic schist. The lower (Reeves) and middle limestones belong to the Maitlen Phyllite, the upper to the Metaline Formation. All three limestones are very similar except for thickness: the Reeves is from 15 to 30 metres thick, the unnamed Maitlen limestone 30 to 150 metres thick, and the upper, or Metaline limestone is in excess of 365 metres thick. All are medium-grained recrystallized limestones that preserve no sedimentary structures other than bedding, which ranges from well-bedded to poorly bedded. Bedding commonly is identified by color variations that range from white to light gray to medium gray, but it is also identified by concentrations of thin foils of muscovite that lie parallel to the color layers. Compositionally they are low magnesian limestones, containing on the average less than 5 percent of magnesian carbonate. According to the few samples analyzed (table 1) the Reeves (no. 2) and the unnamed limestone (no. 1) are the most impure. The upper unit, the Metaline limestone (no. 3) is a high quality limestone suitable for many industrial uses (Mills, 1962, p. 70).

The Reeves and unnamed Maitlen limestones are separated by a greenish gray phyllitic schist about 30 metres thick. This rock has the well foliated fabric of a schist and the grain size of a phyllite. It is composed predominantly of muscovite and lesser amounts of chlorite and quartz plus minor amounts of recrystallized calcite and sparse grains of detrital heavy minerals and grains of euhedral magnetite. It has no bedding, but a well-developed cleavage along the foliation. In the Red Top mine the unit encloses a thin layer of limestone, which was not recognized elsewhere.

Table 1.--Chemical analyses, in percent, of limestones from the south slope of Red Top Mountain.

	1	2	3	4	5	6
SiO_2	2.3	2.9	32.6	1.2	9.2	2.5
Al_2O_3	.92	1.1	.22	.18	.15	.94
Total Fe as Fe_2O_3	.46	.92	.06	.04	.02	.66
MgO	8.8	.82	1.4	1.3	1.4	.72
CaO	43.8	52.5	35.7	54.2	49.4	53.1
Na_2O	.10	.08	.04	.08	.07	.08
K_2O	.24	.24	.02	.04	.02	.22
TiO_2	.04	.03	.02	.02	.02	.02
P_2O_5	.14	.07	.08	.03	.02	.05
MnO	.01	.01	.00	.00	.00	.01
CO_2	43.4	41.6	29.00	43.3	39.6	41.8
Total	100.	100.	99.00	100.	100.	100.

1. First limestone above Reeves Limestone M. } Mid. E. edge sec. 25, T. 40 N., R. 42 E.
2. Reeves Limestone Member }
3. Top of Metaline Formation; SW 1/4, NE 1/4 sec. 25, T. 40 N., R. 42 E.
4. Mid. N. edge NE 1/4 Sec. 26, T 40 N., R. 42 E. } Maitlen limestones of uncertain position within formation.
5. SE 1/4 sec. 23, T. 40 N., R. 42 E. }
6. S. edge NE 1/4 sec. 26, T. 40 N., R. 42 E }

[Chemical analyses by rapid rock method; analysts: W. W. Brannock, Paul L. D. Elmore, Paul W. Scott, and Katrine E. White]

The schist that separates the Metaline and unnamed Maitlen limestones is also a fine-grained rock that has textures of both schist and phyllite. It is well bedded and composed mainly of muscovite and quartz. Chlorite is present only in minor amounts, but locall, particularly near its base and top, it is calcareous. The average thickness is about 60 metres.

Thickness of the two schist units and those of the Reeves and unnamed limestone can be measured with fair accuracy, but this is not true for the limestone of the Metaline Formation, which is nowhere present in its entirety. Nor can its total thickness be restored by assembling the faulted segments because of the homogenity of the unit. Its upper contact was seen only on Red Top Mountain and on the north slope of the canyon of O'Hare Creek.

On Red Top Mountain, the Metaline Formation is thinner bedded, finer grained, and contains as much as 30 percent of detrital quartz. The predominant black argillite of the overlying Red Top argillite encloses occasional beds of limestone through a distance of several hundred feet above the contact, but these limestone layers, black and fine grained, sharply contrast with the light colored, coarser-grained beds assigned to the Reeves Limestone Member. The contact between the limestone and the black argillite is conformable as well as sharply gradational, which suggest that the change in depositional conditions, although sudden, does not necessarily represent a hiatus.

The faulted section of Metaline rocks in the Red Top Mountain block lacks the thick middle dolomite member present in all other sections of Metaline Formation. Whether this has been removed by faulting or was never deposited is not known. The absence of any Metaline rocks on the same structure (Columbia anticline) to the north in British Columbia favors the latter.

In the Lime Creek Mountain and Red Top Mountain blocks on the geologic map of the Deep Creek area (Yates, 1964) the Reeves is shown as a multiple limestone unit with interunits of phyllite or schist. At the time this map was made, the relations between the carbonate and phyllite components of the Maitlen Phyllite was not appreciated. After mapping the Northport quadrangle (Yates, 1971) where the Maitlen consists of an alternation of limestone and phyllite units, with phyllite in the minority, it was realized that the Reeves Limestone Member was a name that should apply only to the limestone unit immediately above the quartzite and that the other limestones should go unnamed--at least for the time being. This correlation has been made on figure 2.

Argillite of Red Top

The argillite Red Top is a dark gray to black very fine-grained siliceous rock that ranges in layering from micro-bedded to massive. Some varieties are microquartzites and siltites that have 80 percent or more quartz. Muscovite occurs evenly distributed throughout the rock or is concentrated in certain microbeds. The dark color comes from a finely divided black pigment, probably amorphous carbon. Some beds are medium-grained, slightly dolomitic quartzite. A few thin, fine-grained, dark gray dolomite beds were seen. One thin section showed microlayers rich in apatite, but no systematic sampling was made that would indicate whether such microlayers were common or rare. Pyrite in small amounts is a common constituent, but is seldom preserved in the surface rocks. The unit is estimated to have a minimum thickness of 600 metres.

The Grass Mountain sequence

The Grass Mountain sequence is shown in the explanation for figure 2 as of questionable Carboniferous age, an assignment more permissive than positive for these unfossiliferous beds. A lower Paleozoic age, although unlikely, is not inconceivable and is discussed in a following section. The age assignment depends upon whether it is more closely related to the Red Top argillite, or to the Devonian(?) argillite, of the Lime Creek Mountain block. The basic lithology of all three is quite similar. According to the interpreted movement pattern on the O'Hare and Black Canyon faults, which separate the three units, the sequence has the greatest affinity to the Lime Creek Mountain block and hence to the Devonian(?) argillite. The age assignment to the Carboniferous, therefore depends largely upon the correctness of a structural interpretation.

The Grass Mountain Sequence is divided into five units, but only one, an argillaceous limestone, is distinctly different from the other four, which are dark colored pelites composed largely of quartz and muscovite. These four noncalcareous units could not be differentiated without the minor variants of thin quartzite and conglomerate. These variants make it possible to separate a lower unit of black slate containing thin quartzite beds (Map symbol Cg_1) from the supposedly successively overlying units of argillaceous limestone (Cg_2), black and gray slate unit (Cg_3), black slate and argillite with conglomerate lenses (Cg_4), and siliceous black argillite (Cg_5). This sequence is believed to be arranged chronologically in the explanation of figure 2, but the arrangement is unsupported by fossils or depositional features that conclusively determine the top of beds.

The arrangement shown in the explanation on the previous geologic map of the Deep Creek area (Yates, 1964) is revised in figure 2 and text of this report. What is now labeled Cg_4 was formerly unit Cg_2 and what was Cg_4 is now Cg_2. However, this new arrangement is neither entirely dependable nor very real. The only demonstrable relations are between units Cg_4 and Cg_3 and here it is assumed the section is overturned and Cg_3 is the older. There is an indication of gradation or intertonguing of units Cg_3 and Cg_3 and Cg_2 and, following the relation "established" between Cg_4 and Cg_3, unit Cg_2 is the older. Unit Cg_1's position in the core of the anticline favors its age as greater than Cg_2; Red Top argillite was observed in depositional contact with Metaline limestone of the core sequence and therefore is awarded the basal position. By default, unit Cg_5 is awarded the uppermost position, which is consistent with the predictable relations occurring on an overturned and thrust faulted anticline.

The dark, fine-grained rock that forms the great bulk of the Grass Mountain sequence differs but little from one unit to another. Slate, argillite, and phyllite in varying proportions dominate all but the limestone unit (Cg_2), but even this unit contains minor interbeds of slate. Colors vary from intense black to the much rarer silver gray of sheared and leached rocks. Bedding, commonly obscure, is most prominent in the siliceous black argillite (Cg_5). Composition varies as much within as between units. Quartz and muscovite are the predominant minerals and compose 80 percent or more of the rock; the quartz content ranges from 20 to 80 percent, the high content being typical of the siliceous black argillite (Cg_5). Calcite is commonly absent, but rare beds may have as much as 10 to 25 percent and some of these grade into limestones beds similar to those in unit Cg_2. Chlorite is uncommon and rarely totals more than 10 percent. Pyrite altered to hydrous iron oxide is common. The black finely divided pigmenting material, although a carbon compound, is not graphite because it readily leaches in the presence of oxidizing pyrite. Description of the principal lithology, black lutite, of each unit would be redundant, therefore the units are described below in terms of their variants.

Unit Cg_1, black slate with quartzite beds, is set off from the sequence by the presence of beds of black quartzite. These occur through an exposed stratigraphic thickness of 360 metres. Individual beds range from 15 centimetres to 3 metres thick and are discrete units with very little internal layering. The quartzite is a fine to medium-grained rock that contains 95 percent or more quartz, which occurs as poorly sorted grains in all degrees of roundness. Some grains are recrystallized chert and a few are albite and microcline, but the most common accessory mineral is muscovite, which, however, is nowhere abundant.

The quartz is strained and mortar structures and other cataclastic structures are common. In addition to the quartzite beds, unit Cg_1 also has a few 5 centimetre beds of green gray phyllite that contain chips of black phyllite.

Unit Cg_2 is the limestone member of the Grass Mountain sequence. It is a black fine-grained limestone, poorly bedded and rarely thin-bedded. It contains an estimated 25 to 50 percent of noncarbonate material, which is also fine grained and dominantly sericite and quartz.

The boundaries between unit Cg_2 and adjacent units are faults or are hidden by surficial deposits. The position of the unit in the Grass Mountain sequence is more speculative than factual because of the ambiguities of structure. Outcrops of the limestone in Seriver Creek grade upward into argillite but the lack of outcrop prevents positive identification of the overlying unit as the inferred Cg_3.

The Cg_3 unit, the most widespread unit in the Grass Mountain sequence, is the residium of the sequence; it is what remains after the black argillites that contain conglomerate and quartzite are split off. It is a monotonous unit of fine-grained clastics in the textural and structural form of slates, argillites and phyllites that range from black to medium gray in color. In places it contains a few beds, several inches thick, of black, fine-grained limestone and punky porous beds from which carbonate minerals have been removed by solution. The composition is quartz and muscovite in varying proportions plus minor amounts of chlorite and iron oxides. The fine grained character of the rock prevents greater refinements of the mineralogy through thin section study.

Unit Cg_4, although physically separated from unit Cg_1, is interpreted as younger, but the two are so similar that it is conceivable that they may be lateral variations. Instead of containing beds of fine- to medium-grained quartzite, unit Cg_4 contains beds and lenses of pebble conglomerate which is so siliceous in clast and matrix and so thoroughly cemented that it is itself a quartzite. The principal difference between the two quartzites is grain size, but the pebble conglomerate also differs in containing angular clasts and whisps of black argillite.

The pebble conglomerate beds range from a few centimetres to more than 3 metres in thickness; individual clasts range seriately, from coarse sand to pebbles 3 centimetres in diameter. Contacts between the pebble beds and the enclosing argillite are knife edge. In places the conglomerate contains thin lenses several inches across of black argillite, in addition to the fairly common argillite chips. The thin lenses of argillite may be larger chips elongated by deformation. In addition to the black argillite chips or clasts, which are less than 5 percent of the total volume of the conglomerate, the rock contains several percent of muscovite, chlorite, and hydrous iron oxide. The pebbles are found only in the pebble beds; there are no erratics in the argillite.

The shape of individual pebble beds and the possibility that they have a preferred orientation was not determined. Outcrops are inadequate to determine the continuity of individual beds. It is probable that the pebble beds are fillings in channels cut in the argillite and are linear elements. Unit Cg_4 appears to be about 750 metres thick.

The Cg_5 unit, the siliceous variant of the Grass Mountain sequence occurs not only in the Stone Mountain area, but also on Red Top Mountain. It is similar to the black argillites of Russian Creek in the Metaline district that are considered by Dings and Whitebread (1965, p. 26) as belonging to the Ledbetter Slate. The argillites in the Russian Creek area are described as thin-bedded rocks composed of fine quartz grains in undetermined argillic material. In the Deep Creek area the argillic material is finely divided sericite and quartz. It is estimated that many beds are more than 80 percent silica; they approach chert in composition and appearance, but thin sections clearly show that the rock is clastic in origin.

The Cg_5 unit on Red Top Mountain is shown on the preliminary geologic map of the Leadpoint quadrangle (Yates and Robertson, 1958) as two units, "bsp" and "bpc." Unit bpc is described in the map explanation as "Sheared and contorted dark gray phyllite; may be a highly deformed phase of bsp." In this report unit bpc is accepted as a "highly deformed phase of bsp" and accordingly is not separated. Much of unit Cg_5 on Stone Mountain is also highly deformed cataclastic rock, particularly that near the Columbia fault.

The Pend d'Oreille block

The Pend d'Oreille block is that part of the Deep Creek area that lies north of the Columbia fault. It in turn is bounded on the north by the Waneta fault, which loops south of the International Boundary for only a short distance (figure 14). The small area in the United States that lies north of the Waneta fault is not given a "block" name but is part of a large area of Jurassic volcanic rocks, the Rossland Group, that is extensive in British Columbia, and is called in this report, the Jurassic volcanic province.

The rocks of the Pend d'Oreille block are argillites, limestones and minor chert and greenstone of the Pend d'Oreille sequence. The Sheppard granite, which intrudes the sequence is described in the section on structure.

The layered rocks of the Pend d'Oreille sequence (figure 2) are divided into three units: a lower unit of dark gray argillite interlayered with limestone (Cp_1), an intermediate unit, mainly phyllite and argillite, but including greenstone and chert (Cp_2), and an upper unit composed of a distinctive black silty argillite (Cp_3). The upper unit, Cp_3, which occurs only on the west side of the Columbia River has not been previously described, but subunits that comprise the other two units are shown on the geologic map of the Leadpoint quadrangle (Yates and Robertson, MF 137, 1958). These local subunits are not shown on figure 2, being unidentified parts of Cp and Cp_2 Figure 4 indicates correlative map symbols for figure 2 and MF-137.

Although the Pend d'Oreille sequence contains rocks similar to those in the Grass Mountain sequence, the gross characters differ. Fine-grained clastics predominate in both sequences, but the mineralogy differs with muscovite dominant over chlorite in the Grass Mountain sequence and chlorite over muscovite in the Pend d'Oreille sequence. Except for rocks in the northeast corner of the block (fig. 2)--which may more properly belong to the Grass Mountain sequence--they contain less carbonaceous material and are lighter in color, commonly having a greenish cast.

The lower unit, Cp_1, is a black argillite with interlayered black, fine-grained limestone. The argillite superficially resembles the argillites of the Grass Mountain sequence but is distinguished under the microscope because chlorite dominates over muscovite. The limestone beds, which may compose 25 percent or more of the unit, range from ½ centimetre to 1 metre in thickness; most beds are between 1 and 5 centimetres thick. The limestone comparable to the argillite is very fine grained and contains as much as 50 percent fine quartz, chlorite and sericite, as well as the usual black pigmenting material.

The impurity of the limestone is noticeable in outcrop because the limestone beds weather only slightly more rapidly than the enclosing argillite. Their lighter gray tone gives the rock a faint banding. A more pronounced banding is exhibited by the limestone pods, the larger of which are shown on figure 2 as subunits within unit Cp_1. These are medium to light gray, thin-bedded limestones containing only small amounts of quartz and mica. Just south of the Waneta fault on the west side of the Columbia River at the 1,480 foot contour is an outcrop of this limestone that is too small to separate on figure 2. This outcrop is bioclastic limestone that contains fossil "trash" among which are crinoid columnals. According to J. Thomas Dutro, Jr. (written commun., February 25, 1957) who examined material from this locality, "This collection is composed mainly of crinodal debris, considerably recrystallized, but recognizable. There are a number of columnals and fragments with nodose ornamentation. There are a number of columnals and fragments with nodose ornamentation. The age is Paleozoic....." The limestone in Cp_1, sec. 15, T.40N., R.41E., also has poorly preserved crinoid columnals and may be the same horizon.

Unit Cp_2 succeeds unit Cp_1 as an abrupt change from black argillite to an assemblage of green gray chert, greenstone, phyllite and quartzite in what appears to be a conformable succession. Unit Cp_2 also includes a thin layered gray and white banded limestone about 60 metres thick. Rock types in the Cp_2 unit include: (1) chlorite schist interlayered with thin quartzite beds, (2) greenstone, (3) metatuffs, (4) chert, and (5) limestone.

The greenish gray chlorite schist (and phyllite) and interlayered quartzite is the most abundant of the five rock types. It is at least 300 metres thick on the north slope of Red Top Mountain where it is best developed. The ratio of quartzite to schist is about two to one. The quartzite is a medium-grained, light gray rock that has pronounced rehealed cataclastic texture. In places it contains as much as 25 percent albite and minor amounts of muscovite. The foliated layers between the quartzite beds range in grain size from fine-grained schist to phyllite. Beds commonly consist of 50 percent chlorite, 20 percent calcite, and 30 percent quartz and albite. The presence of chlorite is reflected in the gray green color.

The greenstone occurs as both sheared and massive varieties. The sheared variety is chlorite phyllite that originally may have been pyroclastic rocks. The massive variety is highly altered basaltic lava that appears in thin section as felted mats of indeterminable feldspar--probably albite--set with phenocrysts of uralitic hornblende that have rare unaltered cores of monoclinic pyroxene. Throughout the thin sections are irregular areas of chlorite, calcite, epidote and quartz, the common metamorphic products of the conversion of a basaltic lava to a greenstone.

Accompanying the greenstone are very fine-grained, thinly laminated, light green rocks composed mainly of epidote, chlorite, and clinozoisite; these are believed to be falls of mafic ash.

In outcrop the cherts are difficult to distinguish from the metatuffs. Both have a fine-grained texture and light green color. In thin section they are seen to be thoroughly recrystallized to a fine-grained crystalloblastic mosaic of quartz grains. Some contain flecks of muscovite and veinlets and irregular areas of carbonate minerals. In one thin section rhombs of dolomite containing small partly replaced grains of quartz were clearly crystalloblastic in the quartz mosaic. Most chert is the ribbon variety that has thin partings of phyllite between the chert layers.

The Cp_2 unit contains at least one limestone that may have been a widespread part of the unit. It outcrops on the hill west of Hungate Clearing (sec. 7, T.40N., R.41E., and is probably the same limestone that outcrops near the Ranchview mine in sec. 2, T.40N., R.40E., where it is about 60 metres thick. Near Hungate Clearing it appears to be much thicker probably as a result of duplication by folding. It is a banded light and dark gray limestone, whose individual bands range from less than 1 millimetre to more than 10 mm. Except where recrystallized, it is a very fine-grained, and moderately pure. No fossils have been found. A few much smaller exposures of limestone probably indicate other more discontinuous limestone horizons.

Unit Cp_3 occurs only in the northwest corner of the map area (see fig. 2). It is separated from units Cp_2 and Cp_3 by Quaternary unconsolidated deposits. No fossils have been found in it; hence including it in the Pend d'Oreille sequence and assigning it even a questionable upper Paleozoic age is mainly a matter of convenience. It is included here because it resembles none of the rocks older than the Pend d'Oreille sequence. The unit is composed of fine-grained gritty slates and argillites. Bedding, although faint, is well developed and apparent through slight variations in grain size from bed to bed. The rock differs from all others in the Deep Creek area by the presence of flakes of detrital mica and grains of quartz that stand out from a much finer-grained matrix. The minimum thickness is greater than 300 metres.

Rossland Group

Rocks of the Rossland Group outcrop in two small areas north of the Waneta fault along the International Boundary. Outcrops of these rocks are too small and too incompletely representative of the group to merit more than a very brief description. Rossland rocks, however, cover large areas immediately north in British Columbia, where they have been studied by H. W. Little (1962) and other Canadian workers. The Rossland Group, as most recently defined by Frebold and Little (1962), consists of a lower sedimentary unit, the Elise Formation, a middle, predominantly volcanic unit, and the Hall Formation, which together form the lower Rossland Group. The upper Rossland Group includes both volcanic and sedimentary rocks. Ammonite faunas, ranging in age from lower Sinemurian to early middle Bajocian bracket the age of the Elise and Hall Formations; the upper Rossland has not yet yielded fossils. The Rossland Group and the underlying Archibald Formation is a eugeosynclinal volcanic suite representing in the Salmo-Rossland area two strong volcanic pulses characterized by andesitic and basaltic flows and pyroclastic rocks and an intervening non-eruptive period when marine clastic sediments were deposited. Many of the sedimentary rocks are tuffaceous.

In the Deep Creek area lavas predominate over pyroclastic rocks. The lavas are dark green, massive to porphyritic rocks that have the typical chlorite-actinolite alteration that characterizes greenstones of the greenschist facies. In thin section it was not possible to determine whether the rocks were originally basaltic or andesitic because of the alteration of the plagioclase. Some flows have abundant plagioclase phenocrysts, others have abundant monoclinic pyroxene. Commonly the flows are in part flow breccias, which are difficult to distinguish from the much rarer pyroclastic rocks.

Northport block

Only the extreme northeastern end of the Northport block extends into the Deep Creek map area. The isolated exposures near the mouth of Deep Creek that are shown on figure 2 as Gypsy quartzite, Metaline Formation, and Ledbetter Slate are stratigraphically and structurally uninterpretable without knowledge of the geology to the southwest, which is described in the regional section of this report. The Northport block is a northeast-trending anticline that has been badly broken and distorted by stresses applied across the axis of the structure. This anticline and the rocks within it are described in a latter section. Although no fossils have been found block, comparison of lithologic sequences places these rocks in the Paleozoic. They are most closely resemble rocks of the Red Top-Grass Mountain block.

Rocks south of the Spirit pluton

The rocks within the part of the Deep Creek area south of the Spirit pluton lack stratigraphic and structural unity. Three blocks appear incompletely represented; two are segments of blocks that lie north of the Spirit pluton and the other is the northern extremity of a block that extends far to the south into adjacent quadrangles.

The rocks south of the Spirit pluton are metamorphosed to the amphibolite facies accordingly fossils are not to be expected. Because the metamorphism was accomplished without strong deformation, bedding is well preserved and in places, actually enhanced by the metamorphism. Bedding is paralleled by the mimetic foliation of the pelitic rocks.

The rocks south of the Spirit pluton are described under four headings: (1) Precambrian and Cambrian(?) rocks of the Ione Hill and Huckleberry Mountain areas,[2/] (2) Cambrian and Ordivician rocks of the Magma mine area, (3) Cambrian rocks on the north slope of Mount Rogers, and (4) Cambrian rocks of the Sierra Zinc mine area. These four areas are separated by faults. The rocks[2/] of the Sierra Zinc mine area are related to rocks of the Lime Creek Mountain block; those of Mount Rogers and the Magma mine areas are related to rocks of the Hooknose-Baldy blocks.

[2/] On Map I-412 (Yates, 1964) these rocks are shown as Precambrian in age, an assignment the writer no longer considers valid.

Precambrian and Cambrian rocks of the Ione Hill and Huckleberry Mountain areas.--The rocks of Huckleberry Mountain and Ione Hill areas have been arranged in the explanation of figure 2 in two sequences: (1) the Byers Creek sequence, which is in the Huckleberry Mountain area and (2) the Ione Hill sequence. The Ione Hill sequence questionally regarded as, the younger. The two sequences are separated by a fault which cuts out an unknown thickness of section. Both sequences have similar pelitic schist lithologies, with a little quartzite and limestone. The Byers Creek sequence, however, is distinguished by a unit of hornblende-plagioclase schist, undoubtedly of volcanic origin. This unit is considered equivalent to the Leola Volcanics of the Metaline quadrangle (Park and Cannon, 1943, p. 9-11) and the equivalent Huckleberry Greenstone of the Chewelah quadrangle (Bennett, 1941, p. 8).

In the Metaline quadrangle the Leola Volcanics are overlain by the Monk Formation, basically a phyllite unit, and rest unconformably in the southern part of the quadrangle on the Priest River Group, a "complex sequence of metamorphic rocks that includes phyllites and schists, limestones, dolomites, quartzites, and volcanics." In the Chewelah quadrangle (Miller and Clark, 1975) the Huckleberry Greenstone is unconformably overlain by either the Addy Quartzite (equivalent to Gypsy Quartzite) or by an intermediate unit that is considered a thin version of the Monk Formation.

To extend any of these names into the area south of the Spirit pluton, with the exception of Leola Volcanics, is, perhaps, premature until the area south of the Deep Creek area is mapped in detail. Although the correlation of the hornblende schist unit with the Leola Volcanics is valid, the formalization of this correlation automatically enhances the credibility of correlations of rocks overlying and underlying the hornblende schists with those in neighboring areas. The schists overlying the hornblende schist would be, on the basis of stratigraphic position, an abnormally thick Monk Formation; however, the work of Gerald W. Thorsen (19) to the south of the Deep Creek area indicates that a fault separates the volcanic unit from the overlying rocks. The nearest Leola Volcanics are to the east in the southern part of the Metaline quadrangle where they are likewise separated by a fault (Harvey fault, a west dipping reverse fault) from the overlying rocks. The overlying formation in this case is the Maitlen Phyllite, which in many respects resembles the Byers Creek and Ione Hill sequences, the rocks overlying the volcanic unit in the Deep Creek area. It is possible that the fault mapped by Thorsen is an offset extension of the Harvey fault and that the overlying rocks are the Maitlen Phyllite. This correlation with the Maitlen Phyllite is supported by the tentative conclusion reached during reconnaissance mapping in the Aladdin quadrangle where the rocks overlying the volcanics project into similar rocks that are demonstrably Maitlen Phyllite. A corollary of this conclusion is that the rocks underlying the volcanic unit may belong to the Priest River Group of the Metaline quadrangle (Park and Cannon, 1943, p. 6).

The Byers Creek sequence is predominantly shale metamorphosed to the hornblende hornfels facies by heat from the Spirit pluton and Kaniksu Batholith (see Index map, Fig. 2). The sequence is subdivided into three units: (1) a lower sedimentary unit composed of quartz mica schist interlayered with dolomitic limestone and a quartzite, probably equivalent to the Priest River Group, (2) an intermediate volcanic unit of hornblende-plagioclase schist, and (3) an upper and much thicker sedimentary unit of quartz mica-andalusite schist containing thin interlayers of quartzite and limestone probably equivalent to the Maitlen Phyllite. The sequence is at least 3,350 metres thick.

The Kaniksu batholith engulfed all but several hundred feet of the lower unit of the Byers Creek sequence. On Huckleberry Mountain, it consists, from the top to the intruded base, of a thin biotite schist about 30 metres thick, a white to light gray, bedded quartzite about 75 metres thick and about 150 metres of mica schist interlayered with limestone. The schists are similar to those in the upper part of the Byers Creek sequence described below; the limestones are white to light gray, fine- to medium-grained, medium bedded and dolomitic, as evidenced by abundant diopside; the quartzite is medium- to coarse grained, white to light gray, and contains a few percent of muscovite.

The middle unit, the hornblende-plagioclase schist is a dark green fine- to medium-grained rock commonly laminated with thin white layers or spotted with circular white areas a few millimetres in diameter. The rock is predominantly green hornblende, substantial plagioclase (An_{40}), and lesser amounts of quartz and rare biotite. It is locally veined by epidote and clinozoisite. A strong to weak foliation is emphasized by the tendency of the plagioclase to concentrate in layers. Nests of plagioclase are what give the rock its spotted appearance.

The rock is thoroughly recrystallized, having no trace of an original igneous texture, although its composition clearly indicates an igneous, probably basaltic, origin. Its foliation and plagioclase-rich lamellae lie parallel to the bedding planes of adjacent sedimentary rocks, therefore the recrystallization is mimetic. The compositional layering could be either controlled by the primary bedding structure of a tuffaceous rock and pseudomimetic, or the product of metamorphic differentiation controlled by cleavage that developed parallel to the bedding planes of the rocks that enclose the volcanic unit. The interpretation that the compositional layering represents bedding is supported by the difficulties inherent in applying the process of metamorphic differentiation in the field of thermal metamorphism, and is more positively supported by the occurrence of well preserved bedding in the limestones and quartzites of the same sequence.

Above the volcanic unit is about 3,000 metres of schist with thin layers of quartzite and limestone in the upper part. The bulk of the unit is a biotite schist that contains variable amounts of andalusite. Some layers have an estimated 30 percent or more of andalusite--particularly in the lower part of the section--others have less than 10 percent. The schist is brown to gray with both massive and thin bedded varieties. The more massive varieties are less schistose and have more abundant porphyroblasts of andalusite, which swarm with fine inclusions, largely quartz and biotite. The matrix for the porphyroblasts is a fine-grained mixture of quartz and brown biotite. Locally along the contact with the pluton the rock is gneissic; microcline, sillimanite, and cordierite are prominent.

The upper part of the sequence is both more limy and more siliceous. Discrete zones of thin layered limestones and quartzites a few feet thick are interlayered with the schist. Neither limestones nor quartzites are as pure as those of the Cambrian Gypsy quartzite.

Only one limestone unit is thick enough (30 m) to show on the geologic map (pl. 1); it is easily seen at an elevation of 4,900 feet on the road to the Huckleberry Mountain lookout. This is a very thin-layered, almost microlayered rock, white to pale green in color. It is largely calcite, but contains as much as 20 percent quartz and some varieties contain hornblende, zoisite, and garnet, as well as biotite.

Rocks of the Ione Hill sequence are very similar to those of the Byers Creek sequence; they have the same predominance of andalusite-biotite schist, contain similar thin layers of limestone and a unit of gray quartzite. It was not however, possible to match any quartzite or limestone unit with any in the Byers Creek sequence.

Cambrian and Ordovician rocks of the Magma mine area

Between the Ione Hill rocks and the Spirit pluton is a narrow belt of unfossiliferous limestone, dolomite, phyllite and black slate that resemble the Maitlen, Ledbetter, and Metaline Formations to the north of the Spirit pluton. The rocks are believed to belong to the assemblage of the Hooknose-Baldy block, but the evidence for this is largely inferential. Breccias similar to the intraformational breccias of the middle dolomite unit of the Metaline Formation on Gladstone Mountain occur sporadically in the dolomite in this area and are the strongest support for the correlation. Representatives of the lower limestone and the middle members of the Metaline Formation, the Maitlen Phyllite, and the Ledbetter Slate were identified.

Although these rocks have been as intensely metamorphosed as the Byers Creek and Ione Hill sequences, they reacted differently, mainly because of compositional differences. The shale facies of the Maitlen Phyllite is converted to a quartz-biotite hornfels, the lower limestone member of the Metaline Formation near the granite contact to a calc-silicate hornfels, the middle dolomite member is converted to a coarse-grained magnesian marble and wherever silica was available to tremolite, diopside, or diopside-forsterite rock, but the Ledbetter Slate remains a dark gray, fine-grained rock that shows very little megascopic evidence of metamorphic change. Exposures are too discontinuous and too small and the sample too incomplete to reconstruct a stratigraphic section, or to make positive interpretation of structure.

Cambrian rocks of the north slope of Mount Rogers

In the extreme southwestern corner of the Deep Creek area are Cambrian rocks representing the Gypsy, Maitlen, and Metaline Formations. These rocks are the northerly termination of a northeasterly trending anticline that extends as far south as Colville (map by W. A. G. Bennett, plate 6 in Mills, 1962). They are believed to belong to the same stratigraphic assemblage as those of the Hooknose-Baldy block. They fall within the hornblende hornfels facies of thermal metamorphism.

Gypsy Quartzite

The Gypsy Quartzite exposed on Mount Rogers, includes both schist and quartzite, which were mapped separately. The schists lie immediately beneath the Reeves Limestone Member. The boundary between schist and quartzite is moderately sharp, although a few schist layers are interbedded with the uppermost quartzite and a few quartzite layers are present as boudins in the lowermost schist. All but the upper 45 metres of the lowermost quartzite unit is cut out by a fault. This lower unit is overlain successively by 180 metres of schist, 640 metres of quartzite, and the uppermost 300 metres of schist.

The mineralogies of the schists are similar to those of the Precambrian rocks, mainly quartz, biotite, and muscovite, with variants containing andalusite, or, where adjacent to the pluton, sillimanite and cordierite. They are fine to medium-grained rocks, gray to brownish gray in color. The quartzites are gray to white, medium bedded to massive, medium to coarsely crystalline rocks.

The white coarsely crystalline quartzite about one mile west of the Sierra Zinc mine is believed to belong to this assemblage. This quartzite has no schist partings and being in contact with granite on three sides is completely recrystallized.

Maitlen Phyllite

The Reeves Limestone Member of the Maitlen Phyllite on Mount Rogers is a single unit, as it is in the Hooknose-Baldy block north of the Spirit pluton. From a thickness of 90 metres at the south boundary of the quadrangle it pinches northeastward to 30 metres. This thinning may be either depositional or tectonic.

The limestone is a medium-bedded, medium-grained, light gray to buff colored rock. It is slightly dolomitic as indicated by the presence of small amounts of diopside in the presence of free quartz.

In this area most of the Maitlen Phyllite consists of schist, rather like those in the Precambrian and in the Gypsy Quartzite. All in the map area belong to the amphibolite hornfels facies of thermal metamorphism. Two varieties are present: (1) Biotite andalusite schists from alumina-rich shale and (2) biotite-quartz schist from the more siliceous shales. Interbedded with the schists are a few 3-metre-layers of buff limestone.

Metaline Formation

The Metaline Formation is represented by two mappable units that are equivalent to the lower limestone member and the middle dolomite member. The lower limestone member is an impure dolomitic limestone with thin siliceous dolomite interlayers. It is a faintly greenish, cream colored, medium-grained rock having dark brown streaks. The siliceous interlayers are thin, fine-grained rocks composed of diopside and quartz.

The middle dolomite member is a white to light gray, medium to coarse grained marble with poorly preserved bedding. It is composed chiefly of dolomite, diopside, and forsterite. The lead-zinc ore that occurs in prospects along the southern margin of the Spirit pluton, and at the Van Stone mine, a mile west of the quadrangle is in this unit.

Cambrian rocks of the Sierra Zinc mine area

The preliminary map of the Deep Creek area (Yates, 1964) shows the rocks about the Sierra Zinc mines to be Precambrian. Since the publication of that map, the evidence for correlations was reassessed and it was concluded that this group of rocks is most logically correlated with the Cambrian rocks that underlie that thrust plate on Lime Creek Mountain. Under this new interpretation, the Blue Ridge sequence of quartzite, schist and limestone, becomes the Gypsy Quartzite and the Maitlen Phyllite. The thickness is estimated to be in excess of 900 metres.

Only the uppermost 150 metres of the Gypsy Quartzite is exposed in the Sierra Zinc mine area. They are fine- to medium-grained, gray to dark brown quartzites interlayered with lesser amounts of quartz-mica schist. The lowermost exposed beds are massive to thick, light gray to white quartzite. The purity of the quartzite increases downward. The purest quartzite is a white, vitreous rock consisting of interlocking quartz crystals with less than one percent of muscovite and microcline. As the quartzite becomes less pure grading into schist, the amounts of biotite, muscovite and microcline increase to a point where the platy minerals dominate over the granular. The interlayered schists are similar in appearance and mineralogy to those associated with the overlying limestone units.

The rocks that lie above the quartzite can be divided into two groups. The lower group is broken by faults and is poorly exposed; information concerning it comes from small and scattered outcrops, mine workings and drill logs. The upper group is much better exposed and has more discrete lithologic units. According to a composite section assembled by Charles D. Campbell (manuscript report in U.S. Geological Survey files), the rocks in the lower group are schists, chlorite phyllites (chlorite schists in this report) and one marble unit 60 metres thick, another 27 metres thick, and three, 4 ½ to 6 metres thick. The larger limestone units contain interlayered impure quartzite and some beds are sandy, as attested by diopside-bearing varieties. None of the marble units are pure carbonate and some schist is calcareous. Although the carbonate rocks are limestones, the abundance of magnesium silicates suggests that dolomitic phases were present before metamorphism.

In contrast to the limestones of the lower group, those of the upper group are pure and lack quartzite intercalations. They are médium- to coarse-grained recrystallized limestones whose bedding is moderately well developed. They range in color from white to dark gray. The schist units that separate the limestone units are largely biotite quartz schist, and include varieties that contain combinations of andalusite, sillimanite, cordierite, microcline, and plagioclase. The schists near the Sierra Zinc mine have a higher feldspar content than those in surrounding areas. Although this may in part reflect original composition, at least some feldspar was metasomatically introduced, as common cross cutting replacement veinlets of microcline indicate.

The only unfaulted section of the upper group of limestones is in the northwest quarter of sec. 30, T.38N., R.41E. This 550 metre section from bottom to top is divided as follows: 90 metres of schist, 115 metres of limestone, 90 metres of schist, 75 metres of limestone, 30 metres of schist, and 150 metres of limestone, and an unknown thickness of schist.

Structure

Introduction

The procedure used in the preceding section, block by block description, is also used to describe the structure. Although this section is primarily concerned with the description of structures within the limits of the Deep Creek, area, it is to the reader's advantage to be aware of the regional tectonic history, of which these structures are a part. Accordingly, the section begins with a summary of this history, not all of which could have been learned in the Deep Creek area--although most events can be recognized in retrospect--but comes from studies made in surrounding areas, where recognizable chapters of the tectonic history are preserved. In a later section of this report the tectonic history outlined below is developed in terms of both stratigraphy and structure. Although some features such as the Brodie-Sullivan kink fold and décollement faults are important structural features in the Deep Creek area, they are regional in nature and are discussed in the section on regional geology.

Precambrian rocks indigenous to the Deep Creek area, either exposed or buried, have undergone at least six tectonic events of greatly different intensities. Three of these events are epeirogenic, the fourth involves intense folding, the fifth, crossfolding, and the sixth and last, block faulting without folding. These events are dated with varying degrees of precision. The two epeirogenic events of the Precambrian, the first gentle folding at the beginning of Windermere time (Late Precambrian) and the other, high angle faulting and volcanism in mid-Windermere time, did little to deform or metamorphose the rocks. The third epeirogenic event consists of uplift of the panhandle of Idaho and adjacent Montana and westward décollement thrusting of essentially unfolded lower Paleozoic rocks. The décollement faulting is inferred to explain anomalous rock distribution and cannot be accurately dated. It occurred sometime after the Devonian and before the Jurassic. A late Paleozoic age is favored.

The principal period of folding also cannot be dated with precision. The northeasterly trending folds produced by this event dominate the structure of the Deep Creek area; only locally are these folds destroyed by later deformations. To the northeast in British Columbia, rocks as young as Middle Jurassic are folded along northeast axes; however, the folding recorded in Jurassic rocks may not represent the total length of time that northeast trending folds were developing in the Paleozoic rocks. In the Deep Creek area, the 100 million year old Spirit pluton clearly crosscuts the folds and thus puts an upper limit on their age. A mid- to late Jurassic age is believed most probable but earlier Late Triassic folding along similar axes is a possiblity. It would be convenient if the folds could be called Jurassic or Nevadan, but until their age is more precisely established we have to be satisfied with "Arc folds," a name here introduced in reference to their importance as the defining structure of the Kootenay arc, the northeasterly trending segment of the Cordilleran fold belt, in which the Deep Creek area is located.

From the time of Park and Cannon's work in the Metaline quadrangle on deformed cleavage in the Maitlen Phyllite (1943, p. 16) it has been known that the deformation in the Kootenay arc is polyphase and that a later deformation is superimposed on the northeast folds. Bowman (1950) recognized this in the Orient quadrangle, which lies at the western margin of the arc. The nature of this superimposed deformation was not clearly understood by the writer until he studied structures in the Northport quadrangle (1971). In the Northport quadrangle the northeast (arc) folds when subjected to compression across their axes were buckled into kink folds and broken by south and north dipping thrust faults, both of which are of near east-west trend. Hereafter these near east-west folds are referred to as crossfolds and the event, as "the crossfold event." These buckles and thrusts were contemporaneously torn by "tear faults" that broke across the beds and through the thrust faults, allowing some segments to move further northward than other segments. The tear faults, most of N.10°E. trend, may in part be along shear planes that extend into the crystalline basement rocks, because the last event, that of block faulting, was dominated by high angle faults of similar trend, but without measurable horizontal movement. This high angle, block faulting during the Eocene was accompanied, and followed by, the extrusion of volcanic rocks. The emplacement of plutonic rocks occurred as early as Triassic and as late as the Eocene.

All the above structural history was not interpreted from outcrops in the Deep Creek area, where neither the earliest nor the latest events can be dated. Part of the difficulty in resolving this structural history came from the isolation of outcrops. Areas critical in establishing the relations between structures or between structural units, are commonly covered by alluvium or glacial fluvial deposits, consequently the hidden relations must be inferred or interpreted from the better exposed areas. Fortunately, there are enough of these to establish the structural pattern and nature of fault intersections, which is illustrated in figure 5A by the use of structure contours without numerical values. More freedom of interpretation is expressed in this map than in figure 2.

Although effort was made to observe and record as many minor structures as possible without sacrificing other aspects of the mapping, the coverage is not uniform. Perhaps the most meaningful structures that were recorded are minor folds, but unfortunately these are almost entirely restricted to the thin-bedded and fine-grained pelitic rocks. The more massive units, the quartzites and carbonate rocks, are almost everywhere free from wrinkles. An exception is the metamorphosed rocks south of the Spirit pluton, which although thin bedded and fine-grained before thermal metamorphism, are almost free of minor folds. This, however, is possibly the result of the tectonic position of the area within the stress field and not a different reaction to stress.

Figure 5A.--Structural geology of the Deep Creek area as represented by structural form lines. The form lines are linear representation of bedding planes without corrections for topography. Relative steepness of structural slope is indicated by spacing of form lines: the closer together, the steeper the slope. Some of the more interpretive structures here are not shown on figure 2.

Cleavage, particularly well developed in the fine-grained rocks, occurs in most rocks, but unfortunately it was not recorded consistently by all workers who contributed to figure 2. Many outcrops exhibit multiple cleavages; in fact, in some outcrops the strikes appear to "box the compass" and defy interpretation. In some cases, only the dominant cleavage was recorded. In most outcrops, cleavage and bedding coincided.

On the maps and in the text of this report foliation is the term used for a megascopic parallel fabric of oriented planar minerals whose growth was metamorphically controlled. Foliation may parallel bedding or cleavage or both, but is used here to refer to only the more strongly metamorphosed rocks, such as schists.

The "rib and groove" lineation seen on figure 2 is a structure particularly well developed in the argillites of the Red Top-Grass Mountain block. It consists of a series of flat topped, vertical walled ridges alternating with parallel and similar shaped grooves, both lying along the bedding plane. It is interpreted to represent movement direction parallel to the ridges and grooves.

Hooknose-Baldy block

The Hooknose-Baldy block, bounded on the northwest by the Leadpoint fault and on the southwest by the Spirit pluton, extends far beyond the eastern boundary of the Deep Creek area into the Metaline quadrangle, where it was called the Flume Creek-Russian Creek block by Park and Cannon (1943, p. 28). The structure of this block is dominated by the Hooknose anticline, (fig. 6) a prominent fold of northeast strike, which can be identified on figure 2 by its core of Gypsy Quartzite. The Hooknose anticline is broken into two parts by the Ridge fault, which cuts diagonally across the axis of the fold. In the Deep Creek area, that part of the fold on the southeast, upfaulted, side of the fault is tight and arcuate; that on the northwest side is broad, blunt nosed and fragmented by normal, reverse, and thrust faults.

The northwestern segment of the Hooknose anticline contains rocks ranging stratigraphically from the Maitlen Phyllite to the Ledbetter Slate and both upper and lower plates of the décollement thrust. The thrust lying at the surface within the lower limestone unit of the Metaline Formation. The location and character of this fault (surface traces of which are indicated on figure 2 by a double fault line) was not recognized at the time the geologic map of the Deep Creek area (fig. 2) was published (Yates, 1964), because it was not realized that the juxtaposition of unlike facies occurred before folding; consequently faults that are here interpreted as thrust faults were indicated as high angle reverse faults contemporaneous with the arc folding.

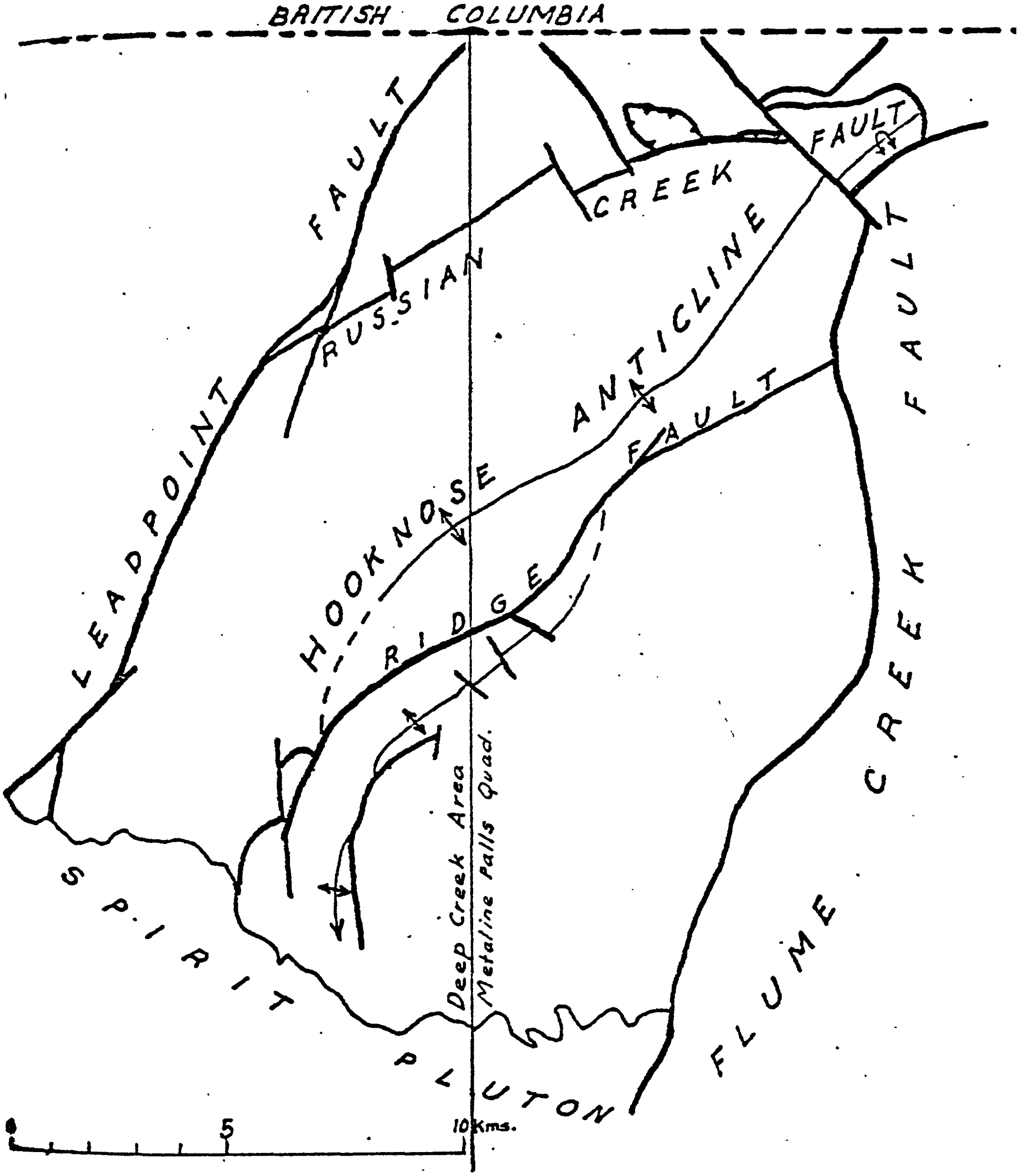

Figure 6. --Sketch of the Hooknose-Baldy block showing the location of the axis of the Hooknose anticline and its possible extension southeast of the Ridge fault.

The blunt nose of the northwest segment of the Hooknose anticline is believed to have resulted from compression applied on the arc folds in a north to northeasterly direction during the crossfold event. This resulted in down buckling and creating a nose on the fold, accompanied by probable strike-slip movement on the Ridge fault. The thrust fault that parallels the flattened nose of the fold east of Deep Lake may be contemporaneous.

The northwest segment of the Hooknose anticline is flanked by a faulted syncline, which crudely parallels the trace of the anticline from the head of Hartbauer Creek to the mouth of Republican Creek. What was once a continuous structure is so broken by faults that it is hardly recognizable as a syncline. The southwest part is identified by down-folded, downfaulted Ledbetter Slate; the northeastern part is recognized by facing dips. The two sets of dips are separated by a northeast trending fault. The beds northwest of the fault all dip to the southeast, an attitude, which on an earlier version of figure 2 (Yates, 1964) was represented as overturned, an interpretation now regarded as wrong and inconsistent with the movement direction on the Russian Creek fault.

The northeast trending faults, in particular those along the north-west flank of the Hooknose anticline, may be products of the arc folding, representing failure of the beds of fold. One fault, the Russian Creek fault, may be as old as the arc folding and as young as the Eocene mafic dikes that locally fill it.

Cleavage, so well developed in the Maitlen Phyllite, is, for the most part, an axial plane cleavage related to the Hooknose anticline. The average strike is N.60°E. and dip, 35° to the SE. In the Maitlen Phyllite, near the Ridge fault, an additional cleavage, averaging N.10°W. at 15° to 20° to the west is prominent. It is possible this is related to the crossfolding and movement on the Ridge fault.

The southeast segment of the Hooknose anticline contrasts with the northwest segment by being the tighter more compressed part of the fold, by the lack of an overturned northwest limb, by having a blunt nose, and by the presence of auxiliary synclines on both flanks. The trace of the axis of the southeast segment of the Hooknose anticline changes through a rotation of 80 degrees from a trend of S.50°W. at the east edge of the quadrangle to S.30°E. where it is lost in the faulted rocks that border the Spirit pluton (fig. 6). Beds of the lower limestone unit of the Metaline Formation that are on the northwestern limb of the anticline are bent around to strike N.60°E., a rotation of 110°. This bend in the anticline is a result of the cross buckling that produced the kink fold discussed in the section on regional structure.

The Ridge fault (fig. 6), which divides the two segments of the Hooknose anticline, is contemporaneous with the late stage of arc folding and accordingly should be bent around with the anticline, but unfortunately, it was not possible to positively identify the fault through the zone of bedding. It can be traced southwesterly for about one kilometre from where it crosses Republican Creek; further to the southwest of the course and nature of the fault is uncertain. In general, it is high angle angle reverse fault whose southeast side moved up during the arc folding and whose northwest side moved northeast during the cross folding.

The Russian Creek fault separates the lower limestone unit of the Metaline Formation from the dolomite intraformational breccia unit. It is interpreted to be a high angle reverse fault dipping to the southeast. It has much less stratigraphic displacement in the Deep Creek area than it has 80 kilometres to the northeast in the Metaline quadrangle, where it separates Gypsy Quartzite from limestone of the upper unit of the Metaline Formation. It is possible that the Russian Creek fault has moved during the arc folding and later during the Tertiary normal faulting; it is discussed more fully in a later section of this report.

The two segments of the Hooknose anticline appear to have reacted differently to the stress that was applied during the cross folding. The southeast segment, containing the core and southeast limb of the anticline bent freely; bedding, cleavage, and fold axes were reoriented to a stable position in the new stress field. The northwest segment, containing the northwest limb of the fold, for some undetermined reason was unable to bend freely. Perhaps this was because the northwesterly dipping bedding planes were poorer planes of movement than the southeasterly dipping bedding and cleavage; or perhaps it was because the Ridge fault permitted the northwest segment of the anticline to move northeastward without bending or kinking but by down buckling of the fold to produce the flattened nose; or perhaps adjustments were more easily made by thrusting.

Lime Creek Mountain block

The Lime Creek Mountain block is the overturned southeast dipping limb of a northeasterly striking fold. Erosion of the fold has exposed a section of Maitlen Phyllite, Metaline Formation and Ledbetter Slate that closely resembles the section of these formations in the Metaline quadrangle and contrasts with the section in the Hooknose-Baldy block. Also present is an assemblage of Gypsy Quartzite and overlying limestones and phyllite that closely resembles the section that occurs in the Red Top Mountain block. These rocks occur in two discrete areas which are shown on figure 2 as bounded by a thrust fault which may be part of the overturned lower plate of the décollement thrust present in the Hooknose-Baldy block.

The hook structures and bent beds of the crossfold event that occur along the south margin of the Hooknose-Baldy block also occur along the south margin of the Lime Creek Mountain block. The change from northeast to northwest strike is associated with a synform at the Deep Creek mine and with thrust faults and a large number of north-south faults.

Although much of the limb of the fold is overturned, the southwestern part is not. This right-side-up section does not seem to have escaped overturning, but to have been overturned and later righted. This restoration to a right-side-up position is interpreted to be the result of refolding the rocks by buckling the arc folds during the crossfold event. This buckling produced the synform at the Deep Creek mine, the strike faults and reverse faults in the synform, and the north-south "tear faults."

The triangular area of Metaline and Maitlen rocks north of the mouth of Lime Creek in the southeast half of sec. 28, T.39N., R41E. is difficult to fit into any movement pattern. The simplest explanation (fig. 7) is that it is a faulted part of the buckled rocks two miles to the southwest and that it has moved northeastward on the northeast trending fault that bounds the northwest side of the triangle.

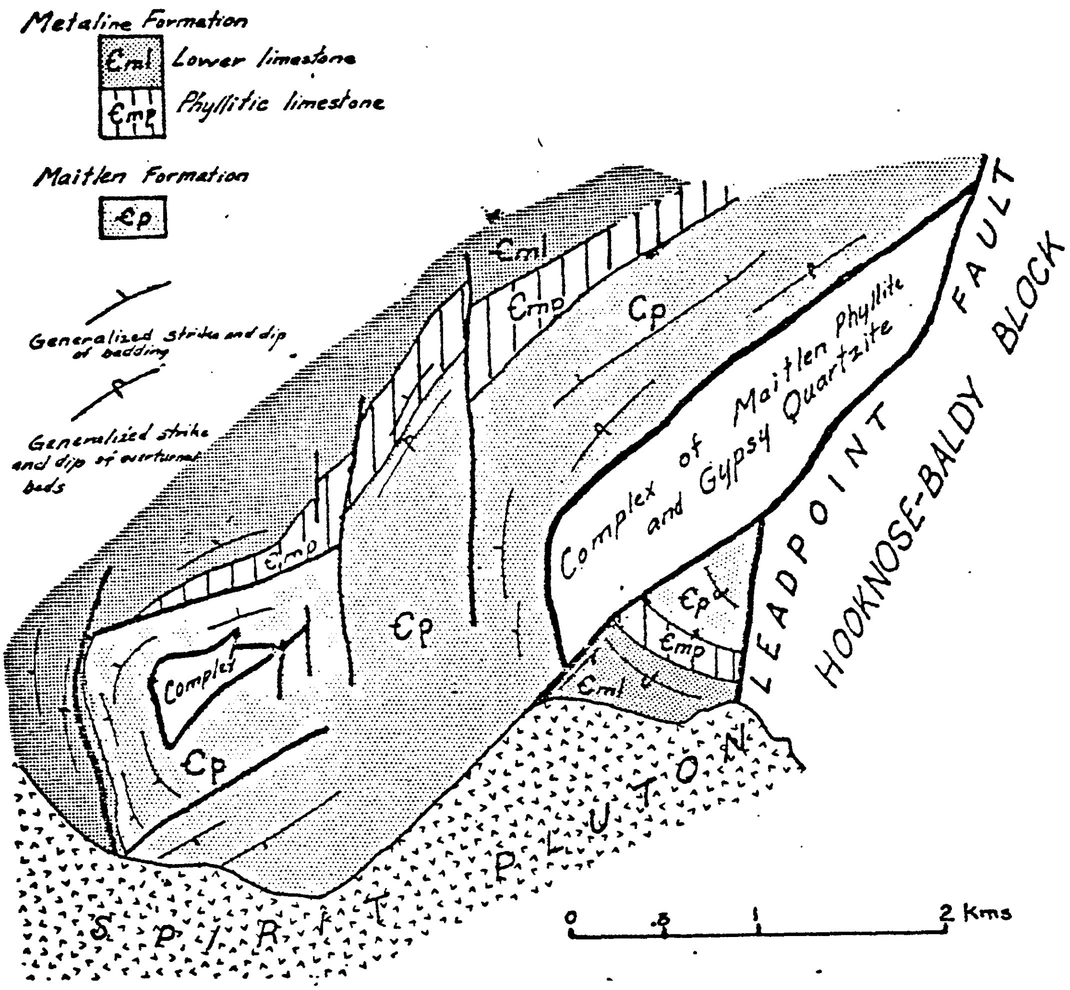

Figure 7.-- Generalized structural relations at southeast end of Lime Creek Mountain block.

The synform at the Deep Creek mine (see fig. 7A) is a secondary fold that was superimposed on the arc folds during the cross-fold event. The tendency to buckle the beds evenly and produce a symetrical and continuous fold, extending from the Deep Creek mine S.60°E. to the southwest quarter of sec. 30, T.39N., R.41E., was resisted and prevented by horizontal movement on faults or slip planes along bedding that were either relics of the earlier folding or faults that developed early in the crossfolding. One of these slip planes is the arcuate fault indicated as present under the alluvium in the Deep Creek valley, another, the Great Western fault, striking north-south lies immediately west of the portal of the Great Western mine. The beds west of the fault that lies along Deep Creek are tightly compressed into the synform, which is a normal flexural slip fold developed by slip evenly distributed along bedding planes. In contrast, the beds east of this fault are not tightly compressed but are bent into an open arcuate fold. The rocks between the Deep Creek and Great Western faults adjusted to the compression of the crossfolding by strike-slip movement along these two faults and upward movement on the south side of the northeast trending fault that separates the middle and upper units of the Metaline Formation. As a result the fault bounded block moved northeastward relative to the rocks on either side. Compression continued after this northeastern movement and the beds and faults were bent to the southeast. The last result of the crossfolding was the northwest trending faults, most of which have left lateral displacements.

Conceivably the crossfolding was contemporaneous with the emplacement of the Spirit pluton. If so, stress would have been transmitted to the edges of the beds north of the pluton by a fluid medium; consequently the beds under compression could adjust by bending with greater latitude than if the interface had been solid against solid. I at first believed that the buckling of beds along the north border of the Spirit pluton was solely the result of forceful emplacement of the pluton. However, when it became known that the buckling or "kink fold" extended for considerable distances both east and west of the pluton, this hypothesis was untenable. It is, nevertheless, possible that the crossfolding and emplacement of the pluton are essentially contemporaneous.

The two areas on figure 2 in the southern part of the Lime Creek Mountain block shown as segments of a thrust plate are believed to be part of the autochton that composes the core of the Columbia anticline (fig. 14) of which the rocks of Red Top Mountain are a part. The thrust is believed to have moved the rocks of the Lime Creek Mountain block westward over those in the Columbia anticline. The arc folding later brought parts of the autochthon within the reach of erosion.

The syncline in the Ordovician rocks near the Advance mine on the south side of Black Rock Canyon (see fig. 2) is difficult to relate to neighboring structures. Interpretation is handicapped by the general sheared character of rocks near the Black Rock Canyon fault. These sheared rocks are possibly a faulted counterpart of the syncline in the area mapped as "argillite with quartzite lenses and limestone beds," symbolized by "Da" on figure 2. Under this interpretation, the fault along the west side of the syncline is an east dipping normal fault along which the syncline has moved eastward.

An alternate interpretation is illustrated in the sketch (fig. 8) that shows the syncline as a drag fold that has been thrust northward. This explanation seems closer in agreement with the rock distribution.

Red Top-Grass Mountain block

The Red Top-Grass Mountain block is part of a large, highly broken anticlinal structure--here referred to as the Columbia anticline--that extends from the Salmo mining district of British Columbia southwestward down the Columbia River to the vicinity of Kettle Falls where it terminates against the upfaulted block of gneissic and granitic rocks that composes the northern part of the Kettle River Range. This fold is everywhere "cored" by Cambrian rocks and everywhere enveloped by post-Cambrian, probable Paleozoic, dark argillites.

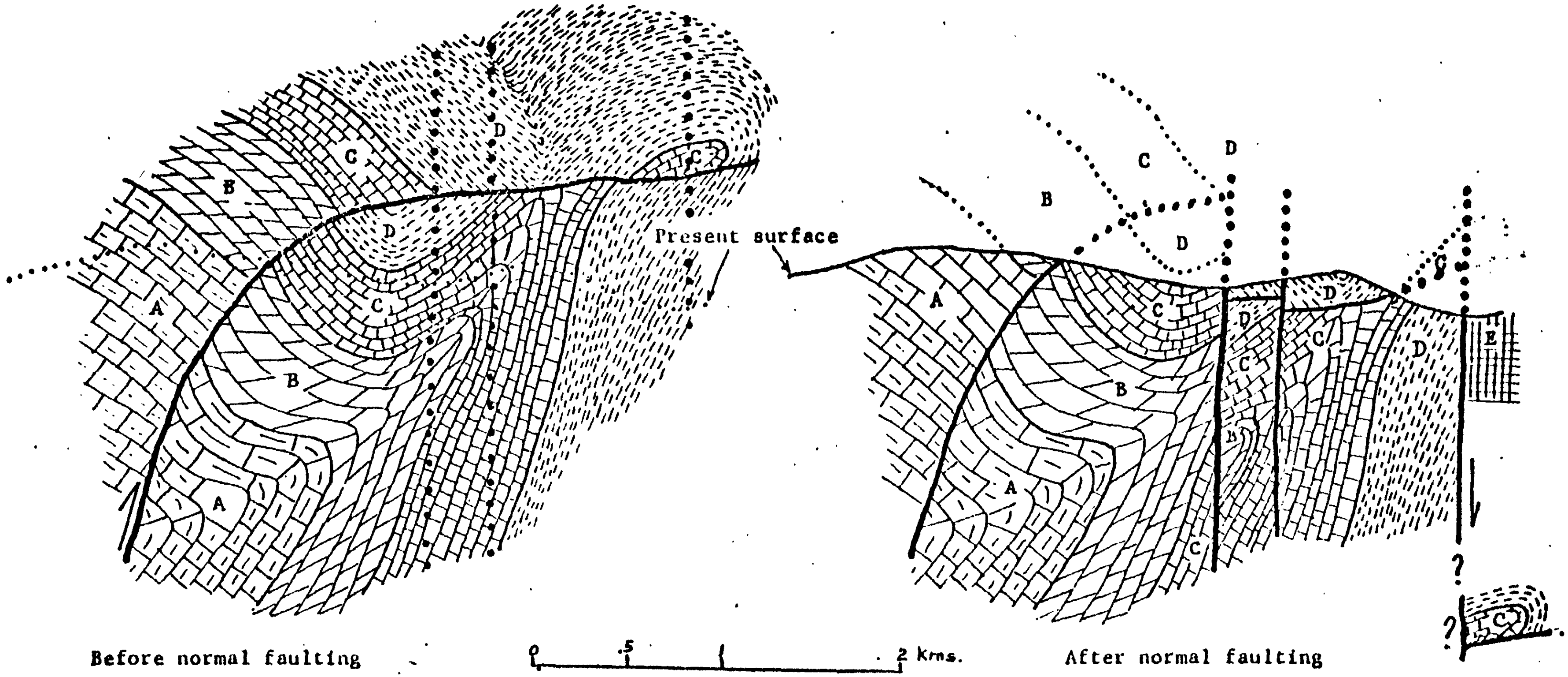

Figure 8.--Sketch cross sections interpreting structure in Lime Creek Mountain block south of Black Canyon. Section extends S. 3 E. from elevation 2720 ft. in stream bed of Black Canyon. Rocks in sections: A, B, C, = lower, middle, and upper units of Metaline Formation, respectively; D = Ledbetter Slate, and E = Grass Mountain sequence. Black Canyon fault not shown.

The Columbia anticline was broken into numerous segments during the crossfolding event. During this deformation, sections of the fold were upthrust or upbuckled bringing to the surface the Cambrian core rocks. One and a part of another such segment is in the Deep Creek area; the Red Top Mountain segment is exposed in the northeastern half of the Red Top-Grass Mountain block, 2) the other segment extends only a short distance into the map area that part where it is called the Northport block. The segment in its entirety is described in the section on Regional structure as the Northport segment.

Only the core rocks can be mapped with enough precision to delimit the basic structure of the Red Top-Grass Mountain block; the envelope rocks are so lithologically homogenous that even gross structures can go unrecognized. The core rocks are divisible into units of quartzite, chlorite schist, and limestone, which are moderately thin and lithologically distinctive. The envelope rocks are subdivided into thick unit of uncertain relative age and questionable unity. Whether or not the same structural pattern developed in both core and envelope is difficult to judge. The core rocks appear to have tighter, more compressed structures and have a stronger tendency towards assymetry than the rocks at higher structural levels. This may be partly the result of differences in depth and partly differences in the physical character of the rocks. The tendency to shear incompetently was strong in the envelope rocks; the tendency of fold, strong in the core rocks.

The Red Top-Grass Mountain block thrusts over the Pend Oreille block to the north and is upfaulted against the Black Canyon fault to the southeast. Its eastern end butts against the Leadpoint fault.

The core rocks exposed in the northeastern part of the block are broken into three parts by north-south vertical faults. The eastern part, on Red Top Mountain, is a complex fold; the middle part, on the eastern slope of Cedar Creek valley, is a pair of anticlines; and the western part on the western slope of the Cedar Creek valley, is an overturned anticline.

The fold on Red Top Mountain is indicated on figure 2 as a syncline on the basis of dips of bedding planes. Such a structure is undeniably present; however, it may be that this syncline is superimposed on a recumbent thrustfaulted anticline, whose axial plane before the refolding dipped gently southeastward. Alternate interpretations are shown in figure 9. Although such two phase folds are not evident in the rocks to the west, several features of the fold strongly favors this interpretation. The axial planes of associated minor folds shown on the northwest limb of the syncline strike from N.60°E. to E-W and dip southward. The axis of the syncline plunges S.60°W. at about 45°, which is the axial orientation of the minor folds. The axial plunge of the arc folds is typically very gentle except where steepened by crossfolding. The steep plunge and open character of the syncline is anomalous in comparison to the folds to the west, which are overturned isoclinal folds with gentle axial plunges. It is easier to relate the crinkled limestone beds in the trough of the syncline to a refold fold than to an open plunging syncline.

Figure 9. Geologic map and cross sections of the Red Top Mountain-Cedar Lake area. Map is a simplified version of the geology shown on plate 1.

On the east side of the Cedar Creek valley are two anticlines separated by a faulted syncline (fig. 9, sec. B-B'). Both are overturned with axial planes dipping southeastward. The northern anticline could be the up-faulted part of the postulated refolded anticline of Red Top Mountain (fig. 9, sec. C-C'). The western part of this fold--separated by a N.10°E. fault under the alluvium of the Cedar Creek valley--occurs on the hill west of Cedar Lake (fig. 9, sec. A-A'). This is the best exposed and the least broken of the three fold segments. It also is an isoclinal anticline. The trace of its axial plane is sinuous.

The fragmentation of the core of the Columbia anticline in the Red Top-Grass Mountain block during the cross fold event was accomplished by the combination of thrust faults striking near east-west and tear faults striking north to N.10°E. The thrusts have dips estimated to be about 30 degrees south. The tear faults both terminate at and cut through lower thrust plates.

The structures observed in the argillite envelope rocks have some of the characteristics of those in the core rocks, but because of the lack of useful stratigraphic reference units, the closeness of this relationship is obscure. In sec. 29, T.40N., R.40E (fig. 2) the boundary between the argillaceous limestone of the Grass Mountain sequence and the argillite of the Pend Oreille sequence show the thrust and tear deformation characteristic of the core rocks. Structures such as the "thrust and tear" structures would be difficult to recognize in the large areas of argillite devoid of marker beds. However, because of the differences in physical characteristics and because the two rock assemblages are separated by thrust faults--or thrust faults are in the argillite a short distance above the contact--it seems reasonable to assume that there would be more fluid shear structures in the envelope rocks and tighter more fractured structures in the core rocks. Doubtless the core rocks had less tendency to fracture in the deeper parts of the anticline.

Pend Oreille block

The Pend Oreille block, the northernmost block, is less rigidly bounded and less understood both stratigraphically and structurally than the other blocks. Its southern boundary is a shear thrust zone, the Columbia fault, that extends eastward across the map area; its northern boundary is the south dipping Waneta fault, that has been traced many miles in British Columbia. The block contains only three stratigraphic units, which are more properly assemblages of related lithologies rather than homogeneous stratigraphic units. If differs principally from the other blocks by a predominance of east-west instead of northeast trends. Albite granite, as a sill-like intrusion, is found only in this block.

If the stratigraphic units and interpreted relations are valid, a N.75°W. trending anticline overturned to the north is defined. The axial plane of this fold appears to closely parallel both the Columbia fault and the Waneta fault, which is adjacent to the International Boundary in British Columbia. All three structures may have been formed during the cross fold event.

An anomalous structure, apparently chronologically inconsistent with this interpretation of an overturned anticline, is the southwest plunging syncline that lies above the Sheppard Granite on the ridge between Cedar Lake and the Columbia River. Although the northeast trend of this fold places it in the arc fold system, it is obviously superimposed upon the N.75°W. fold supposedly related to the cross fold system. A close inspection of the geologic map (fig. 2), however, indicates a genetic relation between the syncline and the Sheppard Granite: the syncline is not present in the rocks below the Sheppard Granite; therefore the syncline is the result of variations in the thickness of the granite sill, and is independent of any regional fold system.

The Sheppard Granite, a medium grained leucogranite composed of albite and microcline plus very small amounts of pyrobole altered beyond positive identification, is a pinching and swelling sill-like intrustion. It has two unusual characteristics: (1) although unquestionably an igneous rock, it had remarkably little thermal effect on the rocks it intruded, and (2) almost all specimens exhibit cataclastic structures such as albite lamellae and much of the outcrop surface shows randomly oriented streaks of mylonite.

Outcrops west of the Columbia River on the slopes of Mitchell Mountain are inadequate to give a structural interpretation that can be incorporated with that given rocks on the east side of the river. Except for the northwesterly striking rocks that border the Sheppard Granite on Moraski Mountain, the general strike of bedding, cleavage, and foliation is northeasterly and dips, southeasterly.

Fold axes generally plunge to the south, in contrast with the common southwest plunge of folds in the remainder of the quadrangle. The rocks in the Pend Oreille block are more intensely sheared and deformed than elsewhere in the Deep Creek area.

Area south of the Spirit pluton

The area south of the Spirit pluton contains the northern end of several large tectonic blocks that are under investigation in the south half of the Colville 30 minute quadrangle by J. Eric Schuster of the Washington State Division of Mines and Geology. Although relations between the units cannot be completely resolved in the Deep Creek area, enough is known to make it possible to place them within the framework of the structural history that has been established for the area north of the Spirit pluton.

The area is divided into two principal parts by the South Fork fault, a N.10°E. trending high angle fault. East of the South Fork fault are two tectonic units separated by the Magma fault, the Magma tear fault, and the Meadow Creek fault. West and north of these faults are Cambrian and Ordovician rocks, south and east are Cambrian and Precambrian rocks. The rocks to the south and east form a homocline that strikes northeast and dips 60° and 70° to the northwest. The homocline is cut by the Byers Creek fault, trending N.75°W. and displacing rocks on the south side of the fault 3,000 metres to the west. The Paleozoic rocks to the north that contact the Spirit pluton have strikes ranging from northeast to N.60°W. with most, nearly east-west. Dips are both to the north and to the south. The Magma fault is interpreted as a pre-pluton, south dipping reverse or thrust fault. Its eastern extension would fall within the Spirit pluton. The Magma tear fault is believed to be a tear fault similar to those of Red Top Mountain.

The South Fork fault is nowhere exposed: It is a structure essential to the separation of stratigraphically and structurally unlike rocks on opposite sides of the valley of the south fork of Deep Creek. It possibly is the southern extension of the Leadpoint fault. The lack of offset of the contact of the Spirit pluton indicates that it is preintrusive.

The rocks that are west of the South Fork fault are separated by faults into three tectonic units. The quartzite, limestone and schist units correlated with the Maitlen and Gypsy Formations that are grouped around the Blue Ridge (Sierra zinc) mine are believed equivalent to the lower plate rocks exposed in the Lime Creek Mountain block. The patch of white Gypsy Quartzite on hill 3159 is believed to belong with the rocks south of the Magma West fault, being a faulted outlier of these rocks.

The Cambrian rocks south of the Magma West fault form an anticlinal fold that extends 24 kilometres southwest to the vicinity of Colville. This anticline has the same stratigraphic assemblage as the Hooknose anticline is considered an extension. The Magma West fault, an east-west southward dipping thrust fault, is considered to be the complement of the Magma East fault, although the two may have been developed independently on both sides of the South Fork fault. The small area of rocks of the Metaline Formation in the southwest corner of the map area (fig. 2) is correlated with upper plate rocks of the Lime Creek Mountain block.

Faults that bound the blocks

The faults that bound the blocks are the least exposed of all the large structures in the area. The finely comminuted rocks of fault zones do not crop out in areas such as this where the climate favors ready development of soil and plant cover; consequently, the character of the faults is largely determined indirectly and inferentially. The Leadpoint and Black Canyon faults are high angle extensional faults; the Columbia fault is a thrust fault, or more properly a thrust zone; the South Fork fault is as described in the preceeding section, a possible extension of the Leadpoint fault. The Leadpoint fault was seen in only one place, being largely covered by valley fill; the Black Canyon fault plane, or planes, was never seen, but its position could be located on the map with reasonable accuracy; the Columbia fault is basically a shear zone between two rock sequences that share the same lithologies, consequently the fault is recognized by a shear instead of lithologic contrast.

Leadpoint fault

The Leadpoint fault extends in a N.30°E. direction from the north edge of the Spirit pluton to the northeast corner of the map area (fig. 2). A fault that may be the southward continuation of the Leadpoint fault is the South Fork fault, which extends S.10°W. from the south edge of the pluton. The extension or connection of these two faults through the pluton would be hidden under the alluviated valley of the South Fork of Deep Creek. As neither the north or south contacts of the pluton appear to be offset where the faults would cross, it is highly unlikely--assuming the projection to be correct--that there was movement on the fault after the emplacement of the pluton.

The trace of the Leadpoint fault north of the pluton is also largely under Quaternary unconsolidated deposits. Only east of Joe Creek, a tributary of the East Fork of Cedar Creek, can the shattered rocks along the fault be clearly seen. Here highly brecciated and veined Gypsy quartzite and limestone of the Pend Oreille sequence are dragged northward along the west side of the fault thus indicating either a relative upward or northward movement on the west side of the fault. The quartzite and limestone are separated in turn by a thrust fault, the Columbia thrust fault, which is involved in this drag.

The fact that this thrust fault is involved in the drag, unfortunately does little to help establish the relative ages of the two faults or direction of movement on the Leadpoint fault. Five interpretations (illustrated on fig. 11) must be considered: (1) the Columbia thrust fault is the older fault and was bent by the block east of the Leadpoint fault moving downward--this hypothesis assumes an eastward dip on the Leadpoint fault, (2) the thrust fault is the older and was bent by a northward horizontal movement of the block east of the Leadpoint fault, (3) the thrust fault and Leadpoint fault are contemporaneous, genetically related products of the cross fold event, and movement on the Leadpoint fault is horizontal and rocks on the east side of the fault moved at a higher velocity than rocks on the west side, (4) the Leadpoint fault is older than, and independent of, the cross fold event and during crossfolding rejuvinated movement on the Leadpoint fault was that of interpretation 3, and (5) movement on the Leadpoint fault was in two stages: first development of a thrust-tear combine, and second, normal faulting on tear (Leadpoint) fault and extension of the "tear" northward. The conclusion of the writer is that drag can be an unreliable criterion to indicate direction and time of fault movement. Consideration of the regional structural relations favors interpretations 1 and 5, with 5 having the edge. The Leadpoint fault is clearly younger than the arc folding, older than the Spirit pluton (100 million years) and probably as old as the cross fold event. The vertical component of movement can be interpreted to be as much as 6,000 metres (see section, fig. 25), with the east side down.

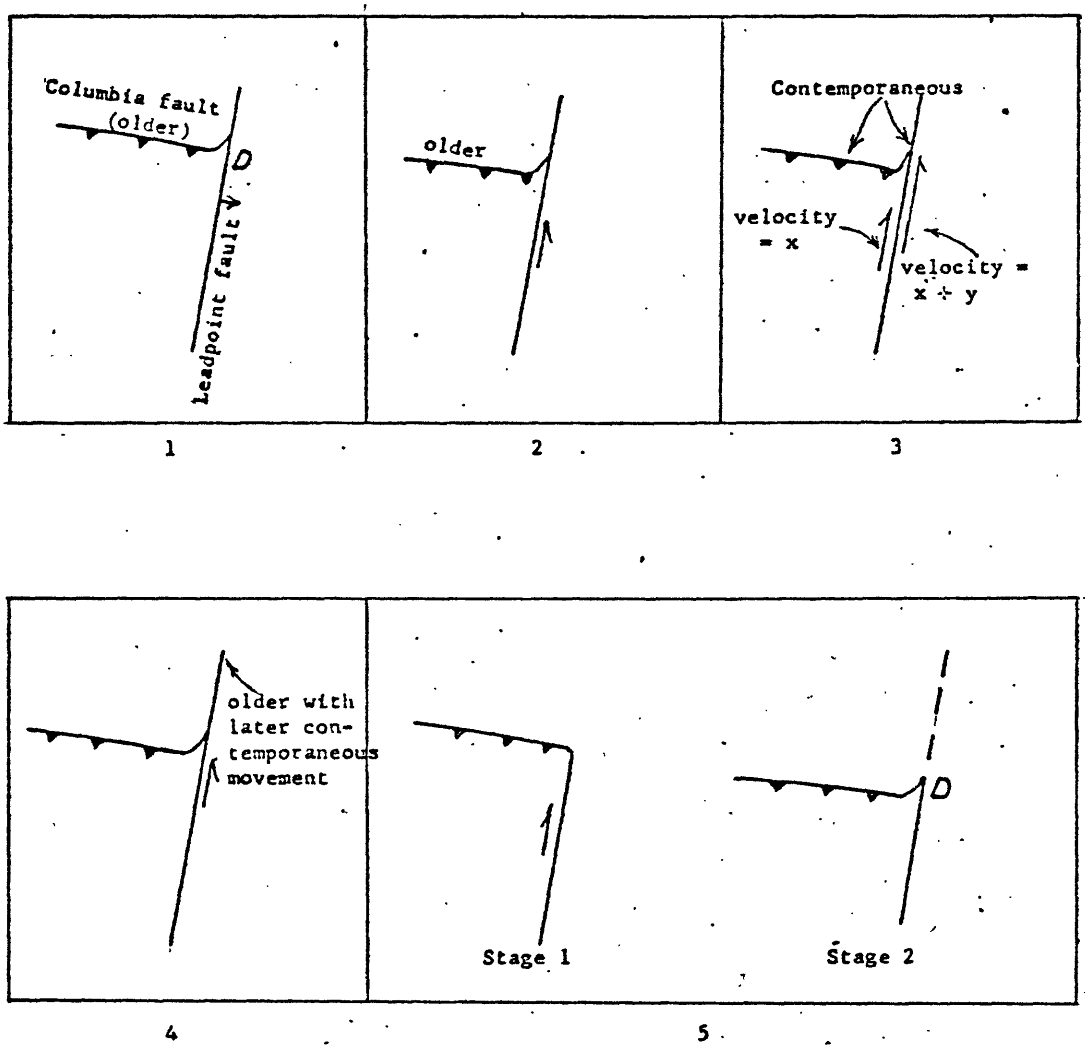

Figure 11.-- Diagrammatic sketche maps indicating hypothetical movement patterns of the Leadpoint fault.

Because there is so much freedom in positioning the Leadpoint fault under the Quaternary deposits in the Deep Creek valley, the character of fault intersections has to be inferred from relations in the better exposed areas. Although the preliminary edition Yates (1964) of figure 2 projects many faults under the alluvial cover and shows inferred intersections between faults, it was decided, because of the lack of data the interpreted intersections should be omitted from figure 2. Some indications of the nature of the intersections of the Leadpoint fault with northeasterly (N.40-75°E.) faults is shown in two places on figure 2. The N.40°E. jog in the indicated trace of the Leadpoint fault northwest of Deep Lake appears as a right lateral offset in the Leadpoint fault along a N.50E. fault that cuts both Hooknose-Baldy and Lime Creek Mountain blocks. A similar right lateral apparent jog in the overall N.30°E. general trend of the fault occurs on a N.75°E. fault on Hartbauer Creek four miles to the north. This occurs in an area of poor exposures where three blocks, Hooknose-Baldy, Lime Creek Mountain, and Red Top Mountain, converge, consequently the nature of this apparent offset is unknown. It is not a simple offset along the N.75°E. fault, because this fault does not extend beyond the strands of the Leadpoint fault.

Black Canyon fault

The Black Canyon fault, which separates the Lime Creek Mountain block from the Red Top-Grass Mountain block, extends N.60^{o}E. from the midpoint of the west side of the quadrangle to about one kilometre south of Cedar Lake, where it appears to terminate against a N.10^{o}E. trending fault (see fig. 2). What is interpreted as a northeastern offset extension of the fault bounds the upper unit of the Metaline Formation on the southeast slope of Red Top Mountain. If this interpretation is correct, the offset occurred during the cross fold event and accordingly the Black Canyon fault is contemporaneous with, or older than the cross fold event. The Black Canyon fault does not extend east of the Leadpoint fault, unless the Russian Creek fault is its extension (fig. 12).

Between Black Canyon and the Cedar Lake valley, the Black Canyon fault was traced on the basis of structural discordance (terminated units), sheared and otherwise deformed rocks, and missing stratigraphic units; southwest of Deep Creek it was projected through a heavily forested area along the course of a strong lineament visible on aerial photographs. By relating the trace of the fault to topography it is a high angle fault; if interpreted as a postfold fault, the rocks on the southeast side are inferred to be displaced downward for an estimated kilometre or more. The strike of the beds on the southeast side of the fault parallels the trace; that of beds on the northwest side both parallels and is discordant to the trace.

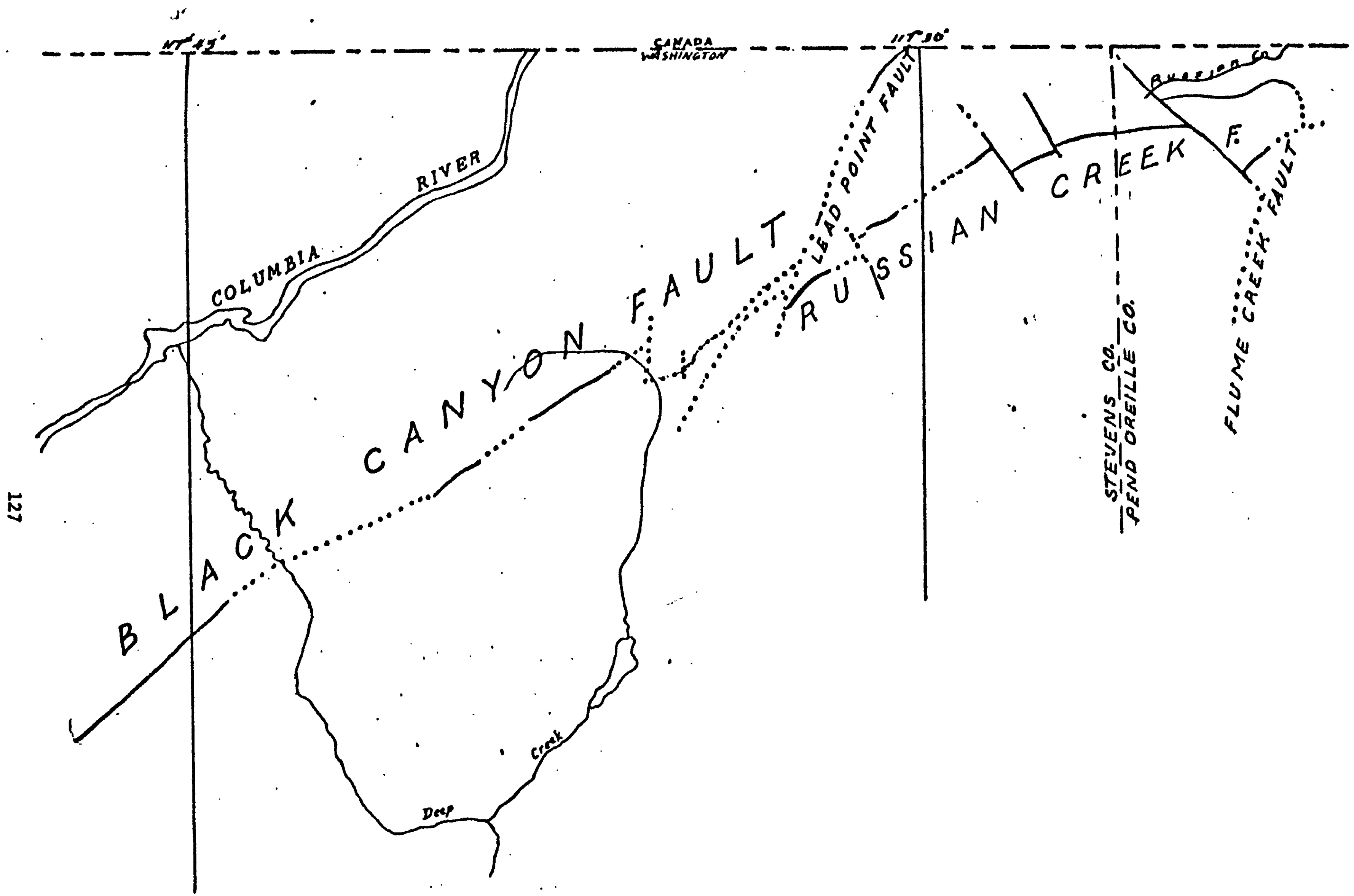

Figure 12.-- Trace of Black Canyon and Russian Creek faults. Only other faults shown are those that disrupt the trace.

To understand the character and function of the Black Canyon fault as a block-bounding structure, it is necessary to understand the contrasts and affinities between the Red Top-Grass Mountain block and the Lime Creek Mountain block. The two blocks, although stratigraphically different, have common features. Most important of these is that the core assemblage of quartzite, limestone, and phyllite of the Red Top-Grass Mountain block also occurs in the Lime Creek Mountain block where it is separated by a thrust fault from the section of Cambrian and Ordovician rocks that closely resembles that in the Metaline quadrangle. The argillite sequences of the two blocks apparently have no common units, although both sequences have depositional contacts with the uppermost beds of the Metaline Formation. The contrast between the Cambrian sequence of Red Top Mountain and the sequence above the thrust fault in the Lime Creek Mountain block is sufficiently great to justify the conclusion that the two sequences must have been deposited farther apart than they are now. The two sections could have been most easily juxtaposed by a thrust fault that moved before or early in the arc folding and before the Black Canyon fault (see fig. 12a). The thrust fault in the Lime Creek Mountain block is permissive to these requirements, but no counterpart of this fault has been identified in the Red Top-Grass Mountain block. In this block the thrust would have to lie above the Metaline Formation, either in the Grass Mountain sequence or in rocks eroded from above that sequence.

If the postulated thrust fault tectonically divides the Grass Mountain sequence into two groups, these rocks cannot be the sequence of closely related rocks that I have described; the group in the lower plate would be in depositional conformity with the Metaline Formation and the group in the upper plate would be rocks that lie above the upper, and missing, argillite facies rocks of the Lime Creek Mountain block. Nevertheless, as emphasized in the section on stratigraphy, the relations between rock units of the Grass Mountain sequence are so tenuous, that one must entertain the suggestion that they may be divided by a thrust fault into two tectonic groups. If one must speculate, it is least speculative to suggest that all units above the Red Top argillite belong to the upper plate, but it is almost as conceivable that unit Cg_5 may belong with the Red Top argillite in the lower plate. As the postulated thrust is interpreted to be prefold structure it would be difficult to identify it in the confusion produced by the arc folding, cross folding, and high angle faulting.

On the other hand, if the lost thrust fault was in rocks now eroded from above the Grass Mountain sequence, the units of the sequence have been correctly interpreted as closely related. Whether or not the thrust fault lies within or above the Grass Mountain sequence does not change the movement direction on the Black Canyon fault.

The Lime Creek Mountain block is the overturned northwest limb of an anticline and the southwest limb of a syncline; the Red Top-Grass Mountain block is interpreted to be an overturned anticline. The area between the two structures, which should be synclinal, is cut by the Black Canyon fault. Although the exposed rocks do not define a syncline, such a structure can be restored if we assume downward movement on the southeast side of the Black Canyon fault as sketched in figure 12a.

This, the favored interpretation, however, does not tell us whether the thrust was in rocks now eroded from the Red Top-Grass Mountain block, or whether it exists within the preserved Grass Mountain sequence as an unidentified fault fragmented by the cross fold event. The interpretation is, nevertheless, compatible with the hypothesis that the Black Canyon fault is a segment of a larger fault, possibly the Russian Creek fault.

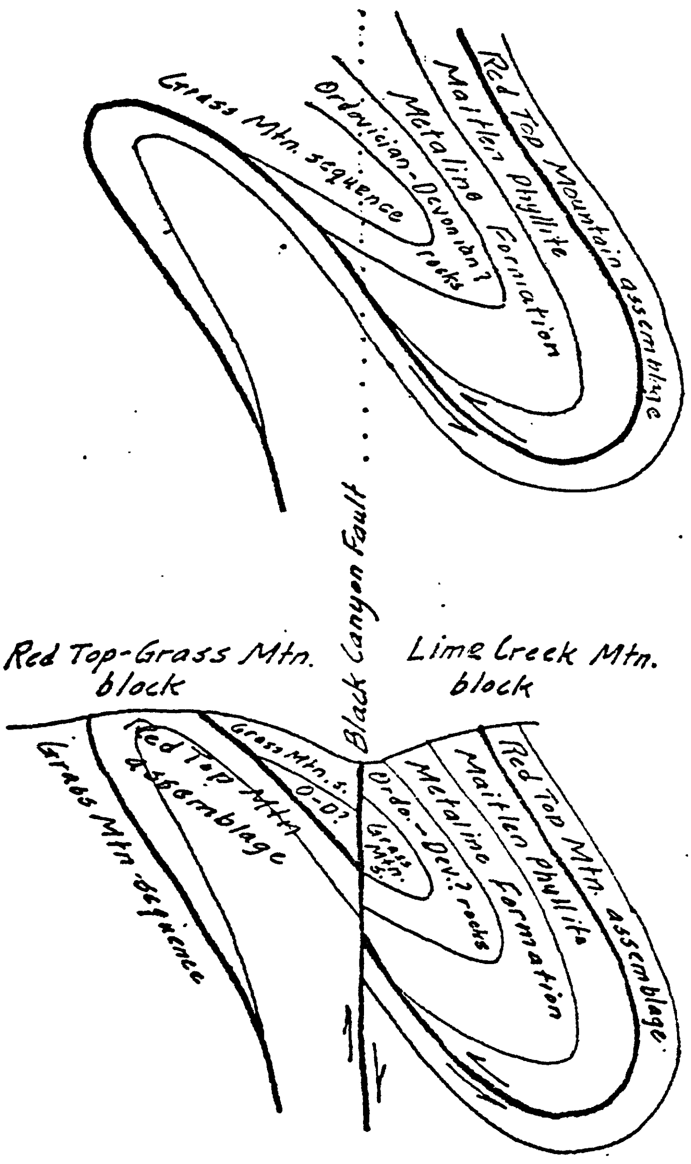

Figure 12a.-- Sketch showing interpreted relations between Lime Creek Mountain and Red Top-Grass Mountain blocks. A. Before Black Canyon fault. B. After Black Canyon fault and with erosion to present level.

Columbia fault

The Columbia fault is a broad zone of shearing that is interpreted as a thrust fault that brings the overturned anticline of the Red Top-Grass Mountain block northward to rest on the south flank of the overturned anticline of the Pend Oreille block. The fault trends about N.80°E., terminates to the east against the Leadpoint fault, and extends westward under the terrace deposits and into the Northport quadrangle. The dip to the shear zone probably average about 30 degrees to the south.

The nature of there termination of the Columbia fault against the Leadpoint fault is discussed and interpreted on page 123. The favored interpretation is that the Leadpoint fault is contemporaneous with and related to the Columbia fault, but had a later independent normal movement.

Because of its strike, dip, and association with a belt of smaller but similar thrusts that clearly are superimposed on the arc folds in the Northport quadrangle to the west (Yates, 1971), the Columbia fault is considered to be a product of the Cross Fold event. The shear zone representing the thrust fault is only to be seen on the north slope of Red Top Mountain and on the divide between Cedar Creek and the Columbia River; elsewhere the fault is covered by glacial fluvial and terrace deposits. Rocks involved in the thrusting are mainly black argillites and fine-grained quartzites, which are strongly cataclastic having finely ground quartz laminated with thin seams of recrystallized muscovite (fig. 13). The shearing and cataclastic structure extends for more than 100 metres beyond what is considered the principal plane of movement.

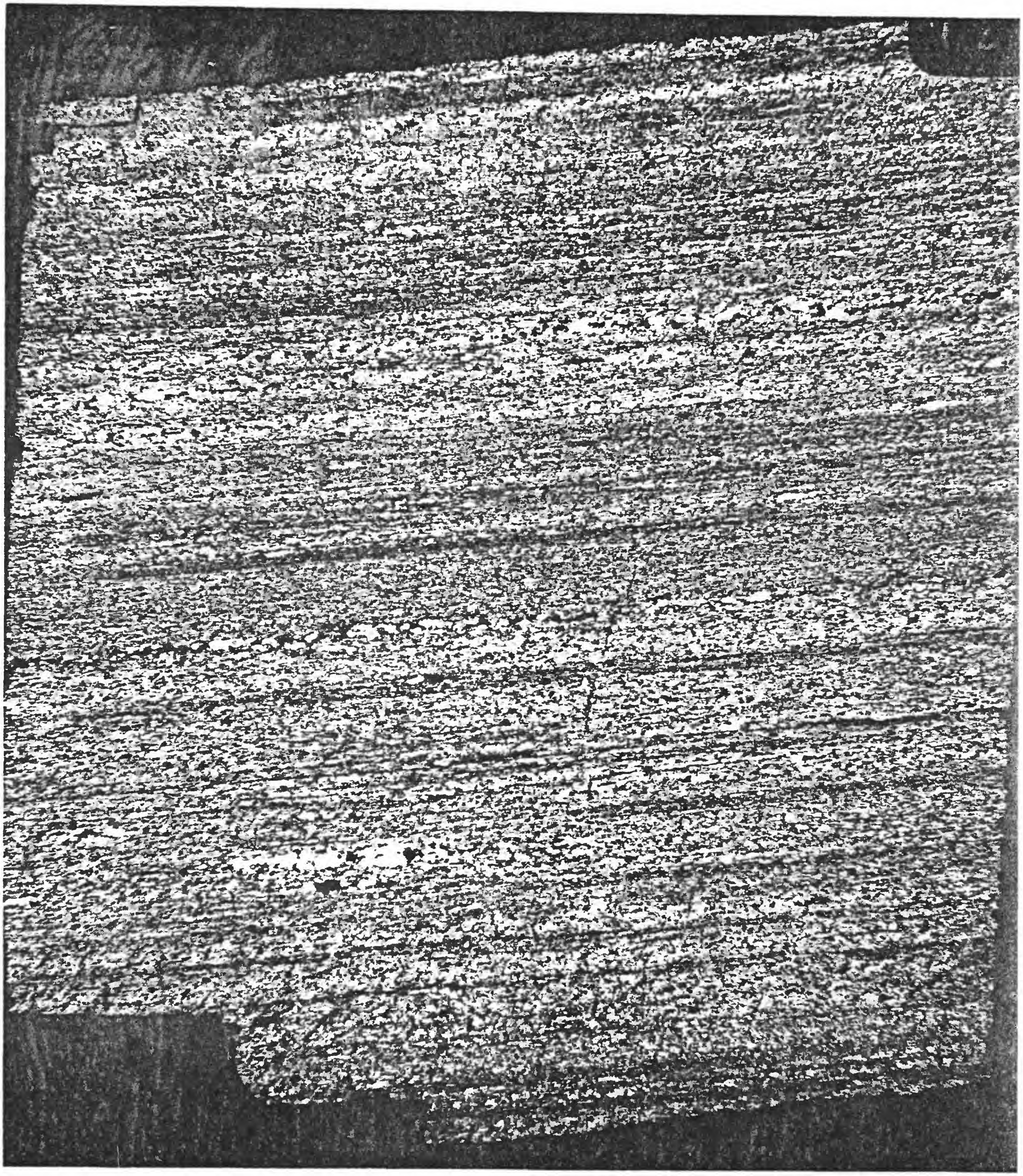

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Figure 13A and B. -- Photomicrographs of cataclastic black argillite. Quartz-muscovite pelite from north slope of Red Top Mountain.

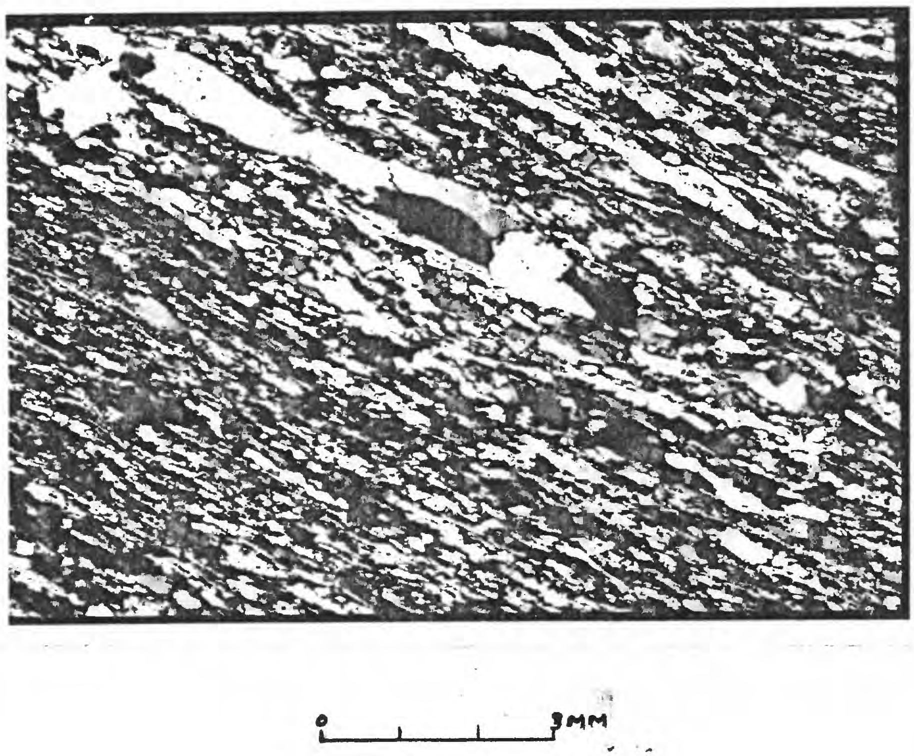

Figure 13B

On the preliminary map of the Leadpoint quadrangle (MF 137) (Yates and Robertson, 1958), a zone of phyllite with an outcrop width of 600 metres immediately above the thrust, was separated out as a separate unit, although it was recognized that the phyllite "may be a highly deformed phase of bsp," (unit bsp is what is called Cg_5 of the Red Top-Grass Mountain sequence on fig. 2).

The Columbia fault is not a typical example of the thrust faults of the Cross Fold event; indeed it is difficult to determine the precise character of the fault, because of poor exposures and because it separates noncontrasting lithologies. Most faults of the Cross Fold event are characterized by numerous "tears" that offset the fault and may or may not pass into the lower plate. These appear to be absent in the stretch of the fault crossing the Deep Creek area, except where the fault crosses the valley of Cedar Creek, but such faults could be hidden beneath the terrace deposits along the Columbia River. The trace of the fault may not be as uninterrupted as indicated.

Regional setting

The rocks

Introduction

From the preceding descriptions of rocks and structures, it is possible at this point to make interpretations of the geologic history of the Deep Creek area, but at best the interpretations would be little more than generalizations giving but a sketchy account of the history. The account can be greatly enhanced through an examination of the geology of the surrounding region, because in the larger area it is possible to relate the stratigraphy to regional cycles of deposition and erosion and the structure, to tectonic events that molded the continental framework. An areal (fig. 14) that includes the Deep Creek area (fig. 2) the Metaline quadrangle to the east (Park and Cannon, 1943, pl. 1), the Northport quadrangle to the west (Yates, 1971), and a strip of British Columbia, 15 minutes of latitude wide, (Little , 1965, 1962, and 1963) immediately north of the International Boundary contains rocks representative of all periods of geologic time from the Precambrian to the Cenozoic. By integrating the published knowledge of the many from whose work this map is compiled and my unpublished observations, it is possible not only to fill gaps in the history of the Deep Creek area, but to use the geology of the Deep Creek area as a means to arrive at a reasonably coherent and consistent account of the geologic history of the region.

p. 137 follows

The system used in the preceding section to describe strato-structural units in the Deep Creek area breaks down when this area is combined with those surrounding it. Although each strato-structural block is a legitimate entity useful in describing details of structure and stratigraphy, the system is too detailed to apply to the regional stratigraphy and structure. Instead the rocks will be described in the traditional order from older or younger, which fortunately has some system in the region in a geographic sense, as well as a chronologic sense. In the area of figure 14 the oldest rocks, Precambrian, are the farthest east; the lower Paleozoic rocks are intermediate in position; and the late Paleozoic, Mesozoic and Tertiary rocks are farther west. Also--in a very general sense--this age grouping corresponds to tectonic provinces, which are, from east to west, the Belt-Purcell anticlinorium, the Kootenay arc, and the Pacific Borderland (fig. 15). Only the area in the eastern margin of figure 14 is occupied by rocks of the Belt-Purcell anticlinorium and only the western margin by rocks of the Pacific Borderland; the central, major part gives a sample of the geology across the Kootenay arc. Although the geology of the Deep Creek area is primarily concerned with the rocks of the Kootenay arc, the relations between the arc and neighboring provinces are discussed because they are vital not only to an understanding of the arc rocks but also to an understanding of this part of the North American continent.

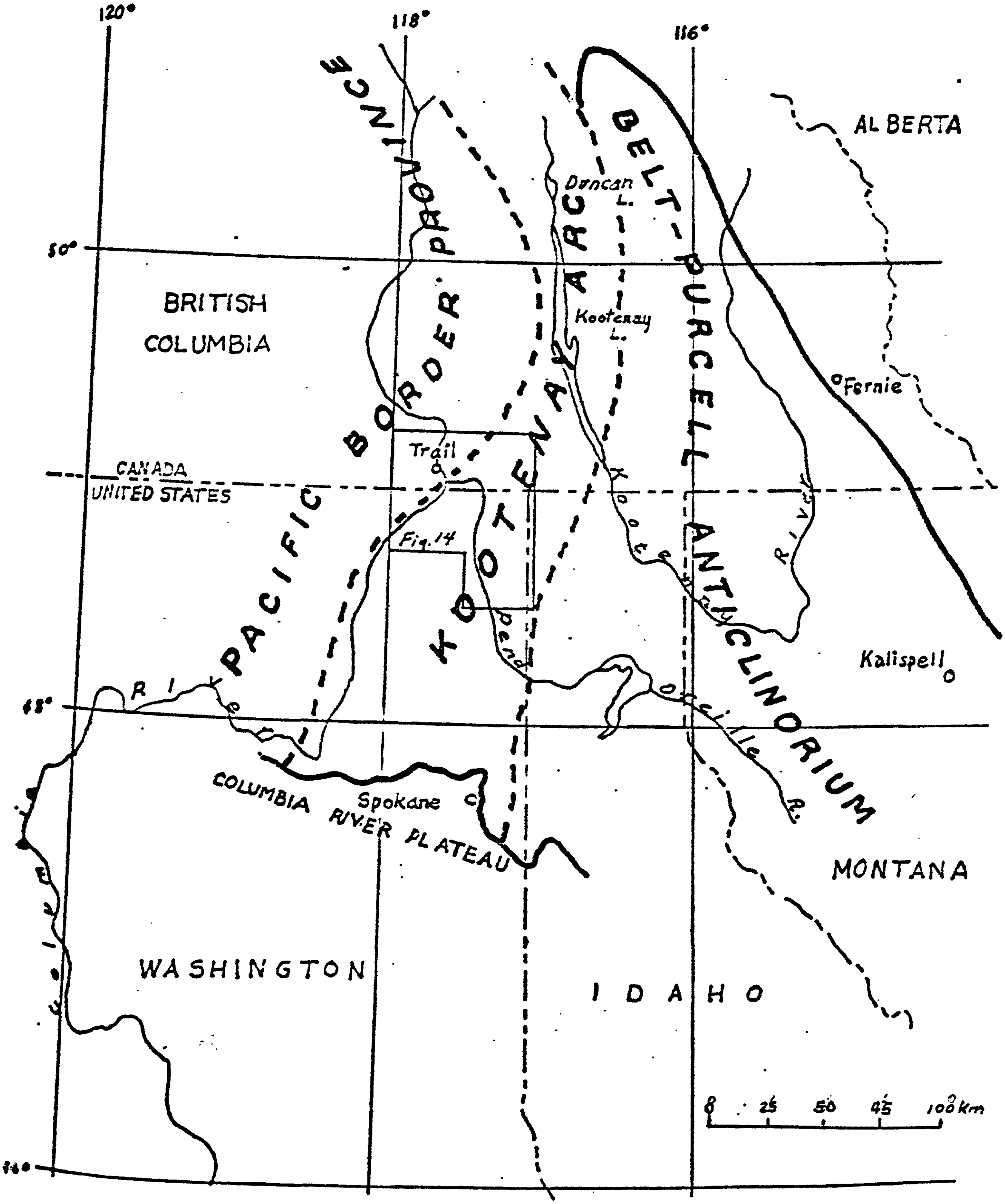

Figure 15.-- Index map showing location of Belt-Purcell anticlinorium, Kootenay Arc, and Pacific Border Province. Area of Figure 14 outlined.

The three provinces, Belt-Purcell anticlinorium, Kootenay arc, and Pacific Borderland are geologic provinces whose tectonic behavior created environments that favored the accumulation of distinctive rocks assemblages. The Belt-Purcell rocks of the anticlinorium, dominantly fine- to medium-grained clastics, accumulated on the craton that consisted of a crystalline basement complex whose internal tectonic activity was limited to downwarping and block faulting during the 600 million years it took the Belt-Purcell sediments to accumulate. The Kootenay arc, basically a foldbelt, represents a zone of change from this stable, continental basement environments to the unstable environment at the oceanic border. It is the hinge zone where the sediments of the craton thicken to a miogeosynclinal prism. The Pacific Borderland consists of the Cordilleran eugeosyncline and its volcanic-bearing deposits. It represents the crustal unrest typical of this environment. Its thick covering of sedimentary and volcanic rocks are believed to rest upon an oceanic basement.

Although these are three basically different environments, they are not sharply divided by space, time, or lithology, but are transitional. The divisions between provinces therefore are arbitrary: as for example, the Kootenay arc in an earlier paper (Yates and others, 1966, p. 52) was separated from what is equivalent to the Belt-Purcell anticlinorium by the Purcell Trench, with the following justification, "the northern part of the trench has been mapped as a fault; the southern part where it separates paragneiss terrane from mildly metamorphosed Belt rocks, is assurely a fault, but where it crosses granitic rocks its structural existence is in doubt. Whatever its true structural nature, we can wholeheartedly accept the Purcell Trench as a hypothetical structure that very usefully separates the predominantly Precambrian terrane to the east from the structurally distinctive Precambrian-Paleozoic terrane of the Kootenay arc to the west." The above choice of a structural boundary as the division between two geologic provinces that were defined principally on a structural basis unfortunately does not precisely fit the requirements of a province boundary when the provinces are defined principally upon their characteristics as depositional environments with inherited tectonic properties.

One outstanding objection is that by placing the boundary at the Purcell Trench, one excludes from the anticlinorium large areas of Belt Supergroup rocks, as well as the closely related Deer Trail Group and Priest Group rocks. Conversely, if these Precambrian rocks are included in the anticlinorium--which is a structural as well as stratigraphic feature--the structural integrity of the arc is infringed upon so that in the Chewlah quadrangle to the south of the Metaline quadrangle the two provinces, arc and anticlinorium, become meaningless distinctions. It becomes obvious that the Belt-Purcell anticlinorium must share some of its rocks with the Kootenay arc. Accordingly, the Purcell Trench is accepted here as the tectonic boundary between two geologic provinces and the Belt and Priest River rocks that fall within the area of the map of the regional geology (fig. 14) are considered as belonging to the Kootenay arc province.

The western border of the arc is only approximately defined. In theory, it should separate the miogeosynclinal facies and the northeasterly folds of the arc from dissimilar trends and the eugeosynclinal facies of the Pacific Borderland. In practice, such a neat separation is not possible, because the fold trends of the arc slop over into the Jurassic volcanic units of the eugeosyncline. The temporal transition from miogeosyncline facies to eugeosyncline facies complicates the position of any boundary as does the lack of well-defined structure juxtaposing the two. Although fold trends in the eastern fringe of the Pacific Borderland--the only part of this province that falls within the map area (fig. 14)--parallel those of the arc, they are atypical of Borderland trends, which are predominantly northwesterly.

Because there are no rigid boundaries between the three closely related provinces, descriptions and discussion also overlap. The discussion of the regional stratigraphy begins with the rocks of the Precambrian and the Belt-Purcell anticlinorium of northern Idaho, proceeds to rocks of the lower and middle Paleozoic of the miogeosyncline of the Kootenay arc province, and ends with the eugeosynclinal assemblage of Late Paleozoic and Mesozoic age of the Pacific Borderland province. The fragmentary record of continental sedimentary and volcanic rocks of the Late Cretaceous and Cenozoic is discussed separately. The nonvolcanic igneous history is mentioned but not discussed. Because many stratigraphic units have two or more names, a correlation table (fig. 15A) is included.

WASHINGTON				BRITISH COLUMBIA		
Deep Creek area and U.S. part of Fig. 14		Chewelah-Loon Lake area	Hunters and Turtle Lake quadrangles	Canadian part of Figure 14		East Nelson map area
Flagstaff Mountain sequence				?		?
Grass Mountain sequence						
Ledbetter Slate				Active Formation		
Metaline Formation		Metaline Formation	Old Dominion Limestone of Weaver	Nelway Formation		
Maitlen Phyllite	phyllite member (informal)			Laib Formation	Upper Laib	Lardeau Series
					Emerald Member	
	Reeves Limestone Member				Reeves Member	Badshot Formation
Gypsy Quartzite		Addy Quartzite	Addy Quartzite		Truman Member	Hamill Series
				Reno Formation		
				Quartzite Range Formation		
				Windermere Group	Three Sisters Formation	?
Windermere Group	Monk Formation	Monk Formation			Monk Formation	Horsethief Creek Formation
	Leola Volcanics	Huckleberry Formation	Huckleberry Formation		Irene Volcanics	Irene Volcanics
	Shedroot Conglomerate				Toby Formation	Toby Formation
Priest River Group		Deer Trail Group	Deer Trail Group			Purcell System

Figure 15A.--Correlative stratigraphic names commonly referred to in this report.

Precambrian rocks of the Belt-Purcell anticlinorium and Kootenay arc

Priest River Group

The oldest rocks under discussion are in the extreme eastern part of the map area (fig. 14). They are rocks, Proterozoic in age, that were named Priest River Group by Park and Cannon (1943, p. 6). The Proterozoic age was determined from their noncrystalline character and unconformable position stratigraphically far below fossiliferous Lower Cambrian rocks. Various workers (Becraft and Weis, 1963, p. 15-18; Hobbs and others, 1965, p. 11; Yates and others, 1966, p. 24) have stated from descriptions of Park and Cannon that the Priest River rocks probably belong to the Belt Supergroup, the thick assemblage of Precambrian clastics and lesser carbonate rocks that underlie a large part of western Montana, northern Idaho, southeastern British Columbia, and adjoining Alberta. But this correlation was based more on stratigraphic position than on lithologic similarity, because the Priest River rocks are described by Park and Cannon in only general terms.

The correlation was placed on much firmer ground by the work of F. K. Miller and L. D. Clark (1975) in the Chewelah-Loon Lake area to the south. Miller subdivided the Deer Trail Group of Campbell and Loofbourow (1957) into two sequences that were physically separated by the covered area of the Chewelah Valley and the Jump-off-Joe fault. The rocks in the hills east of the valley were correlated with units in the Belt Supergroup and named accordingly; units in the lower part of the sequence west of the valley closely resembled Belt units across the valley, but those higher in the section did not. Miller and Clark retained the name Deer Trail for the rocks west of the Valley, concluding that the lower three-fourths of the Deer Trail sequence were facies variants but time equivalents of Belt rocks, and that the upper units were rocks younger than any Belt Supergroup rocks known in Washington, although they were clearly of the same stratigraphic group. The transition from Belt rocks to Deer Trail rocks probably was telescoped by tectonic transport.

Miller and Allan B. Griggs, two workers experienced in Belt stratigraphy, examined some Priest River rocks in the Metaline quadrangle with the writer and, although they were unable to make any positive correlations at that time, it was generally agreed that the Priest River rocks they saw here were more closely related to the Deer Trail Group than to the Belt Supergroup.

The Priest River Group was mapped by Park and Cannon to within 10 kilometres of the International Boundary, where it extends off the east boundary of the Metaline quadrangle into Idaho. The distribution of these rocks on and along the International Boundary was described by R. A. Daly (1912, p. 258-271) under the name Priest River terrane, from which Park and Cannon derived "Priest River Group." Daly (1912, p. 259) regarded the Priest River terrane as Precambrian and pre-Beltian in age. He subdivided the terrane into several "belts", A through G, but in so doing expressed considerable concern as to whether they represented a continuous stratigraphic sequence. Belt A, which occupied the position of what was considered the youngest unit, consisted of a heterogenous group of rocks, largely mica schists, sericitic quartzites and dolomites, listed in the order of abundance. Belt B was more of the same, except that the dolomite was in thicker and more persistent units. The remaining 5 belts are described as varieties of mica schists.

The Priest River terrane that was covered by Daly's boundary survey was remapped along with the East Nelson Map area by Rice (1941). Rice did not regard Daly's Priest River rocks as pre-Belt in age, but instead, recognized five formations of the Purcell System, the Canadian equivalent of the Belt Supergroup. The Purcell System is divided into Lower and Upper Purcell, both of which are present in the Nelson Map area. It is of interest, as well as having important bearing on the regional distribution of Belt-Purcell facies, that Rice (1941, p. 11) was unable to recognize in the Nelson area the subdivisions of the Upper Purcell that were made by Schofield (1915, p. 36-38) in the Cranbrook district, which lies to the east and across the strike, but that he was able to correlate with subdivisions established in the Windermere area to the north (Walker, 1926, p. 7-11). The formations correlative with those in the Windermere area are the Dutch Creek and the overlying Mount Nelson. As identified from the published descriptions, these are the A and B belts of Daly.

Park and Cannon described (1943, p. 6) the Priest River Group as a "complex sequence of metamorphic rocks that include phyllites and schists, limestones, dolomites, quartzites, and volcanics." On their map (pl. 1) only quartzites and "volcanics" are separately indicated. To correlate this assemblage with Belt, Purcell, or Deer Trail rocks it was necessary to establish the chronologic relations within the sequence. Fortunately, since the time of Park and Cannon's mapping--of what was then a relatively inaccessible area--many miles of roads have been built, so that it has been possible with a minimum of effort to learn more about the Priest River Group.

The reconnaissance study of the Priest River terrane in the Metaline quadrangle was too brief to permit division of the group into formations or subunits or to trace out thin limestone or quartzite units that might serve as key horizons, however, from what was seen, it is possible to characterize the general lithology of the group. Because the Priest River rocks north of the Pass Creek fault zone are different from those to the south they are described separately.

The lower part of the Priest River section north of the Pass Creek fault zone is intruded by the Kaniksu batholith and the upper part is eroded and overlain by the Shedroof Conglomerate, consequently what remains of the group are 3000 metres or more of dominantly dark pelites, as argillite and phyllite, in an unknown intermediate position in the section. The lower one-third of the partial section differs from the upper part by being more limy and lacking the thin, but discrete, quartzite, limestone, and dolomite units found in the upper two-thirds of the section. Near the batholith the lower beds are metamorphosed to calc silicate hornfels and the thin limestone beds to marble. The carbonate units in the upper part are both dolomite and limestone, whose surface commonly weathers to a brown spongy crust containing variable amounts of residual quartz. Most quartzites, from light to dark gray in color, contain from traces to 30 percent or more of impurities of mica and calcite. Both limestone and quartzite are fine grained, a feature, along with thin beds, that typifies the Priest River rocks. The phyllites and argillites that form the bulk of the Priest River group are medium to dark gray and commonly laminated, particularly in the upper third of the section.

The Priest River Group south of the Pass Creek fault zone has the same dark phyllite enclosing occasional thin carbonate and quartzite units as does the unit to the north and in addition has at the top of the section over 500 metres of sandy dolomite overlain by 600 metres of coarse grained quartzite. The carbonate and quartzite clasts in the Shedroof Conglomerate are from these two units, which apparently were eroded from the Priest River section north of the Pass Creek fault zone. The dolomite--which also contains limestone--ranges from a carbonate rock that contains only a trace of quartz to a quartzite that contains only a trace of carbonate. This gradation between dolomite and quartzite is both across and along the beds. Another unusual rock in the southern section on the top of Grassy Top Mountain is a thin unit of phyllite containing small clasts of quartzite. This diamictite is very similar to those in the Shedroof Conglomerate.

The section of Priest River rocks south of the Pass Creek fault may be twice as thick as that to the north, but because of structural complexities in an area that has not been mapped in detail, this is uncertain. The difference in the two sections, both of which are overlain by the Shedroof Conglomerate, is compelling evidence for movement along the Pass Creek fault zone during the late Precambrian. The lower part of the southern section was not examined by the writer, nor was the contact of the Priest River and Prichard (Belt Supergroup) rocks.

The Priest River Group of Washington projects across the northern tip of Idaho into British Columbia where Rice (1941, p. 7-13) separated it into three units of the Purcell Series, which are from oldest to youngest, the Kitchener-Syeh Formation, Dutch Creek Formation, and Mount Nelson Formation. Below these units Rice mapped the Creston and Aldrich Formation, Canadian names for the lower Belt rocks, which apparently lie below the Priest River rocks that occur in the Metaline quadrangle north of the Pass Creek fault. The Kitchener-Syeh (Wallace Formation of the Belt Supergroup) may be equivalent with the lower calcareous unit north of the fault. It is described by Rice as variously colored, calcareous and dolomitic argillites and is overlain by the Dutch Peak Formation (Striped Peak Formation of the Belt Supergroup) a laminated argillite having beds of magnesian limestone and quartzite. The Dutch Peak Formation may be equivalent to the upper part of the section in the Metaline quadrangle north of the Pass Creek fault. The uppermost Purcell unit in the Nelson map area, the Mount Nelson Formation, might find its equivalent in the dolomite and quartzite unit that occurs south of the Pass Creek fault.

The Belt-Purcell depositional environment

The Priest River rocks are an extremely small part of a very thick, very extensive assemblage of mildly metamorphosed sedimentary rocks that were deposited in a basin[1] that extended at least as far south as the Snake River and at least as far north as 51° north latitude (fig. 16). In the United States the general distribution, lithology, and origin of the Belt rocks have been summarized by Hobbs and others (1965, p. 10-13) and in British Columbia the Purcell rocks have been summarized by Reeser (1957, p. 150-176) and later by Gabrielse (1972). The description of the Belt-Purcell anticlinorium that follows is largely a paraphrase of--but without specific references to--these summaries.

The limits of known outcrops of Belt-Purcell rocks are indicated on figure 16. This, however, is only the minimum size of the depositional basin, because shore facies are identified only south of Butte, Montana, where basal beds of Belt rocks containing much coarse clastic material rest upon older crystalline Precambrian rocks. The east margin present of the outcrop area, both in Montana and in British Columbia, however, probably lies not very far west of the eastern limits of the Belt-Purcell sea. The southern and northern limits of the basin are hidden by younger rocks. The westernmost Belt-Purcell outcrops are unconformably covered by Late Proterozoic Windermere rocks.

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[1] Basin is used in the sense of Sloss, et al, Geol. Soc. America Mem., 39, p. 100, 1949: "certain areas of the cratons which have been persistently... subject to sinking greater than that of the surrounding shelves."

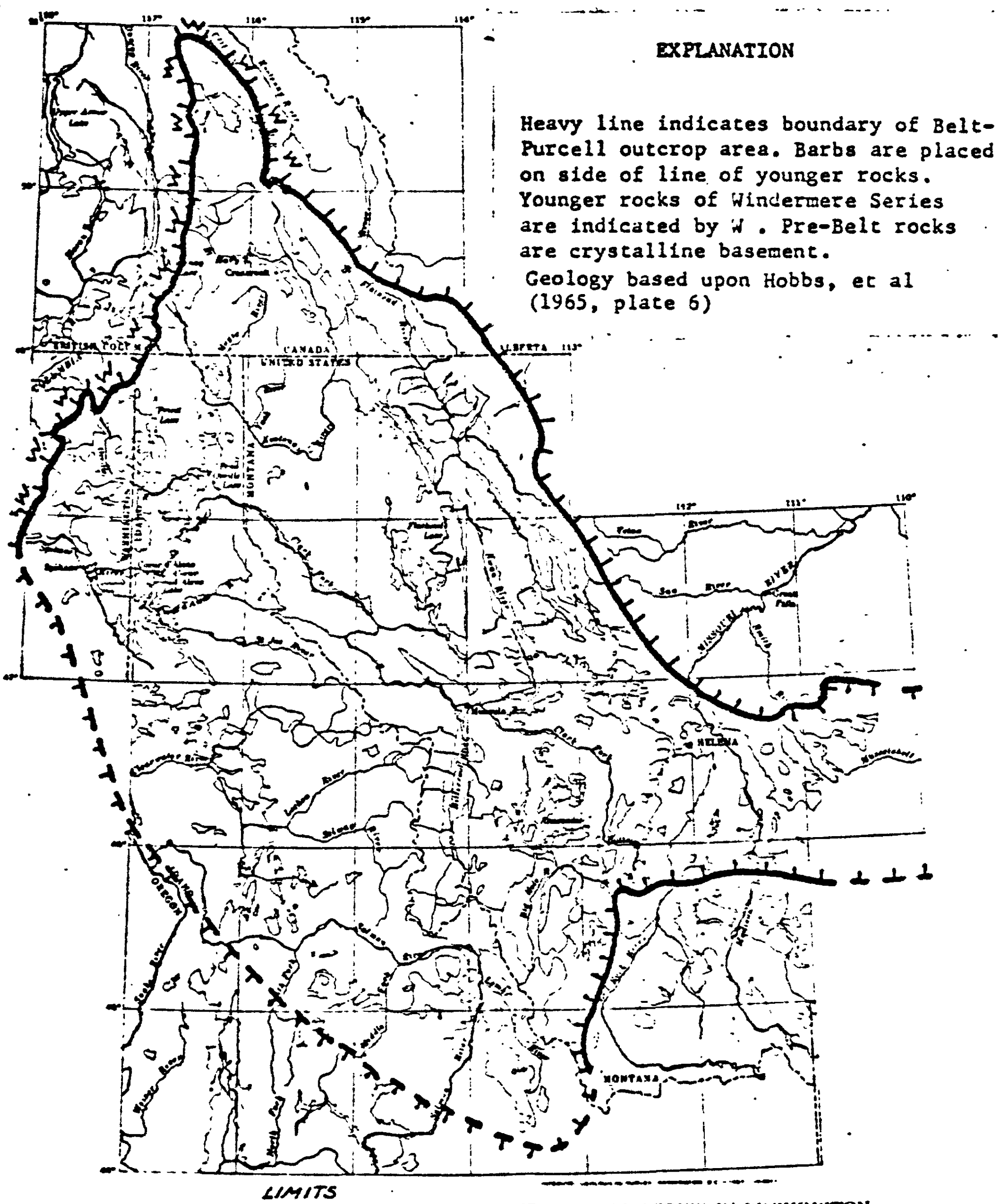

Fig. 16— OUTCROP LIMITS OF THE BELT AND PURCELL SERIES IN WASHINGTON IDAHO, MONTANA, BRITISH COLUMBIA, AND ALBERTA

Within this basin accumulated from 14,000 to 23,000 metres of dominantly clastic sediments. The argillite, quartzite, and siltite that compose the clastic rocks are intercalated with lesser amounts of dolomitic carbonate rocks. Other, much less common, constituents are cherts, which occur in the western, upper part of the section, and basaltic lava flows, which are used to subdivide the Purcell Group into an upper and lower part and which occur in the upper middle part of the Belt Supergroup.

Characteristics of the clastic sediments are fine-grained texture--which applies equally to the quartzites--, thick lithologic units that commonly are gradational into one another both vertically and horizontally, and the presence of sedimentary structures that indicate a shallow water environment of deposition.

The fine-grained texture of the sediments and their simple, almost universal mineralogy of quartz, sericite (and/or chlorite or biotite) and feldspar gave rise to beds of quartzite, siltite, and argillite, which in various combinations form mappable units. To break down the group into formations it is necessary to base division on arbitrary combinations of the three lithologies, a practice which, in an area where horizontal gradation is common, creates problems in nomenclature from one part of the basin to another. Fortunately the fourth lithology, carbonate rocks occur everywhere in about the same position in the sequence.

The nature of the depositional environment is interpreted mainly from the shallow water features, the great thickness, the fine-grained character, the sparcity of disconformities, the monotony of lithologies, and the greater continuity of stratigraphic units in a north-south direction than in an east-west direction. From these characteristics, the authors paraphrased generally agreed that the Belt-Purcell sediments accumulated in a slowly subsiding basin of great tectonic stability and that the indicated environment most closely resembles the delta of a large river, a river that would have traversed great distances across a terrain of low relief.

Those most familiar with the Belt-Purcell rocks do not agree on the provenance of the sediments. Reesor (1957, p. 156-158), after evaluating the evidence found in the Purcell rocks of British Columbia, concludes, "It thus appears that a known, distant, relatively technically stable source area existed throughout the deposition of the Purcell-Belt sediments to the north and east of the known present outcrop area of these sediments. No known source exists to the west and no evidence is found within the stratigraphic column which indicates a western source." The authors of the report on the Coeur d'Alene mining district (Hobbs, Griggs, Wallace, and Campbell, 1965, p. 10-11) favor an opposing view, "The position of the west margin of the basin can only be inferred for it is masked by younger rocks. Furthermore, whether the basin was here bounded by a land mass or by an open sea can only be conjectured. In the Coeur d'Alene district part of the Belt sequence becomes more quartzitic to the west, evidence suggesting a landmass source to the west. A higher percentage of carbonate-bearing rocks throughout the Belt Series section in the east central part of the basin, of which the Glacial National Park area (Ross, 1959) is an example, also suggests that the dominantly clastic rocks to the west had a western source."

Those favoring a western source have the more difficult position to maintain. A western source requires a landmass of crystalline, non-oceanic crust that was large enough to support a major river system; this is, an area continental in size. There is no evidence of, nor need for, such a landmass to exist after the Precambrian, because it is well established that an eastern landmass, the Canadian Shield, was a positive area that was shedding detritus westward not only during the Proterozoic, but during the lower Paleozoic as well. Nothing in the literature indicates that the clastic Belt rocks on the eastern margin of the outcrop area are basically different from those on the western margin, which, if two sources are involved, is unusual.

Windermere rocks

As we move west and cross the unconformity at the top of the Priest River rocks, we pass into one of the most interesting assemblages of rocks in the Kootenay arc. These are some 4,500 metres of young Proterozoic rocks that grade upwards into the Cambrian; they comprise the Windermere Group. The unconformity between the Windermere System and the Belt Supergroup records what White (1955, p. 62-63) refers to as the "East Kootenay orogeny," an event that was, in northeastern Washington, more epeirogenic than orogenic. Later White (1966, p. 186) termed it the mid-Proterozoic epeirogeny. Unfortunately this term does not permit distinction between the three Proterozoic epeirogenic events, consequently to avoid confusion I substitute Kootenay epeirogeny for White's terminology.

Projected into the Metaline quadrangle, Little's (1960) upper limit for the Windermere, the Cambrian-Precambrian boundary, falls within the Gypsy Quartzite, the oldest formation containing Lower Cambrian fossils. In this report for reasons of simplicity the upper limit of the Windermere is placed at the base of the Gypsy Quartzite and the top of the Monk Formation. Therefore in the Metaline quadrangle Windermere includes three of Park and Cannon's formations (1954, p. 7-13), the Shedroof Conglomerate, Leola Volcanics, and the Monk Formation. In variance with this report, Park and Cannon (p. 13) considered the Monk Formation to be of questionable Cambrian age because it "is conformable with and grades upward into the Gypsy Quartzite which is definitely of Cambrian age." Since Park and Cannon's work, Okulitch (1948, p. 340-344) has found early Cambrian archaeocyathids at the top of the Gypsy Quartzite, which is over 2,400 metres stratigraphically above the top of the Monk, a fact that strongly suggests that the Precambrian-Cambrian boundary is somewhere within the Gypsy Quartzite, where Little places it.

It is not fallacious to include the diverse assemblage of rocks that compose the three formations of Windermere age, the Shedroof Conglomerate, Leola Volcanics, and Monk Formation, into a formal stratigraphic group (Miller and McKee, 1973), for despite their diverse and contrasting lithologies, they are relatable parts of a sequence that has a homogeneity, which, although not readily recognizable, is nevertheless very real. This is exhibited by the repetition of distinctive lithologic characteristics throughout the sequence. From oldest to youngest the predominant lithologies represented are the conglomerates of the Shedroof, the mafic volcanic rocks of the Leola, and the grab bag of phyllite, quartzite, dolomite, limestone, and conglomerate of the Monk Formation. Conglomerate, however, is not restricted to the Shedroof, but occurs in the Leola and in the Monk; each younger conglomerate has clasts similar to those of the earlier conglomerates, plus clasts derived from the immediately underlying unit. The consanguinuity of the formations is also expressed in the similarities of quartzites and dolomites that occur in the Shedroof Conglomerate and Monk Formations. This lithologic kinship extends even beyond the Windermere rocks, to the rocks of the Priest River Group, for here the dolomites are also the sandy ferruginous variety and the one conglomerate observed is indistinguishable from the Shedroof Conglomerate. For the above reasons all correlations and origins of the Windermere rocks are given collectively after the descriptions of the individual formations.

The Shedroof Conglomerate

The Shedroof Conglomerate, which unconformably overlies the Priest River rocks, is an unusual conglomerate because of its bimodal character, unsorted clasts--many measured in feet--and a great, but highly irregular thickness. The conglomerate is readily separable into two distinct parts, a matrix of phyllite, and a highly variable concentration of clasts. The matrix is largely a dark gray to green, very fine-grained phyllite composed of muscovite, chlorite, and quartz, with muscovite the most abundant constituent. The clasts, which are of great size range, are largely dolomite and quartzite. The ratio of matrix to clasts ranges between wide limits, from rock of closely packed clasts to clastless phyllite. Bedding is difficult to detect and is commonly masked by one or two cleavages and a strong foliation resulting from tectonic flattening and stretching of clasts. The Shedroof Conglomerate grades into the overlying Leola Volcanics. Gradation is the upward increase in volcanic detritus to the upper beds of the Shedroof.

As mentioned above not all the formation is a conglomerate. Park and Cannon (1943, pl. 1) separated out phyllite, thin quartzites, and greenstone and mentioned (p. 7) observing beds of sandstone and dolomite. An outcrop area more than 2 1/2 kilometres wide of phyllite, located north of the Pass Creek fault, essentially free of clasts is shown on Park and Cannon's plate 1 (also on fig. 14 of this report). This unit pinches out to the northeast. Where it crosses Stony Creek it contains one 30 metre quartzite unit; conglomerate above the phyllite contains two quartzite units and one unit of thin-bedded gray limestone. Apparently clasts were being introduced into an environment where shale, sandstone, limestone, and dolomite were being deposited. The excellent recently-made road cuts show conglomerate everywhere interbedded with clast-free phyllite. It is probable that the depositional pattern consists of tongues of conglomerate extending into continuously accumulating muds.

As mentioned above not all the formation is a conglomerate. Park and Cannon (1943, pl. 1) separated out phyllite, thin quartzites, and greenstone and mentioned (p. 7) observing beds of sandstone and dolomite. An outcrop area more than 2½ kilometres wide of phyllite, located north of the Pass Creek fault, essentially free of clasts is shown on Park and Cannon's plate 1 (also on fig. 14 of this report). This unit pinches out to the northeast. Where it crosses Stony Creek it contains one 30 metre quartzite unit; conglomerate above the phyllite contains two quartzite units and one unit of thin-bedded gray limestone. Apparently clasts were being introduced into an environment where shale, sandstone, limestone, and dolomite were being deposited. The excellent recently-made road cuts show conglomerate everywhere interbedded with clast-free phyllite. It is probable that the depositional pattern consists of tongues of conglomerate extending into continuously accumulating muds.

Based on clast composition there are two basic intergrading lithologic types of conglomerate; that dominated by quartzite clasts and that dominated by dolomite clasts. The dolomite clasts so closely resemble the bedded dolomite in the Priest River Group that there is little doubt that this is the source. The quartzite clasts, which include well-rounded pebbles and cobbles and stretched and flattened clasts, were derived mainly from white to light gray, medium- to coarse-grained quartzites. The quartzite of some clasts consists of well-rounded, well-sorted quartz grains; that of others is so thoroughly recrystallized that original grain boundaries are no longer recognizable.

Although by far the greatest number of clasts are quartzite and dolomite, other types are sparsely represented. Park and Cannon (1943, p. 7) say "a few pieces of black slate or phyllite and a single pebble of granitic rock were seen in the conglomerate." I noted similar clasts of black phyllite, but failed to find any granitic pebbles although they were carefully searched for. Two cobbles of feldspathic gneiss, however, were collected. These are metamorphosed arkosic sediments of obscure provenance.

The clasts are clearly of two origins, sedimentary and tectonic. The well-rounded and the exotic clasts are clasts transported by sedimentary processes. Some flattened and angular clasts are demonstrably boudined from thin beds of dolomite and quartzite. What percentage of the stretched clasts are of this origin is not determinable. In summary, there is little doubt that most sedimentary clasts were derived from the underlying Priest River Group, that very similar clasts were boudined from dolomite and quartzite beds in the phyllite and now appear as stretched clasts, and that the rare gneiss clasts have no known local provenance.

The origin and interrelations of the Windermere conglomerates are regional problems that are discussed at the close of this section; however, certain features and associations of the conglomerate, although genetic in tone, are included here to complete and accent the description of the conglomerate. In searching for an explanation of the conglomerate, Park and Cannon (1943, p. 9) list six points that relate to its origin. These are:

"(1) Its association and gradation into greenstone is nearly invariable.

(2) Its thickness is extremely inconstant.

(3) The fragments comprise (a) quartzite, dolomite, phyllite, schist, rarely granitic rocks, thought to be derived from Priest River Group, and (b) small round quartzite pebbles and cobbles mixed with large angular limestone fragments. Volcanic fragments are absent, except in the transition zones to the Leola Volcanics.

(4) The matrix consists of sericite, quartz, carbonate, chlorite, and biotite.

(5) There is nearly complete lack of sorting; bedding is shown by a few sand and dolomite beds.

(6) There may be a stratigraphic hiatus below the Shedroof Conglomerate."

To this list can be added:

(7) The high ratio of matrix to clasts; much of the matrix is free of clasts; a considerable part of the conglomerate qualifies as a diamictite.

(8) Pebble, cobble, and boulder-sized clasts were deposited in an environment where mud and occasional beds of sand and carbonate were accumulating.

(9) The extreme contrast in clast angularity.

(10) The similarity between dolomite bedded in the matrix, dolomite in the Priest River Group, and the dolomite clasts.

Some of Park and Cannon's 6 points can be amplified with recently acquired data.

(1) The association between greenstone and Windermere conglomerate is a local association which extends only a few miles north of the International Boundary.

(2) The inconstant thickness is in large part the result of contemporaneous faulting and erosion. This is discussed in a later section.

(3) The matrix is compositionally similar to the pre-Windermere and Cambrian phyllite, suggesting similar depositional conditions.

(4) The bimodal character is in itself a crude sorting, but probably not by the same agency.

(5) The stratigraphic hiatus at the base of the conglomerate is well confirmed.

The Shedroof Conglomerate has been correlated with the Toby Formation of British Columbia, which separates the Purcell Group (Belt Supergroup) rocks from the Windermere Horsethief Creek Group rocks. This correlation is discussed on pages 159-161. The Toby conglomerate and its correlatives, Shedroof Conglomerate and Huckleberry Conglomerate of the Chewelah quadrangle was studied regionally by Aalto (1971) who came to the conclusion that it was of glacial origin related to a continental ice mass that moved eastward from the cratonic area. Earlier Canadian workers including H. M. A. Rice, H. W. Litt.e, G. B. Leech and others (Little, 1960, p. 16) considered the glacial hypothesis and rejected it. Reesor (1973, p. 38-39) after studying the Toby Formation in the Lardeau Map area, east-half, British Columbia "does not favor a glacial interpretation" for the origin of the Toby Formation." His argument is that although "lithologic equivalents for many beds of the Toby Formation may be found in Pleistocene glacial and postglacial deposits. ...It is in the search for specific, unequivocal evidence of glaciation ... such as faceted pebbles, striated boulders, or striated pavements that the idea of a glacial origin fails."

Aalto (1971) in presenting his argument for glaciation reduces competing hypothesis to what he considers the most likely, submarine mass flow. He concludes, after analysis of the pros and cons, that evidence for a glacial origin outweighs that for submarine mass flow. He presents none of the positive evidence that Reeser demands. The diamictites of the Metaline quadrangle, because of extreme tectonic distortion do not lend themselves to solving this problem.

Leola Volcanics

The Leola Volcanics along with the Irene Volcanics of southern British Columbia and the Huckleberry Volcanics of the Chewelah quadrangle form a belt of tuffaceous sediments, flows, and breccias that extends discontinuously from 24 kilometres north of the International Boundary southwestward for 137 kilometres. At its north end it is unconformably overlain by the Horsethief Creek (Monk) Formation and at its south end unconformably overlain by the Addy (Gypsy) Quartzite. The volcanic event that produced the flows and pyroclastics of the Leola was an interruption and not a termination of acculation of the Windermere conglomerates of the post-Belt epeirogeny, which began with the uplift and erosion of the Belt-Purcell anticlinorium. The conglomerate at the top of the Leola Volcanics and at the base of the Monk Formation probably includes the last of the coarse clastics derived from the anticlinorium. The quartz sand in the upper part of the Monk Formation and that in the lower Cambrian quartzites was not derived from the anticlinorium but from some more distant source.

The Shedroof Conglomerate grades into the overlying Leola Volcanics through a transition zone several hundred feet thick. The color of the matrix changes from gray to green, the clasts become smaller, and the bedding more distinct as the formation grades upward to clastless green schist or phyllite. Daly (1912, p. 144) describes lava flows along the International Boundary as interbedded with the conglomerate for "a hundred feet or more." Little (1960, p. 15) on reexamining the section along the International Boundary, however, found the top of the Toby (Shedroof) to be "interbedded green argillaceous schist and conglomerate" without recognizable lava flows. Park and Cannon (1943, p. 10) regarded the green schists in the Leola Conglomerate as "of probable tuffaceous origin," whereas Daly's (1912, p. 145) columnar section of 1800 metres of the Irene Volcanic Formation lists the lower 1200 metres as "sheared and greatly altered basaltic and andesitic lavas, largely greenstone schist," but mentions no tuffaceous rocks except a few subordinate layers of basic tuff. In the East Nelson Map area, to the northeast of Daly's and Little's observations, Rice (1941, p. 16) recognized tuffaceous sediments. To the southwest in the Chewelah quadrangle, Miller and Clark (1975) identified but minor amounts of tuff, although flow breccias are common.

The volcanics in the formation have been variously described. Daly ignores the alteration that masks the true character of the lavas by calling them basaltic and andesitic lavas and describing them as containing labradorite, and Rice (p. 16) identifies the "present composition" as andesite with the only identifiable plagioclase as andesine associated with actinolitic amphibole. Little is in general agreement, considering the plagioclase andesine-labradorite in composition. In the Chewelah area, Miller and McKee (1973) found the lava to be a basalt and the plagioclase to be albite.

Chemical analyses of the lavas are shown in table 2.

Table 2.-- Analyses, in percent, of volcanic flows of Late Precambrian age in the Chewelah quadrangle, Wash.

	85	89	90	98	99
Chemical analyses					
SiO_2	47.5	48.4	37.4	44.8	48.9
Al_2O_3	13.4	11.5	12.1	13.2	12.2
Fe_2O_3	2.4	2.8	.47	1.0	4.6
FeO	10.9	10.1	13.0	11.5	7.3
MgO	7.5	3.2	3.3	6.3	3.4
CaO	6.7	9.0	14.5	6.6	9.1
Na_2O	2.1	3.1	2.2	2.8	3.7
K_2O	.90	.00	.05	.85	.34
H_2O-	.43	.15	.15	.18	.21
H_2O+	4.4	3.3	4.0	4.3	3.2
TiO_2	2.3	2.7	3.3	2.8	2.9
P_2O_5	.25	.43	.63	.45	.70
MnO	.22	.19	.22	.21	.25
CO_2	.24	5.4	8.6	4.8	3.2
Total	99.	100.	100.	100.	100.
CIPW norms					
qtz	2.25	16.70	5.29	8.69	10.83
or	5.36		.30	5.03	2.01
ab	17.91	26.16	18.63	23.74	31.31
an	24.67	7.68	13.46		15.68
ne					
wo	2.36	(c) 3.57	(c) 3.50	(c) 7.69	1.95
en	18.82	7.95	8.23	15.53	8.47
fs	14.76	12.10	18.46	16.09	5.28
fo			0.0		
fa			0.0		
mt	3.51	4.05	.68	1.45	6.68
hm			0.0		
il	4.40	5.11	6.27	5.33	5.51
ap	.56	1.02	1.49	1.07	1.67
cc	.55	12.25	19.57	10.74	7.28
Total	95.15	96.58	95.88	95.54	96.63

Analyzed by x-ray fluorescence by Paul Elmore, Sam Botts and Lowell Artis.

The Leola Volcanics, however is not all volcanic rocks. Much of the fine-grained clastic rocks the writer examined near the top of the unit were, upon examination under the microscope, a mixture of volcanic and nonvolcanic detritus. The fact that the green (chlorite) schists of the Leola cannot--as Park and Cannon state (1943, p. 10)--"be distinguished with assurance from certain greenschists or phyllites in the Priest River Group, in the phyllite facies of the Shedroof Conglomerate, in the Monk Formation, and even in the much younger Maitlen Phyllite" indicates that volcanism was of long duration in some nearby area, or that chlorite-bearing phyllites are not, per se, reliable indicators of volcanism. This questionable suggestion that volcanism continued for long periods both before and after Leola time is much less positive than the evidence that sedimentary processes accompanied Leola volcanism in the same basin of accumulation. Similar but less abundant beds of limestone, dolomite, quartzite and conglomerate occur in the Leola Volcanics as in the Shedroof Conglomerate. On the Boundary, Daly (1912, p. 146) mentions a 12-metre unit of fine-grained dolomite. Rice (1941, p. 16) in the East Nelson map area found limestone and conglomerate beds 15 to 30 metres wide.

The tuffaceous beds near the top of the Leola Volcanics grade into a conglomerate, which is included here, as it is in the West Nelson Map area, as the basal member of the Monk Formation. This conglomerate closely resembles the Shedroof Conglomerate, except that it contains volcanic detritus from the underlying volcanic rocks, thus indicating an erosional unconformity between the Leola Volcanics and the Monk Formation.

Monk Formation

The Monk Formation, predominantly phyllite, occurs in the southeastern part of the Metaline quadrangle as a belt that extends from the International Boundary to the Pass Creek fault zone, a distance of 24 kilometres. It continues north of the boundary, across the southeast corner of the West Nelson Map area under the name of Monk, and into the East Nelson Map area where it becomes the unnamed lower part of the Horsethief Creek Series (Rice, 1941, p. 17-18), which has been mapped to the north for many kilometres. The Monk also appears on the west side of the Pend Oreille River, as the lowermost formation brought up along the Flume Creek Fault. Rocks believed correlative with the Monk Formation appear between the Huckleberry Volcanics (Leola Volcanics) and the Addy Quartzite (Gypsy Quartzite) in the Chewelah quadrangle (Miller, oral communication, 1974).

The Monk Formation was named by Daly (1912, p. 147-150) and described from observations made "on a traverse between Monk Creek and the boundary along the top of a ridge running east-southeast from Mt. Ripple." Daly remarks on the poor exposures of this formation. He estimates its thickness as 1700 metres and emphasizes--as all later workers do--the heterogeneity of the formation, which he describes as consisting of predominant quartz-sericite schists and argillites that contain numerous intercalations of quartzite grit and conglomerate. Although he fails to mention dolomite, which is a prominent constituent south of the Boundary, this rock is probably present but not exposed in the heavily timbered area of Daly's traverse.

When Walker (1934) mapped the Salmo Map area (included in the West Nelson Map area), which includes the area of Daly's Monk Formation, he renamed the formation the Horsethief Creek Series, because of its similarity to the Horsethief Creek Series of the Windermere area, 160 kilometres to the northeast. He expanded the unit to include a 15 metre band of schist and a 60 metre band of conglomerate that Daly had placed in the Irene Volcanic Formation. Later workers, including Rice (1941), who mapped the East Nelson Map area, which includes the northeast extension of the Precambrian rocks of the Metaline quadrangle, followed Walker's terminology. However Rice (p. 18-19) found that the unit overlying the Horsethief Creek Series, the Three Sisters Formation, which was separable from the Horsethief Creek Series in the Salmo area, was not separable in the East Nelson area; consequently he included strata equivalent to the Three Sisters Formation under the Horsethief Creek Series and by so doing deviated greatly from the original definition of the series in the Windermere area. Little (1960, p. 18-20) because of this confused use of the name, Horsethief Creek Series, refrained from its use in the West Nelson area and revived Daly's "Monk Formation."

Little (1960, p. 19) describes the Monk as beginning with a 90-metre basal conglomerate (the conglomerate Daly placed in the Irene Volcanics) having clasts of pisolitic magnesian limestone and lesser clasts of greenstone, dark quartzite, and argillaceous schist that has a a pitted surface, followed by a unit of fine-grained argillaceous schist containing a 60 metre band of gray limestone. About 300 metres above the limestone, grits appear and become increasingly more abundant upwards. To the south in the Metaline quadrangle these grit beds are proxied by limestone.

Rice's description of the Horsethief Creek (Monk Formation) rocks in the East Nelson Map area is in close agreement with that of Walker (1926, p. 14) for the type locality in the Windermere area. Walker describes the unit as largely "gray, green, and purplish slate with several lenticular beds of coarse quartzite and pebble conglomerate... numerous thin interbeds of blue-gray crystalline and mostly non-magnesian limestone." Rice (p. 17) found the argillite northeast of Kootenay Lake to be "generally slaty, finely laminated or uniform, mostly dark grey to black but some beds...greenish grey...beds of blue grey, crystalline, essentially non-magnesian limestone are in places conspicuous... Some beds of buff weathering dolomitic limestone also occur." The "series" also includes occasional beds of white quartzite, some of which are limy. Beds of conglomerate occur all through the series. Clasts are rounded pebbles or cobbles of quartzite and occasional slaty fragments in a siliceous matrix. Feldspar clasts are common. Some beds are composed of angular blocks of sandy magnesian limestone in a cement of similar composition. Beds of blue gray limestone conglomerate also occur.

A stratigraphic section made by Rice (1941, p. 18) west from Rose Pass (80 kilometres north of the Boundary) shows that conglomerate--mainly limestone conglomerates--occur throughout the lower 1200 metres of a 2966 metre section, which Rice believed to be nearly doubled by faulting. Accepting the persistence of conglomerates through a minimum section of 600 metres of beds lying on the 90 metre thick Toby Conglomerate (the conglomerate that marks the ast Kootenay epirogeny in southern British Columbia, one becomes aware that the unusual conditions that produced the diamictite of the Toby Conglomerate continued along with the concomitant deposition of fine-grained argillites as well as quartzites and limestone.

It is evident that here in the lower Horsethief Creek (Monk), as in the Shedroof Conglomerate to the south, normal marine deposition and the accumulation of diamictite were geographically contemporaneous processes.

This diversity of lithology also characterizes the Monk Formation in the State of Washington. Park and Cannon (1943, p. 12) describing the Monk lithology in the Metaline quadrangle point out, "the details of lithology shown by nearly any hand specimen or within a limited area of exposure find their analogs in nearby younger and older rocks, particularly in the rocks of the Priest River Group." With this introduction they described the formation as predominantly phyllite that contains numerous intercalations of carbonate rocks, quartzite and grit. The conglomerate that separates the Monk from the underlying Leola Volcanics contains clasts of volcanic rocks but in other respects strongly resembles the Shedroof Conglomerate, except for being much thinner, from 60 to 120 metres. The writer was able to trace the basal Monk conglomerate as far to the southwest as Sullivan Creek, where it was lost in the complex of faults.

The phyllite of the Monk Formation in the Metaline quadrangle is similar to the phyllite of the Shedroof Conglomerate and to the Cambrain Maitlen Phyllite in consisting of fine-grained quartz, sericite, and lesser chlorite. The dolomite closely resembles the white, fine-grained, sandy dolomite of the Priest River assemblage and the quartzites and grits resemble those in the overlying Gypsy Quartzite. The thin bedded, white and gray banded limestone has its counterpart in both the Priest River and Shedroof rocks.

The band of Monk Formation stretching southward from the International Boundary to Sullivan Creek changes markedly along strike. On the ridge that connects Salmo Mountain with Shedroof Mountain, the Monk is excellently exposed in road cuts. The basal conglomerate, 107 metres in thickness, is overlain by a gray phyllite that contains beds of diamictite containing dolomite, greenstone and quartzite clasts, along with beds of intraformational dolomite conglomerate, and beds of banded limestone. This phyllite-carbonate unit, roughly 450 metres thick, contains in its upper half, thin-bedded quartzite and dolomite containing spherical grains of quartzite. Above this lies 210 metres of dark gray slate, microlaminated and containing at least one dark gray limestone unit. More gray limestone and sandy limestone overlies the laminated phyllite. The sandy limestone is succeeded by a 60-metre band of conglomerate composed of rounded quartzite cobbles in a sandy matrix.

The uppermost 600 metres of the section consists of phyllite with intercalations of brown-weathering quartzite. The contact with the Gypsy Formation was placed at the uppermost phyllite beds, but here this coincides with a shear zone that may represent fault movement cutting out a moderate thickness of Monk and(or) Gypsy Quartzite. The thickness of this section of Monk is estimated to be 1800 metres.

The Salmo-Shedroof section described in the preceding paragraph contrasts greatly with the Kinyon Creek section 10 kilometres to the southwest. This section has an estimated thickness of only 1200 metres and the kind and sequence of lithology is drastically different. The lower two-thirds of the section, which rests on the basal Monk conglomerate, is phyllite that contains two or three thin limestone units and one thin quartzite unit.

It has none of the diamictite conglomerates and intraformational breccias common to the northeast. The upper third of the formation is a massive white dolomite that is sandy near its contact with the overlying Gypsy Quartzite.

How much variation in lithology along strike results from tonguing, how much from faulting, and how much from an angular unconformity at the top of the formation will not be known until the Monk is mapped in considerably more detail. Nevertheless, enough mapping is done to clearly indicate that thinning and pinching of lithologic units is a characteristic of the Monk assemblage, and that relations between the Monk and Gypsy are unconformable.

On the other hand the Monk rocks in the Chewelah 30-minute quadrangle have a closer resemblance to the correlative Horsethief Creek Group, of the type Windermere area in British Columbia than do those in the Metaline quadrangle. The slates are characteristically purple and green, a feature Walker (1926, p. 14) describes in the Windermere Map area. The Monk is the Chewelah quadrangle is a fragmented unit, appearing as small incomplete isolated sections. According to Miller and Clark (1975, p. 26), the three outcrop areas of Monk Formation are separated by alluvium and faults. At no place is there more than 150 metres of section, which is everywhere a dominantly slate-argillite lithology that has prominent dolomite beds in its lower part. To the southwest of Chewelah, the greenstone (Huckleberry) and dolomite beds are separated by a conglomerate that appears to thicken to the south.

Problems on the correlation of conglomerates

The Windermere volcanics, which overlay the Shedroof Conglomerate most logically can be interpreted as an eastward pinching wedge whose upper surface, at least in its eastern part, is a surface of erosion, an unconformity, overlain by the basal conglomerate of the Monk and whose lower surface rests upon the Shedroof Conglomerate. In British Columbia, somewhere north of latitude 49°15' before the emplacement of plutonic rocks, this wedge of volcanics, which is here called the Irene Volcanics, disappears beneath an overlap of younger rocks; it is probably present under the rocks west of Kootenay Lake. In the latitude of 40°15' and east of Kootenay Lake, a conglomerate at the base of the Horsethief Creek Series occupies the same stratigraphic position the Monk Conglomerate occupied to the south and--because the volcanics are absent--also occupies the same stratigraphic position as the Shedroof Conglomerate (fig. 16A). This conglomerate, known as the Toby Formation, interrupted only by Mesozoic plutons, continues on to the north into the Lardeau Map area, east half. Here Resser (1973, p. 23-24) reports that in the drainage of upper Frances Creek the conglomerate contains a volcanic flow, and several kilometres to the south contains greenstone cobbles--as does the Monk Conglomerate 130 kilometres further to the south. The Toby Formation is widespread throughout southeastern British Columbia, and has long been considered the temporal correlative of the Shedroof Conglomerate; however, this correlation does not stand up under analysis.

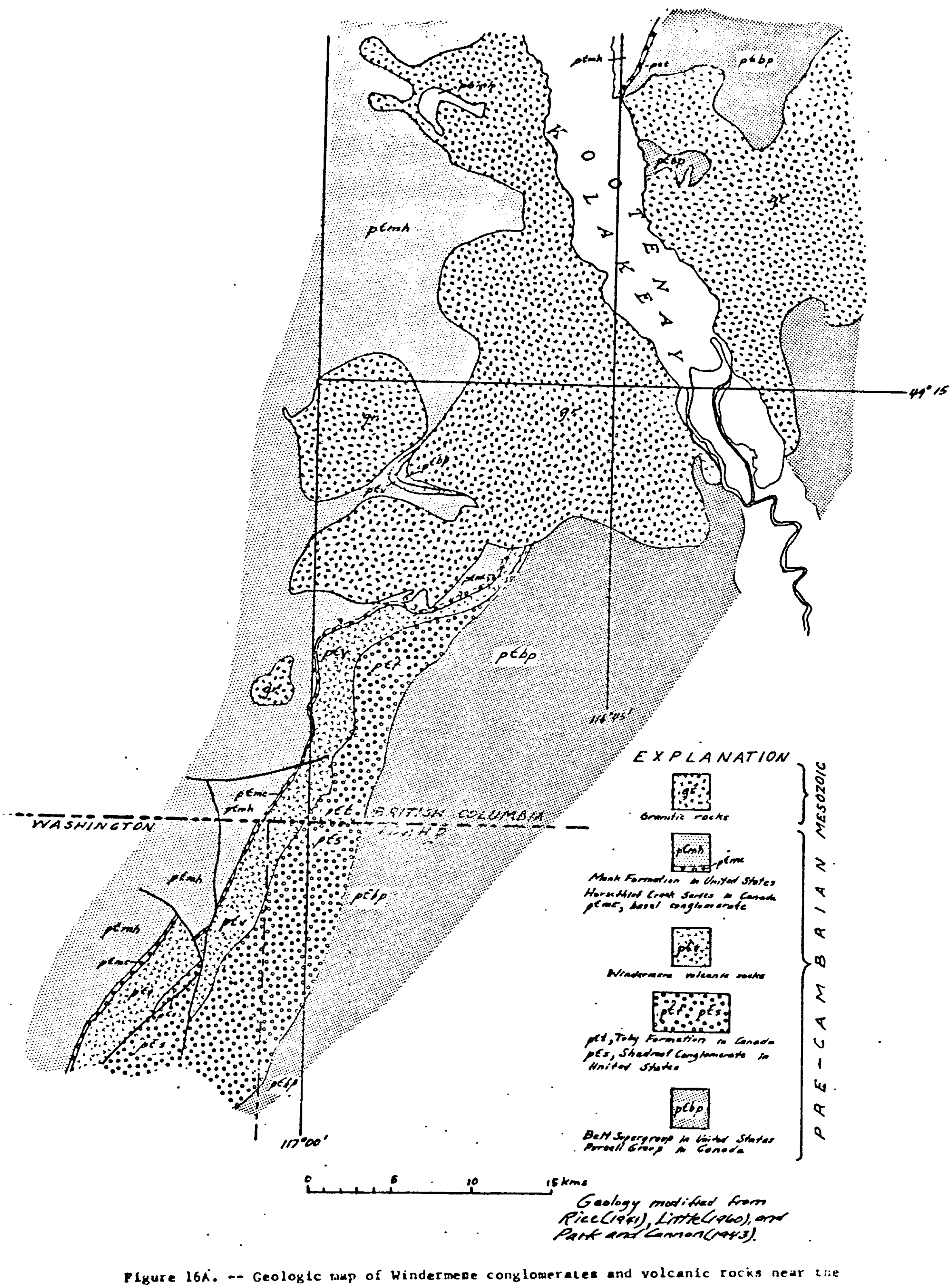

Figure 16A. -- Geologic map of Windermere conglomerates and volcanic rocks near the International Boundary. Basal conglomerate of Monk Formation is projected east of longitude 117° and south of batholith.

A more logical correlation is that the Toby Formation is the stratigraphic equivalent of the basal conglomerate of the Monk Formation. Admittedly there is some pedanticism in this change in correlation, because the removal of the volcanic wedge from the section juxtaposes two similar conglomerates and they become one. The question then becomes, "is the Toby Conglomerate equivalent to all, or to only the uppermost part of this combined unit?" The answer lies in the kind of process that allows the preservation of a conglomerate such as the Toby. The Toby Conglomerate where it appears north of the pluton at the south end of Kootenay Lake as a 60 metre unit occurs free of the stratigraphic confusion created by the wedge of volcanics. It extends from the lake northward for 240 kilometres to the Dogtooth Mountains and eastward for 120 kilometres to the Cranbrook area. Within this triangular area (fig. 17) formed by Kootenay Lake, the Dogtooth Mountains, and Cranbrook--roughly the northern half of the Belt-Purcell anticlinorium--the Toby appears in downfolded and downfaulted structures sufficiently closely spaced to indicate that the conglomerate was once a continuous sheet, which ranged from 15 to 1610 metres in thickness. Everywhere it rests on rocks of the upper Purcell.

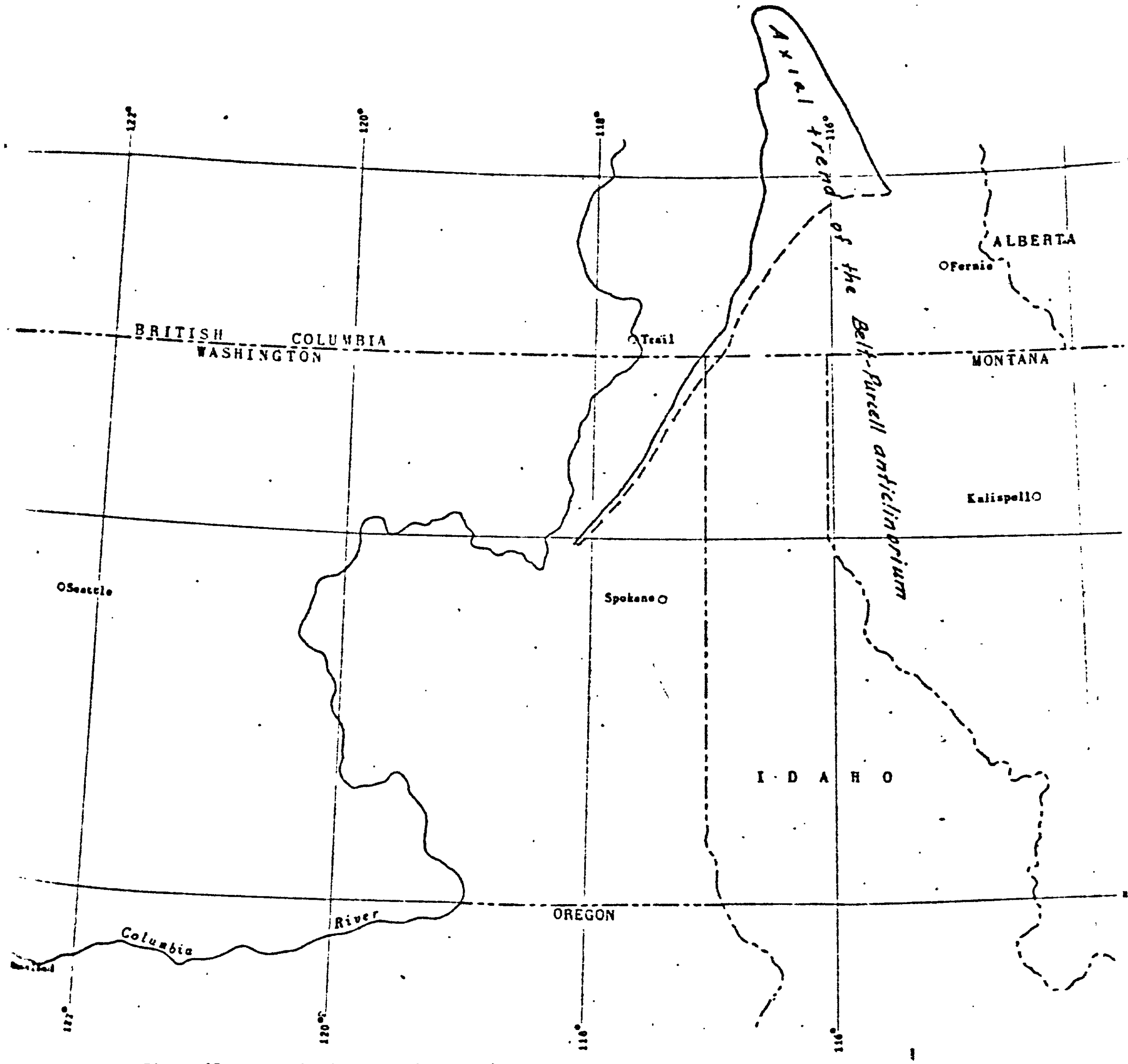

Figure 17.-- Map showing distribution of Windermere conglomerates. Dashed and solid lines enclose area of conglomerate outcrops. Solid line separates the outcrop area from younger rocks; dashed line seperates it from older rocks.

A high energy sediment like the Toby Formation could not have been deposited simultaneously over the whole northern half of the anticlinorium; its broad distribution requires a transgressive sea--regardless of whether deposition was by glacial or other agencies. Assuming transgression eastward, the conglomerate becomes younger in this direction and assuming penecontemporaneous burial--which is essential for preservation--the covering Horsethief Creek beds must also be younger eastward and a transgressive formation. It is highly probable that some of the higher conglomerate beds in the section of Horsethief Creek described by Rice (1941, p. 17-18) at Rose Pass east of Kootenay Lake are composed of clasts from reworked Toby that was deposited further east and redistributed by regressive oscillations. Following this reasoning, one concludes that the Toby Conglomerate represents a late stage of Windermere diastrophism being a basal conglomerate to the Horsethief Creek rocks and that it includes no beds that are time equivalent to those of the Shedroof Conglomerate, which is older, representing the earliest, most westerly, depositional record of Windermere diastrophism.

Cambrian Rocks

Quartzite

In Washington the quartzite facies[1] of the Cambrian is a single formation, known as the Gypsy Quartzite in the Metaline and Colville quadrangles and as the Addy Quartzite in the Chewelah 30-minute, Hunters and Turtle Lake quadrangles. Across the International Boundary in British Columbia it is divided into three formations: a lower conglomerate-grit unit called the Three Sisters Formation, an intermediate quartzite unit called the Quartzite Range Formation, and an upper, much thinner unit called the Reno Formation. Although it is unlikely that these three subdivisions can be identified in Washington with any precision, the general lithologic distinctions can be recognized in many places and sections can be compared with the Canadian units.

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[1] If the term quartzite is restricted to sand-size quartz clastics much of the lower part of what is here called quartzite facies does not qualify, because conglomerates and grits are included. As the conglomerates and grits are composed of quartz clasts and are tightly cemented by quartz--only differing from conventional quartzite by size of grain--the definition used in this section is expanded to include them.

Stratigraphic relations

In the Metaline 30-minute quadrangle the Gypsy Quartzite lies unconformably on the Monk Formation. The boundary, however, cannot everywhere be placed precisely. As for example, where the uppermost Monk beds are limestone there is no difficulty, but where the uppermost beds are quartzite, the separation is less exact and, in places, arbitrary. On the other hand, in the West Nelson map area, Little (1960, p. 20) describes the top of the Monk as sharply separated from the Three Sisters Formation (lower Gypsy Quartzite) being the place where "fine argillaceous schists give way to thick beds of coarse green grit containing a few thin beds of fine pebble conglomerate." As the Three Sisters Formation of quartz sands, grits and conglomerates is traced northeasterly through the East Nelson map area, it interfingers with beds of argillite and limestone, the quartzite becomes argillaceous, and the assemblage becomes indistinguishable from the Horsethief Creek Formation (Monk Formation), with which it is so included by Rice (1941, p. 19).

Southwest of the Metaline quadrangle the base of the quartzite facies is a well defined unconformity. Only in the northern part of the Chewelah quadrangle does the Addy Quartzite (Gypsy) rest on the Monk Formation, which nowhere exceeds 150 metres in thickness. Elsewhere in the western part of this quadrangle it rest on Huckleberry conglomerate (Shedroof) or greenstone (Leola Volcanics); in the eastern part, east of the Jump-off-Joe fault, it rests on rocks of the Belt Supergroup.

Still further to the southwest, in the Hunters and Turtle Lake quadrangles (Becraft and Weis, 1963; Campbell, A. B., and Raup, 1964) the unconformity between Addy and older rocks is even more apparent. Figure 18 modified after Becraft and Weis (1963, p. 16) illustrates the relations along the regional strike in this area. Southwestward the Addy overlaps older and older units, progressively covering all Windermere units and continuing across the Deer Trail rocks to rest in the southwesternmost exposures on the lowermost Deer Trail unit, the Togo Formation. The geology at the base of the Cambrian quartzite is shown on a paleogeologic map (fig. 19).

The top of the Cambrian quartzite facies is an easily recognized horizon in southern British Columbia and in the Metaline and Colville quadrangles, where it is placed at the base of the limestone that named Reeves in the Salmo district and Badshot further north. Relations to the south of 48°30' are more complicated. In the eastern part of the Chewelah 30-minute quadrangle, the only possible depositional boundary between the Addy Quartzite and younger Paleozoic rocks is two and one-half kilometres southeast of Springdale, where fossiliferous middle Cambrian limestone overlies quartzites of the Addy in what appears to be accordant relations (Miller, 1969). But the contact between quartzite and limestone is separated by a 30 metre covered zone; accordingly the contact can be interpreted as either a near-bedding fault or an unconformity. Southwest of the Chewelah quadrangle, in the Hunters and Turtle Lake quadrangles (Campbell and Raup, 1964 and Becraft and Weis, 1963) Cambrian limestone rests directly on the Addy. Neither here nor in the Chewelah quadrangle is the Reeves limestone or the thick phyllite unit that overlies the Reeves present.

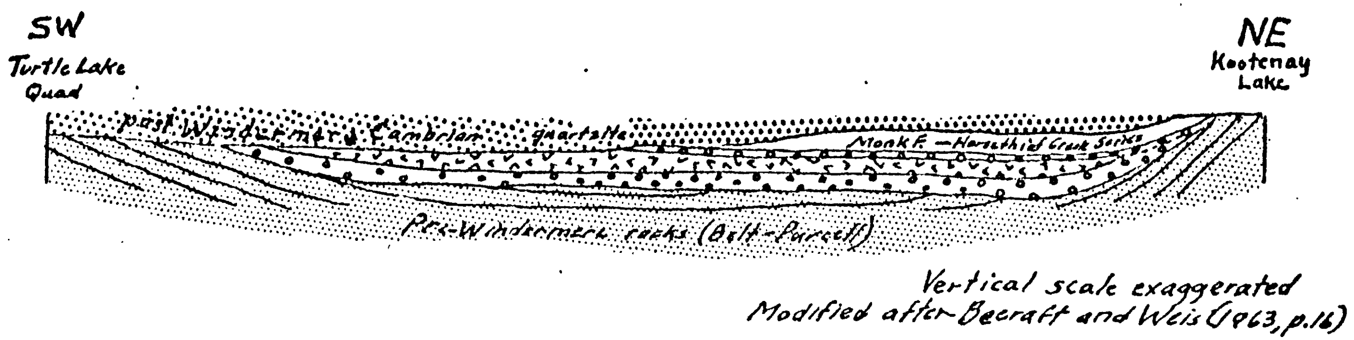


Figure 18. -- Sketch showing relations between Windermere units and between Windermere rocks and overlying and underlying rocks. The relative thickness of the thinner units is exaggerated. Section extends from the Turtle Lake quadrangle northeasterly to Kootenay Lake, British Columbia. Windermere units include Shedroof-Huckleberry Conglomerate, Irene-Leola-Huckleberry volcanics, basal Monk conglomerate, and Monk-Horsethief Creek Series.

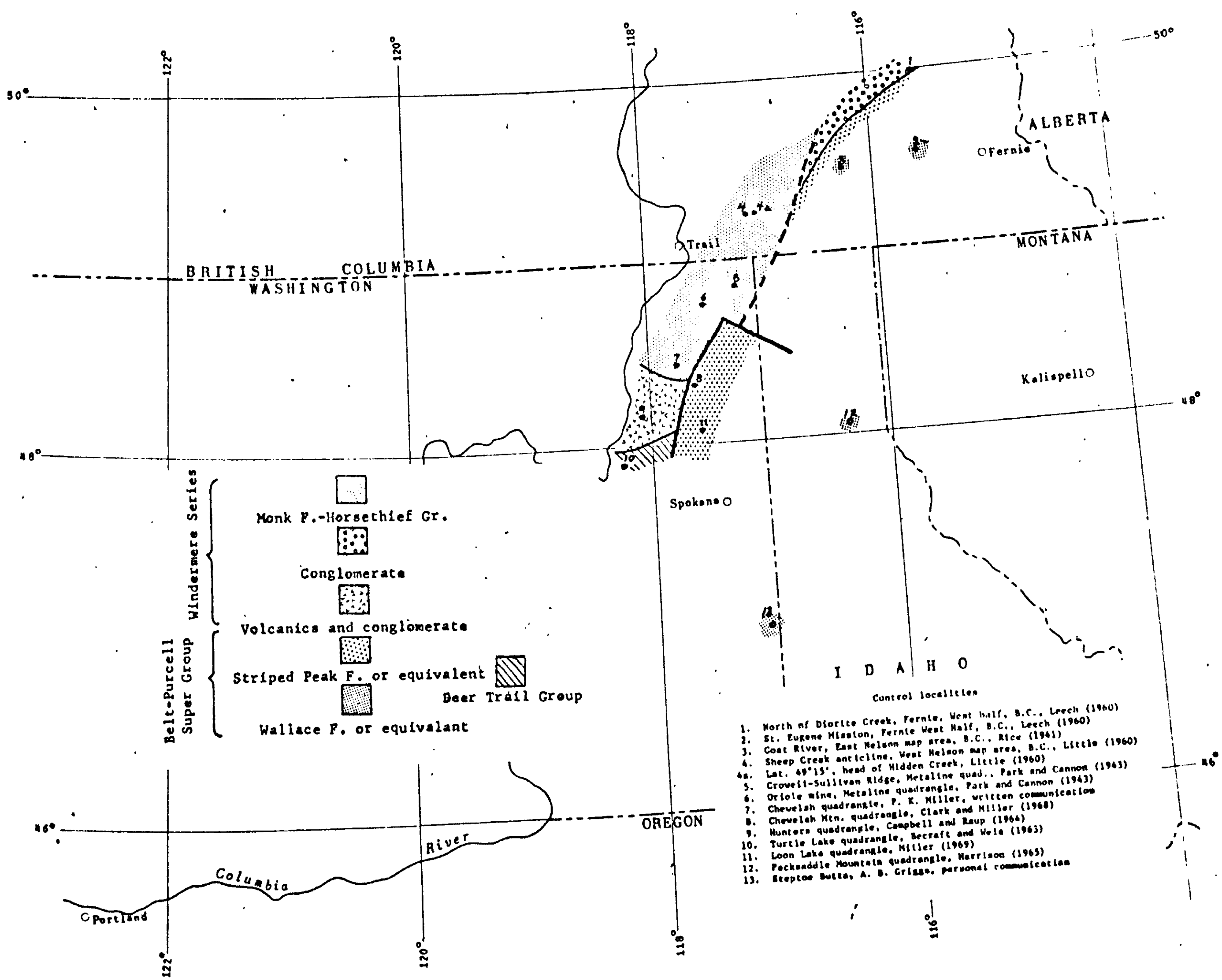


Figure 19.--Paleogeologic map of Precambrian at base of the Cambrian-Precambrian basal qiartzite facies.

Subdivisions

In the West Nelson map area, the oldest of the three formations that correlate with the Gypsy Quartzite, the Three Sisters Formation, is separated from the overlying Quartzite Range Formations at the top of the uppermost gray or green grit bed. Rocks below this bed are predominantly grits and conglomerate; those above are predominantly quartzites of sand-size grains. The Quartzite Range Formation is separated from the overlying Reno Formation (Little, 1960, p. 27) "at the transition of white quartzite to dark argillaceous quartzite and mica schist." The top of the Reno "is sharply defined in most localities against the overlying basal limestone of the Laib Formation (Maitlen Phyllite)."

Although Park and Cannon (1943, pl. 1) did not divide the Gypsy Quartzite into subunits on their geologic map of the Metaline quadrangle, the subunits used in the West Nelson map area can be recognized in the two columnar sections (fig. 3, p. 14) that were measured. In the Crowell-Sullivan section, the top of the Three Sisters Formation would be at the base of the shale that occurs 1364 metres above the top of the Monk Formation, and the top of the Quartzite Range Formation would be at the top of the "platy buff to white quartzite." Park and Cannon's placement of the top of the Gypsy Quartzite, however, does not agree with that selected for the top of the Reno Formation in Canada, which is what is accepted by the writer and used in this report. The limestone selected by Little (1960, p. 30) and others, as the upper boundary of the Reno, is the same as that described by Park and Cannon (p. 15) as a "gray-white limestone band 60 metres thick (not shown on the map)" occurring 30 metres above the top of the Gypsy Quartzite. Because this limestone, the most important stratigraphic marker in the Kootenay arc, variously known as the Reeves Limestone Member in the Salmo district and the Badshot Limestone further north, represents a change from clastic to organic sediments it logically should be placed, not within a formation, but at a formation boundary.

Until the Gypsy Quartzite in the Metaline quadrangle is remapped with these subdivisions, there is no certainty that the above correlations are accurate. Certainly one can not correlate with confidence Canadian subdivisions with parts of the Addy Quartzite (Gypsy) in the Chewelah quadrangle. Nevertheless, it is conceivable--but highly speculative--that the shale horizon described by Miller (1969, p. 3), a well-bedded, light gray, pale green and pale green and pale maroon rock, 15 metres thick, that lies about 300 metres above the base of the Addy Quartzite on Quartzite Mountain in the Chewelah Mountain quadrangle, is equivalent to the 6 metre dark green schist-argillite unit measured by Park and Cannon (1943, p. 14, fig. 3) and considered in the preceding paragraph as the probable division between units correlative to the Three Sisters and Quartzite Range Formations.

The amount of the composite section of Addy Quartzite in the Chewelah quadrangle that is time equivalent to the Gypsy section in the Metaline quadrangle is unknown. Due to the positive unconformity at the base of the quartzite and a possible unconformity at the top, there is little control on time-rock correlation, other than the assurance from fossils that both Gypsy and Addy are in part Lower Cambrian in age. Probably the basal beds of Addy Quartzite west of the Jump-off-Joe fault are essentially contemporaneous with those in the Metaline section, but this tentative conclusion cannot be directly derived. Pertinent to this conclusion is the lack of basal conglomerate or sediments derived from the underlying rocks. The conglomerates that occur in the Metaline section are intercalated through the lower half, and clasts in these, are of rock types foreign to the underlying pre-quartzite rocks, mainly quartzites that closely resemble the Gypsy Quartzite and vein quartz with minor schist and porphyritic igneous rock. Any material that might have been eroded to produce the surface on which Gypsy rests was completely removed before basal beds of the Gypsy were deposited. However, Miller and Clark (1975, p. 28) found in the Loon Lake quadrangle that basal beds of the Addy contained pebbles of the underlying Precambrian rocks. This suggests that a local positive area capable of supplying detritus existed south of the Chewelah quadrangle once subsidence began. The positive area that was the source of the Windermere rocks must have been baseleveled. The great floods of quartz sand that were to be deposited on the erosion surface arrived after the surface was fully developed and from a source other than the Belt-Purcell terrane. According to current concepts, this blanket of sand advanced eastward across the eroded edge of Belt rocks into Idaho, where it is known as Gold Creek Quartzite (Harrison, 1965) and beyond into Montana to be known as the Flathead Quartzite. It thins from 2600 metres in Washington to less than 656 metres in Idaho and Montana, and changes in age from Lower Cambrian in Washington, to Middle Cambrian in Idaho and Montana thus indicating a transgressive sea and a diachronous relation.

The above interpretation should not be construed to suggest that the deposition of the quartzite facies was accompanied by uninterrupted subsidence. Conglomerates in the lower half of the Gypsy Quartzite and well above the base indicate that this was not the case. These conglomerates have clasts composed of quartzite very similar to the purple quartzite at the base of the Addy, a feature that strongly suggests local, near-shore uplift, erosion, and redeposition offshore. Little (1960, p. 22) comes to a similar conclusion in the West Nelson map area (1960, p. 22); he describes the Three Sisters Formation as a lithology of green grit and interlayered conglomerate, and says, "the bands of conglomerate are intraformational and the pebbles are largely of grit from the underlying strata."

General lithology and regional variations

Although considerable variation in thickness as well as minor variations in grain size, color, and depositional features exists, there is very little deviation from the overall high quartz content of this widespread stratigraphic unit that once covered an appreciable part of Cordilleran North America.

The lower part of the Gypsy Quartzite in the Metaline quadrangle and the equivalent Three Sisters Formations in the West Nelson map area are described as predominantly grit, with intercalations of conglomerate and sand size quartzite. In the Metaline quadrangle the conglomerate throughout the lower part of the section occurs as beds in the grit; some conglomerate units are one hundred metres or more thick. These coarse clastics are separated from sand size sediments, which lie above, by a band of schist. East of the Pend Oreille River, the schist is 1400 metres above the Monk Formation; west of the river it is 384 metres above the Monk (Park and Cannon, 1943, p. 14, fig. 3). In the West Nelson map area, the Twin Sisters Formation, which includes all grits and conglomerates in the quartzite facies, ranges from 1853 metres near the International Boundary to 1320 metres in the northern part of the area (Little, 1960, p. 21). The schist horizon described by Park and Cannon is not present at the Boundary or along the strike to the north. However to the west in the Ymir district (McAllister, 1961, p. 6), the Three Sisters Formation is more than one-half schist. This lower part of the quartzite facies is not exposed in the Deep Creek area and Northport quadrangle. Mathews (1953, p. 17), however, reports what may be the same schist in the Sheep Creek anticline, the principal fold in the Kootenay arc. He says, "the grey grits (Three Sisters Formation) are directly overlain by several feet of green schist commonly containing large dolomite rhombs and apparently representing a metamorphosed dolomite tuff. This member, though thin, is distinctive and serves as an effective marker for the top of the Three Sisters in this locality." This green schist in the quartzite facies, interpreted as a volcanic unit, may correlate with a meta-volcanic green schist found by the writer at the top of the grit-conglomerate unit in the Metaline quadrangle at an elevation of 4280 feet on the road to the Sullivan Mountain lookout.

Because of probable faults, the distance of this meta-volcanic unit above the Monk is uncertain. Another thin volcanic unit occurs in the same area 240 metres above the base of the grit-conglomerate member of the Gypsy Quartzite. This is a highly sheared mafic lava less than 15 metres thick. The rock is altered to epidote, albite and chlorite.

Not only does the grit-conglomerate member of the quartzite facies include more and thicker schist units to the west, but it also thins in this direction. It is inferred to also thin eastward, lapping up on the positive area known as Montania, which apparently did not begin subsiding until the Middle Cambrian.

As mentioned in the preceding section the lower unit, Three Sisters Quartzite of British Columbia, changes northward from a coarse conglomeratic quartzite to the lithology of fine clastics typical of the Horsethief Creek Formation (Monk). To the south, in the Chewelah quadrangle, the lower part of the quartzite facies (Addy Quartzite) is described by Miller (1975, p. 27-28) as a unit having a basal 30 to 90 metres of well-bedded purple quartzite, a color variation that has no counterpart to the north except as pebbles and cobbles in conglomerate well above the base of the formation. In the Chewelah quadrangle the grits of the Metaline area are replaced by fine- to medium-grained quartzite; conglomerate appears only in the lower 120 metres of the formation as small pebble lenses. Still further south and west in the Turtle Lake quadrangle, Becraft and Weis (1963, p. 12) describe the Addy (Gypsy) Quartzite as consisting of interlayered quartzite and argillite units, totaling 1190 metres in thickness. The lowest, basal, quartzite unit is about 790 metres thick and is separated from the next quartzite by 210 metres of argillite. Whether this argillite is the same horizon as the shale in Miller's (19 , p.) section or the schist in Park and Cannon's (1943, p. 14) section is unknown. Conglomerate is not reported from the Turtle Lake quadrangle (Becraft and Weis, 1963, p. 12), but graded beds and cross bedding are common. Feldspar is rare.

In the Metaline quadrangle, that part of the Gypsy Quartzite that lies above the shale horizon is equivalent to the Quartzite Range and Reno Formations of the West Nelson map area and probably equals that mapped as Gypsy Quartzite in the Deep Creek area and Northport quadrangle. The extension of the upper Gypsy Quartzite north of the International Boundary is divided into two units, the lower unit, the Quartzite Range Formation, which is largely pure, white quartzite, and the upper unit, the Reno Formation, which is much thinner, consisting mainly of dark argillaceous quartzite, argillite, and mica schist.

Near the northern limit of the regional geologic map (fig. 14) on the west limb of the Sheep Creek anticline Mathews (1953, p. 18) found the Quartzite Range Formation to be on the average about 610 metres thick. This contrasts with the section to the east of the Sheep Creek anticline where McAllister (1951, p. 9) measured 1370 metres of section, which was quartzite except for a 65 metre band of sericite schist about 500 metres above the base. Little (1960, p. 26) sums up Quartzite Range lithology in the West Nelson map area as follows:

"The massive, white quartzite beds that predominate in the Quartzite Range Formation are most striking, and have been traced for nearly 240 kilometres (150 mi) northward to the main line of the Canadian Pacific Railway. These beds are 4-½ to 6 metres (15 to 20 ft) thick, rather coarse-grained, and cross beds are abundant. Near the International Boundary ripple-marks are common. Some 490 metres (1600 ft) above the base is a band of impure argillaceous quartzite and schist 60 metres (200 ft) thick; but otherwise, in the eastern band, only pure quartzites are present.

Further west, the quartzite beds are thinner and finer-grained, several argillaceous members appear. Except in the argillaceous bed, both grains and matrix consist almost entirely of quartz. In the impure beds some sericite is developed."

The Reno Formation (uppermost Gypsy) likewise varies laterally in thickness and composition. Near the International Boundary about 300 metres of beds are included; but this thickness decreases northward and at the latitude of 49°15' is only 225 metres (McAllister, 1951, p. 12). From here the Reno continues to thin and become more schistose to the north. This change is even more pronounced in a westward direction as McAllister (1951, p. 13) found that on the west limb of the Laib syncline, the Reno was only 40 metres thick and less than half of this was quartzite. To the southwest in the vicinity of the Reeves McDonald mine on the lower Salmo River, Fyles and Hewelett (1959, p. 19) measured 73 metres of largely "dark-grey micaceous interbedded with dark-grey to black phyllite." However, they excluded 18 metres of overlying beds that would have been included as Reno Formation in the sections mentioned above.

In the Metaline quadrangle the upper Gypsy Quartzite differs mostly from the lower part by being finer grained, falling in the medium- to coarse-sand range. It consists of massive white quartzite and well-bedded quartzite intercalated with shaley (phyllitic) beds that increase in abundance upwards. Crossbedding is common in both massive and bedded quartzite. The Reno equivalent consists of alternating dark-colored phyllite and quartzite in about equal amounts. It is difficult to evaluate how much of the Addy Quartzite in the Chewelah quadrangle is equivalent to the upper Gypsy Quartzite; consequently specific parts of the Addy cannot be correlated with Canadian subdivisions. The Addy in the Hunters quadrangle is similar to that in the Turtle Lake quadrangle: conglomerates are missing and argillites are intercalated with the quartzites. The argillites are thick enough units to be shown separately on the Turtle Lake map, but too thin to separate to the north in the Hunters quadrangle (Campbell and Raup, 1964). Perhaps the uppermost Addy in the western part of the Chewelah 30-minute quadrangle is immediately east of the town of Addy, where according to Miller and Clark (1975, p. 28) a section about 1200 metres thick is well exposed. Its upper limit is a fault. About 760 metres above the base of the section, trilobites occur in thin-bedded argillaceous quartzite, which is overlain by the thick bedded quartzite typical of the section.

Age

The most precise evidence of the age of the Gypsy (Addy) Quartzite are the fossils mentioned in the preceding paragraph. The quartzite near the town of Addy has three fossil localities which have been described by Okulitch (1951, p. 405) and Miller and Clark (1975, p. 28). The identification of Nevadella addyensis (Okuitch), Kutorgina ? , and other trilobites date the horizon as early Lower Cambrian in age. According to Miller the fossil horizon is about 760 metres above the base of an incomplete section of Addy Quartzite estimated to be about 1200 metres thick. The lithologic unit in which the fossils occur has not been recognized in other parts of the Chewelah quadrangle. It is a thin-bedded, fine-grained quartzite interlayered with 10 centimetre beds of black to gray-green argillite.

One fossil locality in the Metaline quadrangle, furnished fragments of trilobites, which although of "unquestionable" Cambrian age, of unknown position within the Cambrian. Park and Cannon (1943, p. 15) found these fragments on a ridge at an altitude of 6250 ft on the west slope of Gypsy Peak. On Park and Cannon's map (pl. 1) this locality is very near the top of the Gypsy Quartzite, probably in beds equivalent to the Reno Formation; it is doubtful if this fossil horizon is the same as that near Addy.

Correlation of the Gypsy Quartzite of the Metaline quadrangle with Gypsy Quartzite of the Hooknose-Baldy block of the Deep Creek area, and with the Three Sisters, Quartzite Range, and Reno Formations of the Sheep Creek anticline and associated folds in the West Nelson map area is direct because of good continuity of outcrops, structure, and sequence north across the International Boundary and west into the Leadpoint quadrangle. The correlation with the Addy Quartzite of Weaver (1920, p. 61-63) is indirect because the continuity of outcrops is interrupted by plutons and faults; nevertheless, Becraft and Weis (1963, p. 15-17) were able to correlate with considerable assurance by compaing stratigraphic sequences. The lower Cambrian fossils found in the Addy support this correlation because the limestone (Reeves) that immediately overlies the Gypsy Quartzite in the Metaline quadrangle also contains lower Cambrian fossils (Little, 1960, p. 34) indicating a lower Cambrian age for the Gypsy. Quartzites that are correlated with the Gypsy Quartzite in the Lime Creek Mountain block, and particularly in the Red Top-Grass Mountain block of the Deep Creek area as well as in the Salmo area to the north are not as directly correlated. These quartzites are unfossiliferous and associated with other unfossiliferous rock unit; consequently, correlations depend upon similarities of lithologic sequence. Although generally similar, the differ sufficiently to invite questioning the correlations.

One feature of the quartzites convinces the writer of the validity of the correlation. At the top of the sequence of interbedded quartzites and phyllites called the Reno Formation in Canada, is a zone form .3 m to more than 30 metres thick of magnetite-bearing quartzite. This zone, described on page 22, occur in all the strato-tectonic blocks of the Deep Creek area and in the quartzites of the Columbia anticline of the Northport quadrangle as well and was recognized north of the International Boundary by Daly (1912, p. 157), who observed a magnetite bed from 5 centimetres to 2.4 metres thick through a strike distance of 11 km. Daly describes this horizon as occurring in the Beehive Formation, which Park and Cannon (1943, p. 6) consider equivalent to the upper Gypsy Quartzite and lower Metaline Phyllite, and which Walker (1934) considered essentially equivalent to the Reno Formation. Fyles and Hewlett (1953, p. 22) in speaking of the Reno Formation in the Salmo district say "some of the impure quartzite in the upper part of the formations is abundantly charged with magnetitie..." This magnetite horizon has not been recorder in the Chewelah quadrangle or further to the southwest.

The Cambrian phyllites--Maitlen and Laib Formations

The stratigraphic unit on the regional geologic map (fig. 14) designated Cambrian phyllite and limestone is dominantly gray-green shale, though it locally includes thin beds of limestone. The gray-green shale facies dominates both the Maitlen Formation in Washington and the Laib Formation in British Columbia. The Laib Formation extends from the International Boundary 42 kilometres to the northeast in the West Nelson map area and according to Little (1960, p. 35) probably reappears to the northeast as the Badshot Formation and lower part of the Lardeau Series. The Maitlen Formation is present in both Metaline and Colville quadrangles, but apparently is absent in the Chewelah quadrangle, and most probably, in the Turtle Lake quadrangle, although here the uppermost argillite beds in the Addy Quartzite could conceivably be time equivalent to part of the Maitlen.

Stratigraphic relations

The lower boundary of the Maitlen Phyllite (Laib Formation) is the base of the Reeves Limestone Member which was accepted as the upper limit of the quartzite facies of the Cambrian. The upper boundary of the Maitlen Phyllite (Laib Formation) is not so precisely defined, either in British Columbia or Washington. In most places, it is placed within a gradational zone between phyllite and argillaceous limestone of the overlying Metaline Formation (Nelway Formation). The boundary is placed where limestone becomes predominant over phyllite. The transitional zone is everywhere less than 100 metres thick, commonly less than 30 metres.

Subdivisions

In the Metaline quadrangle, east of the Leadpoint fault in the Deep Creek area, and in the Sheep Creek anticline of the west Nelson map area the subdivision of the Maitlen and Laib Formations is simple--but lopsided. The formations are divided into two members, a relatively thin basal unit formalized as the Reeves Limestone Member, and an upper, thick unit informally named the phyllite member. In the Salmo district west of the Sheep Creek anticline, just north of the international boundary Fyles and Hewlett (1959, p. 23-28) have subdivided the Laib (Maitlen) into four members, which are: Truman (phyllite and minor limestone), Reeves (limestone), Emerald (black argillite), and Upper Laib (phyllite). In this report the Truman Member is included under the Cambrian quartzite unit on figure 14, being considered what would be uppermost Reno.

To the south of the Salmo district, in the Red Top-Grass Mountain block of the Deep Creek area and southwestward down the broken core of the Columbia anticline, into and through the Northport quadrangle, the rocks that represent the Maitlen Phyllite are a series of three limestone units and three phyllite (schist) units. The lowest limestone, which rests on quartzite correlated with the Gypsy Quartzite, is considered to be the Reeves Limestone Member. The other two, thin limestone units are unnamed, as are the intervening phyllites. The upper limit of the Maitlen Phyllite is the top of the uppermost phyllite unit and the base of the thick limestone-dolomite sequence believed to be correlative with the Metaline Formation.

General lithology and regional variation

Reeves Limestone Member

Except where dolomitized near the lead-zinc deposits of the Salmo district, the Reeves Limestone Member of the Maitlen Phyllite (Laib Formation) is just what the name signifies--a limestone. It has no specific characteristics of color, fossils, bedding, grain size, or thickness that identifies it from area to area. Its recognition depends largely upon its place in the stratigraphic column, that of a divider between the quartzite and gray-green shale facies.

A few miles north of the International Boundary on the Sheep Creek anticline the Reeves Limestone Member was observed by Fyles and Hewlett (1959, p. 25-26) to be a gray fine-grained massive limestone commonly made up of dark and light gray bands, a few feet to tens of feet thick. Near its base it has a mottled appearance. It ranges from 120 to 140 metres in thickness. Sixteen kilometres further north and on strike it is "fine to medium grained, white or grey...and in places contains rounded dark grey coarsely crystalline spots one quarter to three quarters of an inch in diameter resembling remnants of fossils." Here the limestone is poorly banded and ranges from 30 to 150 metres feet thick.

Southwest from the Salmo mining district, in the Red Top-Grass Mountain block of the Deep Creek area, and further southwest in the Northport quadrangle the correlation of the Reeves is questionable. There is no certainty that the limestone called Reeves is the same horizon as the type Reeves. Although both are the first significant limestone unit above the quartzite, it is possible that the "Red Top" Reeves is a southwestward continuation and thickening of the thin white limestone in the Truman Member. If this is true, the intermediate limestone in the Red Top Mountain section could be the Reeves correlative. This basal limestone that rests on quartzite, regardless of how precisely correlated, can be recognized in the core of the Columbia anticline as far south as Bossburg. It is a thin unit, ranging from 15 to 60 metres in thickness, a coarsely crystalline, white to gray, medium bedded rock that looks very different from the other two limestone units in the phyllite of the Maitlen.

Returning to the eastern outcrops of Cambrian rocks, to the Sheep Creek anticline and the homoclinal structure that lies east of the anticline, we read that Little (1960, p. 33-34) found the Reeves limestone in the West Nelson Map area to range from 60 metres thick at the International Boundary to 20 metres 23 kilometres to the north. This, however, is not a uniform change, because in most places the thickness is nearer the lower figure, and in a few places the unit is absent. South of the International Boundary, Park and Cannon (1943, p. 15) give the limestone unit that lies on the quartzite a thickness of about 60 metres, but give no figure for that west of the Pend Oreille River. Outcrops in the western part of the Hooknose-Baldy block have a thickness of 120 metres indicating that the unit thickens westward in both the Canadian and United States part of the Kootenay arc.

The Reeves Limestone Member is a persistent unit in the Metaline and Colville quadrangle; it can be identified as far south as the town of Colville. It has not, however, been recognized in the Chewelah and Hunters quadrangles.

In the Salmo lead-zinc area the Reeves Limestone Member of the Laib Formation. The Emerald Member has not been recognized in any other part of the central or southern Kootenay arc, although the writer erroneously correlated black argillites in the Red Top-Grass Mountain block (Yates, 1964) with it.

The Emerald Member is defined as the black argillite conformably overlying the Reeves Limestone (Fyles and Hewlett, 1959, p. 26). Its lower contact is sharp; its upper contact is gradational from black argillite to green and brown phyllite of the Upper Laib Member. It occurs in the overturned limbs of folds. The following description is from Fyles and Hewlett (1959, p. 27) "Rocks termed argillite are blocky, have a poor cleavage and grade into phyllite on the one hand and into hornfels on the other. Those termed phyllite...grade into black schist. Most of the phyllites are crenulated but some are slaty." Thickness ranges from 23 to 150 metres.

Phyllite Member (Upper Laib)

The phyllite member of the Maitlen Phyllite is present in the Metaline and Colville quadrangles and present as the Upper Laib in British Columbia, in the West Nelson Map area. It has not been identified south of the Colville quadrangle. It is composed of sericite, quartz, and minor amounts of chlorite. Locally, and rarely, it is calcareous, although it does contain a few widely spaced, thin limestone layers and lenses. It grades upward into the argillaceous limestone of the Metaline Formation (Nelway Formation). Almost everywhere the phyllite is distorted and sheared, a characteristic of this very incompetent rock.

In the Metaline quadrangle, Park and Cannon (1943, p. 16) describe the Maitlen Phyllite as being "gray greenish, fine-grained, and conspicuously banded." This dominant lithology is interlayered near the base by quartzite beds and near the top by limestone beds. The lower part of the phyllite containing quartzite beds is part of the section considered in this report as part of the Gypsy Quartzite. Dings and Whitebread (1965, p. 8) found that in places, the upper part was inclined to be black and free of bedding planes.

The phyllite in the phyllite-limestone sequence that represents the post-Reeves Maitlen in the Columbia anticline ranges from a phyllite to a fine-grained schist. At the northeast end of the fold on the Red Top Mountain, the section consists of three limestone units and two phyllite schist units: the lower limestone is correlated with the Reeves Member and the upper limestone with the Metaline Formation; the phyllite units compose the phyllite part of the Maitlen. The lower phyllite is about 30 m thick and the upper about 60 m thick.

Three kilometres down the river from Northport, the Maitlen Phyllite consists of three phyllite units interstratified with limestone units. The phyllite units are 60, 30, and 170 metres thick respectively from the lower to the upper. Thirteen kilometres further down river, to the southwest, an unfaulted section is absent, and the outcrops neither confirm nor reject the possibility that a reconstructed would be section similar to that near Northport. Some partial sections are similar, others are not; differences may be from localized limestone tongues or by undiscovered near bedding faults.

Age

The Cambrian age of the Maitlen Phyllite (Laib Formation) is well established; Lower Cambrian archaeocyathids have been found in the Reeves Limestone and Middle Cambrian trilobites occur in the overlying lower unit of the Metaline (Nelway) Formation. Assuming relatively constant depositional rates for the phyllite of the Maitlen and argillaceous limestone of the Lower Metaline, all but the upper part of the Maitlen was deposited in the lower Cambrian.

In the western part of the Kootenay arc fossils have been found in the Reeves Limestone only near the town of Colville. Near Colville, archaeocyathids were first found by Bennett (1937, p. 316) and were later collected by Okulitch (1951, p. 405-407) and reported as of Lower Cambrian age. Eastern Kootenay arc localities are in the homoclinal section east of the Sheep Creek anticline, "from a point 300 m south of the International Boundary" to a point 11 km to the north (Little, 1950, p. 17; 1960, p. 34). Okulitch states in comparing faunas (Little, 1960, p. 34), "The similarity to the Donald fauna is rather strong, but there are elements previously known in Nevada, Mexico, and Australia. Your fauna is in all probability, of the same age as the Peyto limestone of the Rocky Mountains." To the west of the localities described by Little (1960) in the Sheep Creek anticline, Fyles and Hewlett describe what are possibly archaeocyathids, "in the northern part of the Sheep Creek anticline, west of the western anticline" where it contains rounded dark-grey coarsely crystalline spots one-quarter to three-quarter of an inch in diameter resembling remnants of fossils."

Correlation

The correlation of the Maitlen Phyllite (Laib Formation) with its equivalents is largely made by matching its basal members, the Reeves Limestone Member, with a stratigraphically equivalent limestone. The Reeves is recognizable throughout the West Nelson map area of British Columbia and the Metaline and Maitlen quadrangles of Washington. South of these quadrangles, in the southern part of the Kootenay arc, it has not been recognized. Here fossiliferous Metaline Limestone rests directly on the quartzite facies of the Cambrian, the rocks, which to the north, directly underlie the Reeves.

North of the West Nelson map area, in the Lardeau map area, the Reeves is recognized as a limestone named the Badshot (Limestone) (Walker, Bancroft, and Gunning, 1929, p. 10). The Badshot, containing Lower Cambrian Archaeocyathids has been recognized as far as 210 km north of the boundary, in the Rogers Pass map area (Wheeler, 1963, p. 6-7).

Carbonate Rocks

The Metaline Formation is the host of essentially all the lead-zinc deposits in northeastern Washington. This unit extends across the International Boundary, where it is named the Nelway Formation, but it has not been productive in British Columbia. The near by Canadian deposits are in the Reeves Limestone Member of the Laib Formation. The Nelway Formation has not been identified more than 34 km north of the boundary. The Metaline Formation, as the Old Dominion Limestone of Weaver (1920), however, extends as far south as the Spokane River, where it disappears under the basalt of the Columbia Plateau.

In Charles E. Weaver's map of Stevens County (1920, pl. 1) he shows five separate formations that are now known to be tectonically isolated outcrops of the Metaline Formation. These are: Northport Limestone, Clugstone Limestone, Old Dominion Limestone, Republican Creek Limestone, and Red Top Limestone. The names were given at a time when the stratigraphy of the county was without fossil control and known in only reconnaisance manner, accordingly it was recognized by Weaver (19__, p. 44-46) that the names had only temporary validity. Regardless of the fact that they supercede the Metaline Formation, I strongly recommend that all the Middle Cambrian carbonate sequence in northeastern Washington be known by the name, Metaline Formation, which is well controlled, stratigraphically and faunally at its type locality in the Metaline quadrangle near Metaline Falls.

Representatives of all five of Weaver's formations are in the Colville 30 minute quadrangle: in the north half, the Deep Creek area and Northport quadrangle all units but the Old Dominion Limestone are present; in the south half, the Old Dominion Limestone as indicated by Weaver includes both the Metaline Formation and the Reeves Limestone Member. Yates (1964, 1971) has redefined Weaver's units in the north half of the quadrangle and W. A. G. Bennett's map of the south one half (in Mills, 1966) abandoned the olds names and subsituted informal lithologic names for divisions of the Metaline Formation and set off the Reeves Limestone as a "lower *Archaeocyathae* limestone reef member."

In the area of the Chewelah 30 minute quadrangle, Weaver indentified two limestone units, the Old Dominion Limestone and "undifferentiated limestone." The Old Dominion is shown on the map only near Addy in (Campbell and Raup, 1964) the northwestern corner of the quadrangle. In adjoining Hunters and Turtle Lake (Becraft and Weis, 1963) quadrangles the compromise name "Old Dominion Limestone of Weaver (1920)" was applied to a formation that clearly is the lithologic equivalent of the Metaline Formation, although its lower part is probably the time equivalent of the upper part of the type section in the Metaline quadrangle.

Stratigraphic Relations

The transition from phyllite to the carbonate facies of the Cambrian is a change from non-limy phyllite to limy-phyllite and thence to interbedded phyllite and limestone. This is the relation in the West Nelson map area and the Metaline and Colville quadrangles, but to the south the phyllite is absent, and the carbonate rocks rest directly on the quartzite of the Addy (Gypsy). In all these areas the relations are regarded as concordant; therefore the lack of phyllite represents either non-deposition, erosion, facies change, or a near-bedding thrust fault. Known facts are inadequate to select a positive interpretation, but the writer favors the postulated erosion because fossils (A. B. Campbell, 1964) collected near the base of the Metaline Formation equivalent, the Old Dominion Limestone of Weaver, are upper Cambrian, thus indicating that an upper Cambrian limestone rests upon a lower Cambrian quartzite.

In the West Nelson map area and Metaline quadrangle the uppermost carbonate beds of the Nelway and Metaline Formations are overlain in sharp contact by the black shale facies of the Ordovician; in the Deep Creek area the change is both sharp and transitional; in the Turtle Lake quadrangle, the carbonate rocks are separated from the Ordovician black slates by a chert unit.

In the Metaline quadrangle an undeformed contact between the carbonate sequence and the overlying black slates of the Ordovician is rarely seen, but considering the deformation the region has undergone, movement is expectable between such contrasting rocks. Dings and Whitebread (1965, p. 21) say, "At most places the Ledbetter rocks appear to rest conformably on the Metaline strata, although a minor erosion surface at the top of the Metaline could easily go undetected along the undulatory and deformed contact where bedding planes are not recognizable in the carbonate rock." It is also possible that an unconformity or disconformity was produced during Ledbetter time, and that this separates in some places middle Ordovician Black slate from Metaline carbonate rocks and in other places separates lower Ordovician black slates from Middle Ordovician slates. This relation is suggested by the anomalous Lower Ordovician age of graptolites that were collected from the "west bank of the Pend Oreille River north of the Pend Oreille mine" (Park and Cannon, 1943, p. 20-21) in a stratigraphic position elsewhere determined to be middle Ordovician. Although this is not a precise location and tells nothing about the distance above the top of the Metaline Formation, the locality is certainly no closer to the contact than the two middle Ordovician localities listed first in the table on page 21 (Park and Cannon, 1943) and the locality collected by Michael Churkin (fig. 19A, locality M-2-6) south of Beatty Lake and less than 100 feet above the contact. All these localities are as close or closer to the contact than the Lower Ordovician locality. A near-bedding thrust fault could also explain this relation.

General lithology, subdivision, and regional variation

The Metaline (Nelway) Formation consists of limestone and dolomite, and essentially nothing else, except in zones gradational to the underlying phyllite, or overlying black slate, or where it merges with other facies to the north and south. Wherever subdivision of the formation has been attempted, the units and members are monolithologic and readily separable. Becraft and Weis (1963, p. 12-13) describe under the name old Dominion Limestone the Metaline Formation in the Turtle Lake quadrangle without mentioning the presence of dolomite, except indirectly, as tremolite bearing marble in the contact aureoles of the Loon Lake granite. Campbell and Raup (1964) in the map explanation for the Hunters quadrangle term it a "heterogenous sequence of thin-bedded light to dark-gray limestone, slate, dolomite, and mixtures of these types," but made no subdivisions. Although Miller and Clark (1975, p. 29) recognized a basal argillaceous limestone and an overlying dolomite with thin shale interbeds, they did not distinguish these on this geologic map.

In this report the section in the Metaline quadrangle and its extension in the Salmo district is considered the standard or "reference" section and all other sections are compared with it. Variations among the several sections of the carbonate rocks have been very helpful in interpreting the structure of the region. It is inconsistencies in juxtaposed sections that forces the interpretation of prefold decollement faulting.

The basic subdivisions of the Metaline Formation were outlined by Park and Cannon (1943, p. 17-18) in the Metaline quadrangle although they never tested their subdivisions by mapping. They divided the section into four parts; the lower two parts have since passed the test of mapping by Dings and Whitebread (1965, p. 1), who combined the upper two because they found that the relative abundance of chert nodules--the basis for Park and Cannon's subdivision--was an unreliable distinction. The three-fold division that results from the combining of Park and Cannon's two upper units is a lower limestone, intermediate dolomite, and upper limestone. This division and the descriptions of the section in the Metaline quadrangle that follows are largely those of Dings and Whitebread (1965, p. 9-22). Major divisions are: a lower, thin-bedded limestone-shaley dolomitic limestone sequence, an intermediate light gray bedded dolomite, and an upper gray, massive limestone.

The lower, dolomitic limestone is the only unit in the Metaline section that is transitional to another lithology. Contacts between the middle dolomite and limestones of the upper and lower units are sharp, as is the contact of the upper unit with the Ledbetter Slate. The interbedding of shale (phyllite) and limestone occurs not only in the lower part of the lower unit but also near the top, where shale may total as much as 10 percent. Although most of the 300 metres of the lower unit is composed of a dark gray thin-bedded limestone and limy shale interbeds there is a subunit of wavy-bedded limestone in which reddish-brown shaly limestone separates elongate eyes of gray limestone. This wavy-bedded limestone is the fossiliferous unit and is easily recognized in all sections except that of the Columbia anticline.

Dings and Whitebread (1965, p. 13) describe the middle, bedded dolomite unit as, "largely composed of a creamy-white to light gray, fine- to medium-grained dolomite that shows no recognizable stratigraphic variations over wide areas." Lithologic varieties include a black and white banded dolomite and a sublithographic banded dolomite. Chemically the rock approaches very closely the composition of the mineral dolomite. Grains of clastic quartz are rare, but vug fillings and irregular masses of introduced quartz are common. Bedding is not everywhere obvious. In the Metaline district only a few ore bodies occur in the middle dolomite, whereas in the Northport district, it is the most productive stratigraphic unit.

The upper unit of the Metaline Formation, the gray limestone unit, although only a minor producer of lead and zinc in the Northport district, is by far the most productive stratigraphic unit in the Metaline district. This upper unit, from 400 to 460 metres thick, is basically a pure limestone, probably averaging 99 percent calcium carbonate; but in many places, particularly in mine workings, this purity is obscured by intense dolomitization, and, in and near ore bodies by silicification. In addition to the silica introduced during mineralization, the uppermost 60 metres of the unit locally contains nodular quartz, probably recrystallized chert. Dings and Whitebread (1965, p. 18) say "rock typical of the gray limestone unit is a light to medium-gray irregular mottled very fine-grained soft massive limestone." Rarely can individual beds be recognized, but within the upper 8 metres of the formation, bedding is indicated by shaly partings and thin shale beds, as indication of the abrupt change from deposition of carbonate rocks to deposition of black slates of Ordovician age that overlay the upper unit.

Figure 19A.--Age and location of graptolite collections from Stevens and Pend Oreille Counties, Washington. X=present (abundance not given), A=abundant, C=common, R=rare

From the upper limestone unit of Dings and Whitebread (1965) McConnel and Anderson (1968, p. 1467-1468) have split off a subunit they named the Josephine unit after the Josephine ore horizon, which it includes. This subunit is the uppermost part of the Metaline Formation and is "an irregularly brecciated stratum" ranging in thickness from a few feet to more than 60 metres (200 ft). McConnel and Anderson describe the Josephine Unit as "a breccia of gray limestone, coarse light gray dolomite, and zebra rock... embedded in a matrix of black dolomite or black jasperoid or some mixture of these two." They relate the dolomite, jasper, breccia, and sulfides to the sedimentary and diagenetic processes of a reef environment. In contrast, Dings and Whitebread believe all these features to be products of hydrothermal processes (1965).

In the Metaline quadrangle, Dings and Whitebread (1965, p. 10) assign a total thickness of 1675 metres to the Metaline Formation, which is in close agreement with the 1370 to 1525 metres estimated by Fyles and Hewlett (1959, p. 30) for the Nelway Formation north of the boundary. The Nelway resembles the Metaline in other respects, having the same three-fold division, a lower limestone, middle dolomite, and upper limestone. The boundaries between the middle dolomite and the limestones are not everywhere depositional planes but tend to cross the bedding.

Near the northeast corner of the regional map (fig. 14) is a small triangular area shown as "lower limestone" (of the Cambrian). These rocks are shown on Little's 1960 map of the West Nelson map area as Nelway Formation after assignment by McAllister (1951, p. 17-18) who originally mapped the rocks and correlated them with the "lower part of the formation." McAllister describes this northernmost exposure of Nelway Formation as occuring in the trough of a syncline on the east flank of the Sheep Creek anticline. The rocks are, "a series of black argillites, phyllites and slates and thin beds of black limestone (that) represents only the lower part of the formation as exposed in the Salmo map area." He goes on to say that the striking character of these rocks is their black color, comparative lack of schistosity, fine grain, and high lime content. The rocks are more calcareous near the top of the section, which is about 300 metres thick and include interbedded black limestone. This section is much more argillacous than the lower limestone unit in the Metaline quadrangle, which it little resembles. An alternate correlation would be with the upper Laib, which could then be interpreted as becoming blacker and more limy toward the north.

In the Loon Lake quadrangle, south of the area covered by the regional map (fig. 14), Miller (1969) found carbonate rocks he correlated with the Metaline Formation. Two units are present: the lower unit is similar to the lower unit in the Metaline quadrangle except that it has numerous beds of black shale, and the overlying unit is similar to the middle dolomite unit, except that it has a few thin shale beds. From 600 to 750 metres of beds may be present in this section, whose upper part is faulted out. Fossils in the lower zone are part of the same faunal assemblage as occurs near Metaline Falls. Southwest of the Loon Lake quadrangle, in the Inchellium, Hunters, and Turtle Lake quadrangles, carbonate rocks overlie the Addy (Gypsy) Quartzite. The carbonate rocks in this area have never been subdivided by mapping or described in enough detail to compare with the standard section in the Metaline quadrangle. However, the same rock types, dark fine-grained shaley limestone, bedded dolomite, and limestone occur and possibly in the same sequence.

Westward from the Metaline quadrangle, across the strike of the rocks, in the Deep Creek area and Northport quadrangle the Metaline Formation occurs in three distinct stratigraphic sequences, only one of which closely resembles the standard section. This section is in the Lime Creek Mountain block and is separated from the other two sections by thrust and high angle normal faults.

All three sections of the Metaline Formation agree in some respects and disagree in others. The section of the Lime Creek Mountain block agrees with the section in the Metaline quadrangle in having the same three lithologic divisions and in having very similar relations to the overlying Ledbetter Slate, but disagrees having an estimated thickness of 2600 metres which contrasts with the 1675 metres thickness of the Metaline section. The lower half of the section in the Hooknose-Baldy block is similar to the Metaline section; the upper half is very different. The section in the Red Top-Grass Mountain block (the Columbia anticline) and that in the Northport quadrangle, is similar in having the three-fold sequence of limestone, dolomite, and limestone but these units differ from comparable units in the Metaline section.

The section in the Hooknose-Baldy block, hereafter referred to as the Leadpoint section, is the most anomalous of all, partly because of its location between the two areas that have Metaline sections. The geometry of its position negates interpreting it as an _in situ_ gradational variation of the Metaline section; horizontal tectonic transport is a postulate essential to an explanation of its location--a concept developed in a following section of this report. The tectonic transport theory explains the Leadpoint section as being deposited east of the Metaline section and thrust westward at some time before the Mesozoic folding.

The lower dolomitic limestone unit of the Metaline section appears in the Leadpoint section with the same basic lithology and the same fossil horizon of Middle middle Cambrian trilobites. The Leadpoint lower unit differs most from the Metaline lower unit by the presence of two or more subunits of fine-grained black dolomite, a feature that characterizes the lower unit and the Leadpoint "facies" as far south as Old Douglas Mountain, a few miles northeast of Colville.

This lower unit is overlain by the middle dolomite unit, which also is very similar to the middle unit of the Metaline section. After 300 metres of this dolomite was deposited, conditions changed to permit the accumulation of a dolomite breccia (see p. 29-32). The breccia continues strong to the northeast, being terminated only by faulting and erosion; to the southeast towards Colville it looses its identity, becoming increasingly less pronounced as a breccia. It is academic whether the breccia belongs to the highest unit or the underlying dolomite. I am inclined to consider it equivalent to the upper part of the middle dolomite.

The upper unit in the Leadpoint section contrasts strongly with that in the Metaline section. Instead of a limestone, it is a medley of black dolomite, limy shale, intraformational dolomite breccia, and mottled limestone, a rock not unlike that of the lower unit. Fossils in this unit suggest an upper Cambrian age or possibly lower Ordovician; the introduction of shale and shaley limestone suggest gradation to the fine clastics of the Ordovician, instead of the abrupt change in the Metaline section.

The Columbia anticline section of the Metaline Formation shows considerable variation along the axis of the fold, being absent in British Columbia and only partly present in the Red Top-Grass Mountain block, that part of the fold in the Deep Creek area. The absence of the Metaline in the Canadian part of the Columbia anticline, may be entirely the result of faulting, because all contacts of Laib rocks with younger rocks are fault contacts; nevertheless, it is conceivable that the carbonate facies in the western part of the arc changes northeastward to a shale facies.

In the Northport quadrangle on the Columbia anticline, the Metaline Formation consists of three units: (1) a lower unit of medium to coarse-grained sugary, white to light gray limestone, (2) a middle unit of medium to coarse-grained white and gray medium- to thick-bedded dolomite, and (3) an upper unit of medium- to fine-grained, light gray to white, medium-bedded limestone. This is a limestone-dolomite-limestone succession typical of the Metaline Formation, but it lacks the characteristics of color, grain-size, bedding, etc. that characterize the rocks in the Metaline section. The dark dolomitic shaley limestone characterizing the lower unit is here a pure, locally sandy light-colored limestone. The middle and upper units have a much greater similarity. All contacts with overlying pelites, except those on Red Top Mountain are faults.

Age

The Cambrian age of the Metaline Formation is well established. Trilobites diagnostic of this age have been found in the Metaline quadrangle (Resser, 1934, p. 1-10; Park and Cannon, 1943, p. 19; McLaughlin and Enbysk, 1950, p. 466-471), in the Deep Creek area (see appendix), in the Loon Lake quadrangle (Miller, 1969, p. 3) and in the Hunters quadrangle (Campbell, 1964) in the Old Dominion Limestone of Weaver. Fossils collected from the lower unit of the Leadpoint section are middle Middle Cambrian, those from the middle dolomite member are unspecified Middle Cambrian, and those from the upper member are probably Upper Cambrian. Middle Cambrian fossils collected from the type locality at Metaline Falls are from near the base of the lower member; no fossils have been found in the middle or upper members. Because Lower and Middle Ordovician graptolitic black slates rest on the upper member, the Metaline type section probably includes both middle and upper Cambrian beds. The age jump from Middle Cambrian trilobites to Lower and Middle Ordovician graptolites through almost 5000 feet of section is not an indication of a hiatus at the formational boundary.

Black shale facies of the Paleozoic

The category, black shale facies of the Paleozoic, is a broad grouping of loosely related rocks that range in age from lower Ordovician to probable Mississippian. Although the grouping includes minor amounts of limestone, quartzite and conglomerate, the dominant lithology is dark gray to black fine-grained clastic sediments that are now slates, argillites, and phyllites. Some other rock units on figure 14 might qualify as members of the black shale facies on the basis of grain-size and color, but are disqualified by the presence of interlayered volcanic material, which is the basic criterion for defining the eugeosynclinal facies. Rock units that are grouped under the black shale facies are: Ledbetter Slate, Argillite A, and Argillite B, a potpourri of odds and ends and the widely distributed Active Formation of British Columbia.

Fossils collected from rocks belonging to the black shale facies range in age from lower Ordovician to middle Devonian, but the facies probably extends into the Mississippian. The Pennsylvanian and Permian rocks immediately to the west of the Columbia anticline, typical eugeosynclinal assemblages, place a tentative late Paleozoic ceiling on the black shale facies. This ceiling is valid only if both black shale and eugeosynclinal facies are autochtonous or else have been tectonically transported without change in their relative positions.

Although gross similarity between the various units exists, no widespread abnormalities of lithology have been recognized that can be used to unequivocally identify regional changes in the depositional history that might permit correlating with confidence the tectonically isolated outcrop areas. The nearest approach to such abnormalities are pebble conglomerates found at several places and possibly the siliceous argillite common to the Russian Creek Basin in the Metaline quadrangle and to the Grass Mountain sequence in the Deep Creek area. There is no assurance, however, that these features are nonrepetitive and represent distinct events. Fossils, particularly useful in identifying the Lower Paleozoic units, are frustratingly absent.

Ledbetter Slate and Argillite A

The basal unit of the black shale facies is the Ledbetter Slate, first described in the Metaline quadrangle by Park and Cannon (1943, p. 19-22), who included in the formation, all black slates and argillites that occur above the Metaline Formation. Northwest of Ledbetter Lake, Park and Cannon's map (1943, pl. 1) shows a large area indicated as Ledbetter Slate, within which is a hill capped by limestone containing Devonian fossils. Exclusion of this limestone from the Ledbetter and consequent inferred restriction of the Ledbetter to rocks deposited during the Ordovician, made it necessary for Dings and Whitebread (1965, p. 23-27) on remapping to exclude from the Ledbetter the black slates that overlay the Devonian limestone.

This restriction of the Ledbetter to the Ordovician part of the facies, makes the physical upper boundary of the formation very difficult to establish, mainly because the upper boundary depends upon the presence of Silurian or Devonian fossils, which are exceedingly rare, consequently the stratigraphic unit listed in the map explanation of figure 14 and informally named, Argillite A of Devonian and Silurian age, was identified in only two places, northwest of Ledbetter Lake in the Metaline quadrangle and provisionally in the Black Rock canyon in the Deep Creek area.

The lower contact of the Ledbetter Slate, however, can be fixed with precision. In the Metaline quadrangle it is in sharp contact with the underlying limestones of the Metaline Formation. Relations in other areas have been discussed in considerable detail in the preceeding section.

The Ledbetter Slate west of the Pend Oreille River in the Metaline quadrangle occurs in close association with rocks that contain both Silurian and Devonian fossils. Near Beatty Lake, graptolites of Middle Ordovician age occur in slates 30 metres above the top of the Metaline Formation; 670 metres stratigraphically above these fossils, in bleached and weathered silty shale, the graptolites are Silurian (Dings and Whitebread, 1965, p. 29-30). Exposures between the two fossil horizons are poor, but float indicates the rock to be dominantly black slate with some light colored silty porous beds in the upper half. A short distance above the Silurian beds, is a bed of black chert and not far above this are lenses of bioclastic limestone, rich in fossil trash. Farther north and to the south are other richly fossiliferous limestone pods and lenses that have all the characteristics of a reef of biohermal origin. Where determinable, the fossils are Middle Devonian in age.

These "bioherms" are with overlain and underlain by black slate. Within the black slates above the limestone are thin beds and lenses of pebble conglomerate composed largely of chert and quartzite clasts in a siliceous matrix. The paradoxical occurrence of conglomerate and limestone in the black shale environment indicates two disparate provenances. My interpretation is that black muds were being deposited continuously from the Ordovician through the Devonian on the continental rise and that the muds were transported by longshore currents from an unknow provenance, probably to the south. The coral-bearing limestones, which clearly accumulated in shallow waters, were formed perhaps at the lip of the rise on a shelf flanking a positive area that was being reborn in the panhandle of Idaho. Continuous uplift of this positive area and contemporaneous sinking of the rise as sediments accumulated resulted in unstable conditions along the margin, causing blocks of limestone to slide off into the muds of the rise. As the uplift of the Idaho high continued into the Late Devonian, a different facies of Silurian and Ordovician rocks from that deposited on the rise were exposed to erosion to become the source of the clasts of chert and quartzite. Gravels derived from these rocks were swept across the platform on which the limestones had accumulated into the deeper water, mud environment of the rise. Turbidity currents may have helped move the gravels down the slope.

In the eastern part of the Metaline quadrangle, a band of Ledbetter Slate extends southwestward from the International Boundary down Slate Creek to the Pend Oreille River and thence down the river to the Pend Oreille mine. This slate band contains lenses, beds, and pods of fine-grained, black to light gray quartzite. The quartzite is associated with Ordovician graptolites and occurs in the lower part of the black shale facies. Rocks younger than Ordovician have been removed by erosion.

To the west in the Deep Creek area, Ledbetter Slate has been identified by graptolites in the Hooknose-Baldy and Lime Creek Mountain blocks. The slate in the Hooknose-Baldy block on Gladstone Mountain--which belongs to the allochthonous Leadpoint section--tends to be more silty than that in the Metaline district, whereas that in the Black Rock Canyon area of the Lime Creek Mountain block appears to be thinner than that in either the Metaline quadrangle or in the Hooknose-Baldy block. The rocks that overlie the Black Rock Canyon graptolitic slates closely resemble those mapped as Devonian-Silurian in the Metaline quadrangle.

Ordovician graptolites were found in only one place in the Northport quadrangle, in the band of black slate along the north edge of the Spirit pluton, but to the south they are preserved in slates on the northwest flank of the Gillete Mountain anticline that extends southwestward towards Colville. This highly discontinuous belt of black slate extends as far southwest as the Inchelium and Hunters quadrangles where black slates have graptolites of Lower and Middle Ordovician age.

The Ordovician black shale facies in the Hunters quadrangle (Campbell and Raup, 1964), is represented by limestone and slate of Middle Ordovician age. According to the map explanation the unit is "chiefly fissile non-calcareous slate; most rocks medium to dark gray; some beds of blue-gray weathering limestone; local concentrations of light-gray to white crystalline dolomite." Graptolites found by Arthur B. Campbell in the slate were determined by Ross (Memorandum, 1964) to be Middle Ordovician (Early Carodoc) age. These rocks are separated from the Old Dominion Limestone of Weaver (Metaline Formation) by a chert unit which is described as "microcrystalline; dark gray to black; laminated to very thin bedded; many secondary quartz veins and numerous small irregular shaped voids that commonly contain red or yellow stained quartz box works. Thinly interlaminated beds and lenses of limestone increase in frequency towards the south." Overlying the limestone and slate unit is another chert unit, this one of unspecified Paleozoic age, which is described as, "thinly laminated, dark green to black, microcrystalline, cut by many limonitic seams and stringers of secondary quartz. Interbedded with thinly laminated siliceous argillite and siltite, which are generally brown but purplish brown locally."

From the above description the Ledbetter Slate appears to be more limy to the south and more closely associated with carbonate rocks than the northern occurrences. Whether the carbonate content in general increases southward, as is indicated for the underlying chert unit of the Hunters quadrangle, is conjectual. The reappearance of chert above the slate indicates that the limy slate was an interlude in an environment where deposition of silica was favored.

To the north of the Hunters quadrangle, in the Inchileum quadrangle, graptolites of both Lower and Middle Ordovician age were collected by A. B. Campbell. The distance of these localities from the top of the Metaline Formation is unknown.

Figure 4A illustrates the relative ages of graptolite collections from Stevens and Pend Oreille County as interpreted by Michael Churkin and Clair Carter of the U.S. Geological Survey (written communication). Churkin and Carter could not always assign the age of a collection to a specific graptolite zone--largely because of poor fossil preservation--and had to consider some collections as belonging to an unknown faunal zone within a broad span. However, of the 37 collections, 15 were given specific zonal assignments and 9 fall in one or the other of two adjacent zones. The specific zonal assignments refer to 6 faunal zones, the earliest, the *Tetragraptus approximatus* zone of the Lower Ordovician and the latest, the *Dicranograptus clingani* zone of the uppermost Middle Ordovician. It seems well established that northeastern Washington was a marine environment during the Early and Middle Ordovician and quite probably during the Late Ordovician, which, as yet is unrepresented by a collection.

Within this great range in age represented by the graptolite faun, it is surprising that no section has been found that has more than two faunal zones and even these, with the exception of the Ordovician-Silurian section on Horsefly Hill in the Metaline quadrangle, do not conclusively demonstrate a time change across the section. It is also surprising that 40 percent of the collections are found within 30 metres of the base of the Ledbetter Slate and that the age of these collections ranges through the total Ordovician time established by the graptolite zones, which include both the Early and Middle Ordovician. The other 60 percent of the collections occur from one to several hundred metres above the base of the Ledbetter and are all of Middle Ordovician age. This means the Ordovician part of the black shale facies is either interrupted by an erosional hiatus or by a thrust fault. Unless a younger over older thrust is postulated, the erosional hypothesis is preferable because Early Ordovician graptolites are found only near the base of the section. The monotonous character of the facies would easily mask any erosional boundaries, particularly if the erosion was the submarine shifting of unconsolidated orsemiconsolidated muds.

The thrust fault explanation finds no support in the geographic distirbution of the collections. The localities of the Early Ordovician graptolites show no systematic variation in position above the top of the Metaline Formation, either from east to west or north to south. Nor can a pattern be produced by restoring tectonic blocks to inferred depositional positions. For example, the Lower Ordovician collections come from the Valley (Metaline quadrangle, Park and Cannon 1943), Hooknose-Baldy, and Lime Creek Mountain blocks, where widespread distribution both before and after the proposed thrust faulting suggests that the deposition of the Lower Ordovician was both extensive and uniform. Nor is there any systematic arrangement of the tectonic blocks containing the Middle Ordovician graptolite zones that occur near the base of the Ledbetter. If thrust faults rearranged the graptolite localities, they were not regional thrusts restricted to the black shale facies.

Argillite B

The unit shown on the map of the regional geology (fig. 14) as Argillite B (Pza) includes rocks that conceivably span a large part of Paleozoic time, although the only fossils found in the unit are Lower(?) Ordovician graptolites. The graptolites are in the Active Formation of British Columbia, which is included here instead of with the restricted Ledbetter Slate, because the Active Formation doubtlessly includes rocks younger than and conceivably older than Ledbetter. The argillites of Russian Creek in the northwestern corner of the Metaline quadrangle are also included because they are probably younger than the Devonian limestones and consequently not part of the Ledbetter. The Grass Mountain sequence of the Deep Creek area is also included and believed to be post-Devonian and correlative, at least in part, with the argillites of Russian

Creek. The age designations for the unfossiliferous parts of the black shale facies are equivocal to the extent that they are debatable.

The Active Formation is a black shale and argillite sequence that overlies the Cambrian Nelway Formation. It was named by Little (1950, p. 21) from its occurrence on Active Creek, a tributary of Porcupine Creek, where Ordovician graptolites were found. Fyles and Hewlett (1959, p. 34) concluded "that the Active Formation in the map area (Salmo lead-zinc district) includes mainly Ordovician rocks, and possibly some rocks that are Silurian and Devonian." At the type locality and elsewhere it closely resembles the Laib Formation and previously had been so assigned. As a black shale facies it has not been recognized more than 32 km north of the International Boundary. Fyles (1964, p. 36) in comparing the rocks of the Salmo area with those of the Duncan Lake area, 120 km to the north, emphasized the striking differences in lithologies of comparable ages, recognizing this as, "one of the major stratigraphic and structural problems of the arc (Kootenay)." He notes that neither the mid-Cambrian carbonate rocks nor Ordovician black shales and argillites appear in the Duncan Lake section. Although correlations are tentative, the Ordovician is probably represented by a quartzite and the mid-Cambrian by a shale. The Duncan Lake rocks can be identified for long distances to the north. In explanation of these relationships, Fyles suggests, "the sedimentary conditions that existed over wide areas in the northern part of the Kootenay Arc and beyond it in early and mid-Paleozoic time differed markedly from those in the southern part of the arc. Whether these conditions produced sedimentary facies changes in a north-south direction parallel to the structural trends or whether contrasting eastern western facies have been brought together by tectonic movements is a problem for further study." The writer agrees with Fyles' analysis and suggests that the trends of facies boundaries could fall anywhere between a direction normal to the trend of the arc and one at an acute angle to it.

The lithology of the Active Formation in the West Nelson map area varies from locality to locality, but the greatest variation is not within but, between the two belts that appear on Little's map (1960, pl. 1). Only the western belt of Active Formation is shown on figure 14 of this report as Argillite B; the eastern belt, which extends only 3 km north of the International Boundary on the east limb of the Sheep Creek anticline, is a continuation of the Ledbetter Slate of the Metaline quadrangle. The west belt, referred to by Fyles and Hewlett (1959, p. 16) as the "Argillite Belt"--one of four structural-stratigraphic belts separated by thrust faults--contains Ordovician graptolites at its northern end, but at its southern end projects into the argillites of Russian Creek in the Metaline quadrangle, rocks believed to be younger than the Devonian reef limestones. The description of the Argillite B unit in British Columbia is therefore a description of the western belt of the Active Formation.

The Active Formation is described by Fyles and Hewlett (1969, p. 34) as composed of argillite and minor amounts of slate, phyllite, argillaceous limestone, and dolomite. Color is almost everywhere black or dark gray, bedding is commonly obscure, cleavage is parallel, or near parallel, to bedding. Detailed descripsions (p. 33-37) are largely concerned with the number, kind, and location of carbonate subunits in the black argillite. Because of unknown structural complexities, no stratigraphic section was determined for the belt, but the distinctive character of some carbonate units is worth repeating.

Limestone appears throughout the Active Formation in the western belt; some is little more than localized limy phases of the argillite, some is as thin interbeds of fine-grained black limestone. Fyles and Hewlett traced some subunits of dark gray banded limestones and dark gray argillacous limestone for as much as 3 km.

Dolomite similar to that in the intraformational breccia unit of the Leadpoint section of the Cambrian Metaline Formation occurs at two places in the western "argillite" belt: (1) on both sides of Sheep Creek, south and east of the Sheep Creek stock (Fyles and Hewlett, 1959, fig. 3, Sheet C), (2) near the head of Harcourt Creek south of the Pend Oreille River (fig. 3, Sheet A). At Harcourt Creek, "dark grey limestone, dolomite and dolomite breccia" have been found as a thin sliver overlying the argillite fault. "Some of the dolomite is massive but most is a sedimentary breccia made up of subparallel tabular fragments of dolomite about half an inch thick and a few inches across in a matrix of dolomite."

The breccia on Sheep Creek was included in the Active Formation because of its dark gray and black color and close association with black argillite. The carbonate unit includes, in addition to the breccia, unbrecciated massive dolomite. The dolomite breccia of Sheep Creek differs from that on Harcourt Creek by including bands and irregular masses of chert and silicated limestone. It is border one the west--probably stratigraphically above--by a band of dark gray limestone and black argillite.

The descriptions of these dolomite breccias and their association with the black shale facies suggest that the part of the Active Formation in which the breccias occur, is possibly correlative with the transitional upper member of the Leadpoint section, and, accordingly, is upper Cambrian in age, which in the Deep Creek area is probably the beginning of the black shale facies. Not only do the descriptions suggest that the black shale facies of the west argillite belt is transitional to the Cambrian carbonate facies, but they support Fyles suggestion that the black shale facies changes northward to a facies of coarser clastics. According to Little's (1960, p. 40) description the most northern exposed Active Formation, at the type locality on Porcupine Creek, includes three limestone bands and argillaceous quartzites. The postulated northward transition of the Active Formation to a correlative quartzite-limestone-argillite in the upper Lardeau Group of Kootenay Lake and equivalent rocks at Duncan Lake is supported by this appearance of quartzite at the northern end of the exposures of Active Formation.

Thirty-two km southwest of the fossil locality on Active Creek, the black argillites of the Active Formation cross the International Boundary into the northwest corner of the Metaline quadrangle, where they become a controversial group of rocks. In this area they are informally named "the argillites of Russian Creek," and have been of uncertain correlation since early studies in the region. As late as 1950, the rocks north of the boundary were considered by Little as part of the Laib Formation. On the United States side they were mapped by Park and Cannon (1943, p. 19-20) as Ledbetter Slate, a conclusion accepted by Dings and Whitebread (1965, p. 26), although they recognized that the rocks were "somewhat different from those found elsewhere" being well bedded, more siliceous, and coarser grained (silt size). Dings explained the differences by interpreting the Russian Creek rocks as a sandy facies of the Ledbetter that was "thrust a considerable distance from its original site of deposition." The writer agrees with this hypothesis of transport by thrusting, but believes the

Russian Creek rocks are post-Ordovician in age, probably equivalent to, or slightly younger than, the pebble bearing argillites that overlay the Devonian reef limestones a few miles to the south.

The Russian Creek argillite is basically a dark gray to black silicious argillite, locally laminated, with a tendency for the laminae to be lenselike. Part, but not all, of the lamination may be structural, produced by movement along and near bedding planes. Black slate, similar to that in the Ledbetter, occurs with the laminated argillite.

Rocks of the Russian Creek type are in the Deep Creek area, on Red Top, Grass, and Stone Mountains and in the Brodie Mountain area south of Northport. All those recognized are within the area indicated as Argillite B on figure 14. Although all occurrences of these argillites were recorded during the mapping, it was not possible to separate them as stratigraphic units, because they are intimately associated with slates, phyllites and other pelitic rocks. They may represent a condition of deposition without regional significance, a condition repeated at different times in different parts of the black shale environment.

Rocks of the Pacific borderland

The boundary between the Pacific borderland and the Kootenay arc is a vague, indefinite boundary both structurally and stratigraphically. The first group of rocks discussed here has lithologies and structures that can be identified with rocks of the arc, but the presence of volcanic rocks sets it off from the miogeosynclinal elements that characterize rocks of the arc. The rocks of this first group, structurally part of the arc, are regarded as a transitional assemblage; the second group is truly representative of the eugeosynclinal environment--and is so labeled.

Rocks transitional to the eugeosynclinal facies

The rock units described in preceeding sections are moderately monolithic, representing fairly uniform conditions of deposition. The transitional assemblage, although retaining the characteristic of predominant fine clastic sediments of the black shale facies, has been modified by changes in depositional conditions to produce two different limestones, as well as such diverse products as basaltic volcanic rocks, chert, and quartzite.

These rocks are mostly cratonically-derived fine clastics but include some volcanics as well. The units included are identified on figure 14--as "Argillite C" and the Flagstaff Mountain sequence, both of questionable Carboniferous age. Stratigraphic position uncertain. They occur northwest of the crest of the Columbia anticline and are in fault contact with other stratigraphic units.

Rock unit, Argillite C, includes the rocks on the lower Pen Oreille River, referred to by Fyles and Hewlett (1959, p. 37-38) as "Sedimentary rocks of the lower Pend Oreille River" and the Pend Oreille sequence of the Deep Creek area described on p. 64-70. Because these two units are continuous across at the International Boundary, they are at least in part equivalent. They include similar rock types, including limestone, chert, greenstone, and quartzite, as well as the predominating black argillite, slate and phyllite. It was not possible to resolve this diverse assemblage into a stratigraphic sequence neither north nor south of the International)undary. Perhaps most of the difficulty is that Argillite C occurs in a 'ly deformed part of the arc, where faults are particularly abundant.

The largest body of limestone Argillite C occurs north of the boundary, adjacent to and a short distance south of the Wanetta thrust facies, which in this area limits the southern extension of the Jurassic volcanics. According to Fyles and Hewlett (1959, p. 38) this belt of limestone is at its west end a discontinuous band of lenses, but at its east end, is a continuous body "ranging in outcrop width from about 150 metres to more than 900 metres." It is light colored, poorly bedded limestone that is "commonly siliceous" and in places dolomitized. It is both overlain and underlain by black phyllite. It contrasts sharply with another type of limestone, which is moderately common in the black argillite as interbeds of fine-grained, impure black limestone. The limestones in the Deep Creek area have neither the thickness nor the continuity of those north of the boundary, however, both above types are represented. The limestones of the transitional facies do not greatly influence the tenor of the assemblage, but are common enough and of sufficient variety to indicate that the overall conditions of deposition were different from any that had previously existed.

The Argillite C rocks that occur in British Columbia, and those placed by Little (1950) in the upper Laib Formation, but Fyles and Hewlett (1959, p. 38) expressed doubt over this decision because of the presence of chert and sheared pyroclastic rocks, which are unknown in the Laib. They however, made no age assignment. Although diagnostic fossils have not been found in this unit either north or south of the boundary, the crinoidal debris mentioned in this report--although indicating to the paleontologist little more than Paleozoic age--is more equitable with the Devonian and younger ages than with Cambrian ages. This inference is supported by the lithologic dissimilarity to any pre-Devonian Formations in the region. If the transitional assemblage is essentially autochthonous, it thus assumes by default a post-Devonian age pre-Permian age; it matches nothing in the Precambrian, nothing in the pre-Mississippian Paleozoic, and nothing in the Permian, which is strictly eugeosynclinal. On the other hand, if it is allochthonous--not completely unrealistic possibility--it is conceivably a lower Paleozoic facies foreign to that exposed in other parts of the arc, although possibly represented in the Lardeau Series to the north.

Flagstaff Mountain sequence

The Flagstaff Mountain is an informal name given to a thick sequence of westward dipping, dominantly pelitic rocks that occur on the west side of Columbia River in the Northport quadrangle (Yates, 1971). The sequence, more than 3,000 metres thick, although broken down into three units, was not formalized as either a formation or group, because it is bounded by faults and has yielded no diagnostic fossils. The only fossil fauna is a crinoid columnal--or echinoid spine-- that gives a time fix that falls somewhere between the Devonian and the early Mesozoic. A further cogent reason for its informal status is the strong possibility that it correlates, at least partially, with the Grass Mountain sequence of the Deep Creek area, which it resembles. Until these two sequences can be assigned more definite ages and their interrelations more precisely established, they should not be enshrined in formal nomenclature.

The Flagstaff Mountain sequence was divided (Yates, 1971) into three subunits, which are: (1) a lower unit consisting of black argillite, phyllite and slate with a few small limestone pods, (2) an intermediate unit, consisting of a siliceous black argillite that contains minor amounts of fine-grained argillaceous quartzite and a thin, altered basaltic rock, and (3) an upper unit, a dark gray to silver gray phyllite that contains several thin limestones, as well as a green (volcanic) phyllite. These three units appear to be a continuous sequence, structurally accordant, without conglomerates or other features that could indicate unconformities. Nevertheless, there is no assurance that they do not embrace a long time span partly represented by unidentified hiatus. Nor is there any positive assurance that the assumed stratigraphic sequence is chronological and not a tectonic reorganization and duplication of strata. The sequence has enough pervasive features to support the belief that it is potentially a distinctive formational unit regardless of the relations between the subunits.

The lowermost unit of the Flagstaff Mountain sequence is dominated by a siliceous black argillite accompanied by silvery gray and black phyllites that are locally limy. Black slates are minor variations; limestone is present, but uncommon. The most prominent limestone located about one mile southeast of Flagstaff Mountain has been replaced at its northeastern end by barite. Bedding in the argillite is commonly obscure; but where present, is recognized through faint color bands, or slight variations in grain size; microbeds were rarely seen. A few argillites resemble those of Russian Creek.

The middle unit is the most siliceous and the most dynamically metamorphosed of the three, but it has no higher grade of thermal metamorphism; it has the same quartz, muscovite, chlorite assemblage as the lower and upper divisions. The basic lithology is a gray to black pelite that has a greater proportion of quartz than the pelites of the other units. Besides the siliceous phyllite schist--which is most abundant in the upper part of the unit, and more dynamically metamorphosed than the other units--are phyllites and argillites similar to those in the other units. The base of the middle unit was placed at the base of the green phyllite schist indicated on the geologic map of the Northport quadrangle (Yates, 1971). Locally within the greenschist and basal to it, is a fine-grained, thin-bedded, gray to greenish gray quartzite.

The greenschist is a highly sheared, almost mylonitic, volcanic rock belonging to the muscovite-chlorite subfacies of the greenschist facies. It has a well-developed planar fabric, derived from the fine comminution of the original minerals, probably calcic plagioclase and pyroxene. The metamorphic assemblage includes albite, chlorite, actinolite and epidote family minerals, that recrystallized from the pulverized rock, apparently in a carbon dioxide medium, because calcite is an important constituent of the rock both as an intimate part of the mass and as crosscutting veinlets. Other veinlets are of epidote and albite.

An analysis and norm of a more massive phase of the greenschist is shown in table 3. The anomalously high alumina and calcium suggest an unusually high anorthite content, possibly indicating a pyroclastic deposit, interpretation supported by the gradation of zoisite bearing rocks into the underlying black argillites.

The upper unit of the Flagstaff Mountain sequence consists of dark gray to black argillites like those of the other units, but is distinguished by numerous thin limestones. The limestones are dense to fine-grained, white to medium dark gray, commonly thin bedded, a few very thin bedded. Some limestones alternate gray with buff beds; others are almost massive, but the thin layered varieties predominate. A few of the limestones, especially those north of Ansaldo Lake, contain "fossil trash," which, although definitely organic, is composed of unidentifiable fragments of fossils. The limestone about one half kilometre northwest of Flagstaff Mountain contained the above mentioned crinoid columnal or echinoid spine.

Table 3.-- Analysis, in percent, of greenschist in Northport quadrangle, Washington

[Rapid rock analysis by Paul Elmore, G. Chloe, L. Artis, D. Taylor, S. Botts, H. Smith, and J. Glenn]

Chemical analysis		CIPW norm	
SiO_2	46.1	qtz	---
Al_2O_3	21.0	or	0.59
Fe_2O_3	.86	ab	10.16
FeO	2.4	an	51.62
MgO	8.5	ne	
CaO	16.5	wo	10.14
Na_2O	1.2	en	11.82
K_2O	.10	fs	2.08
H_2O-	.06	fo	6.55
H_2O+	2.2	fa	1.27
TiO_2	.06	mt	1.25
P_2O_5	.02	hm	..
MnO	.07	il	1.14
CO_2	.92	ap	0.047
Total	100.	cc	2.09
		Total	97.7

Interbedded in the argillite of all three units, are siliceous detrital rocks that superficially resemble cherts. Microscopic examination indicates that they are composed largely of minute angular grains of clastic quartz, grains too small to be rounded by water transport. Because these sediments represent a very different depositional environment from chert formed by chemical or organic processes, they should not be used to interpret the environment of the Flagstaff Mountain sequence as a "chert environment." The most accurate name for them would be siltite.

The upper unit also includes a greenschist, which is very similar to that in the lower unit. It is a highly sheared rock composed of albite, chlorite, epidote and(or) zoisite. The upper and lower units are possibly the same.

Rocks of the eugeosynclinal facies

The eugeosynclinal facies of the Pacific borderland includes all the rock units shown in the explanation of figure 14 that extend in time from the Pennsylvanian to Middle Jurassic. These are: (1) clastic sedimentary rocks and mafic volcanic rocks of the Mount Roberts Formation, (2) sedimentary and volcanic rocks of Unit X, a poorly understood unit of unknown, but probable Permian age, and (3) volcanic and sedimentary rocks of Jurassic age known as the Rossland Group, and (4) limestones of inferred early Mesozoic age.

The Mount Roberts Formation

The Mount Roberts Formation was named and described from outcrops in the Rossland Map area of British Columbia, where fossiliferous limestones were identified as upper Paleozoic. The age of these fossils is "most likely Pennsylvanian but a lower Permian age cannot be entirely ruled out"

(Little, 1960, p. 51). This conclusion is consistent with that of Bowman (1950) who correlated fossiliferous Permian rocks in the south half of the Orient quadrangle with unfossiliferous rocks in the north part. Bowman considered the lithologic similarity of the rocks in these two separated areas to be close enough to warrant the use of the same formational name, the Churchill Formation. Mapping of the Northport quadrangle, which lies immediately east of the Orient quadrangle and south of the Rossland Map area, demonstrates that the Mount Roberts Formation and the Churchill Formation are one and the same unit. Fossils found in the Mount Roberts Formation occur only in the lower member and quite possibly those found in the Churchill Formation occur only in the upper part of that unit; therefore most of both formations may have been deposited in early Permian time. The name Mount Roberts Formation was given to the rocks in the Northport quadrangle, because this name has priority.

The Mount Roberts Formation in British Columbia is composed of two members (Little, 1960, p. 45-53). The lower member, well exposed near Paterson, is composed of limestone, andesite and banded tuff. The limstone which contains the fossils, is crystalline and contains masses of chert; the andesite occurs as flows with brecciated tops; the tuffs are interlayered with the limestone and locally grade into massive black argillite. The upper part of the formation consists of both fissile and massive siliceous black argillite interbedded with a few lava flows, sills and tuffs. The volcanic rocks are most common near the limestones. Rarer rocks in the upper member are intraformational conglomerate and chert pebble conglomerate.

The limestone contains corals, brachiopods, bryozoan, and crinoids, an assemblage that indicate a moderately shallow water environment, a conclusion supported by conglomerates and by plant impressions, including one of a lepidodendron, collected from beds of black argillite.

North of Paterson, near the town of Rossland, the formation consists mainly of "soft, dense, black slates" locally grading to an arenaceous facies. West of Rossland the common rock type is "dark grey to purplish, fine-grained, argillaceous quartzite," which would be a siliceous argillite in the terminology of this report. These rocks grade into paragneiss. The base of the formation is nowhere exposed and the top is an unconformity overlain by Jurassic Rossland rocks.

Rocks of the Mount Roberts Formation extend across the International Boundary into Washington. The limestones containing the fossils, however, extend only a short distance and do not reappear among the fairly widespread outcrops of Mount Roberts rocks in the northwest quarter of the Northport quadrangle. Probably only the upper part--or more probably, the intermediate part--of the Mount Roberts-Churchill sequence is present in the Northport quadrangle.

Well bedded, medium to dark gray siliceous argillites of the upper member are the dominant rock type in Washington, as well as in British Columbia. Interbedded with the argillites are both graywackes and tuffs. Some rocks that resemble lava flows in the field were found to be tuffs in thin section. Minor rock types included in the formation are pods and lenses of chert, fine-grained dark gray limestone, and chert-pebble and intraformational conglomerates. Fine-grained clastics resembling chert are fairly common; "true" nonclastic cherts are rare. The conglomerate pods, rarely more than one foot thick, are composed of clasts of chert and quartzite. Some quartzite clasts have a well-developed planar structure of sutured grains.

By far the most "instructive" rocks in the formation are the feldspathic graywackes, for they are evidence of a significant change in conditions of sedimentation, the beginning of volcanism as a dominant process in the depositional environment of the map area. The graywackes are interlayered with the much more abundant argillites, which are too fine-grained to determine by thin section examination the amount and character of any volcanic detritus that is certainly present; consequently study of the coarser clastic is far more rewarding.

The graywackes range in grain size from fine to very coarse sand, and in color from light to medium gray. They have two pronounced characteristics, highly angular clasts and an abundance of plagioclase. The clasts include plagioclase and quartz, and fragments of chert, limestone, argillite, grit, quartzite, and volcanic rocks. The most abundant clasts are quartz and plagioclase, which together compose over 60 percent of the rock. Rock fragments, commonly subrounded, are rarely more than 25 percent of the total clasts; possibly clasts of volcanic rocks are the most abundant. The quartz is sharply angular, strained, rarely multiple grained, and not sized. The plagioclase (albite) is as individual crystals and broken fragments of crystals. Crystal faces on the plagioclase are generally sharp and well-defined; however, some crystals are rounded with rounding continuing into embayments indicating magmatic resorption of phenocrysts and not rounding by erosional transport. The broken crystals show no effect of attrition; sharply angular corners are common. The rock fragments include fine-grained varieties, some basaltic in composition, others glassy, and a few with trachytic texture. None of the rock fragments contain plagioclase similar to that of the crystal clasts. The clasts of sedimentary rocks, including the rare quartzites, are fine-grained.

The feldspathic graywackes are interpreted as the products of erosion of a volcanic terrane largely composed of pyroclastic rocks, containing abundant loosely held plagioclase crystals. All material had subaqueous transport to its place of deposition. The hypothesis that the plagioclase crystals were deposited directly as a "crystal fall" is untenable, because such an explanation requires that each crystal fall coincide with a high energy pulse of current action necessary to transport the coarse nonvolcanic clasts of the graywackes--which is a coincidence highly unlikely. The lack of rounding of the clasts must indicate a short rapid transport.

Unit X

The rocks listed under Unit X are in the southwest corner of the map area (fig. 14). Unit X is part of the stratigraphic division shown on the Geologic Map of the State of Washington (Huntting, 1961) as a narrow belt extending from the Colville River northeasterly to Brodie Mountain and labeled "Pu" the symbol for "upper Paleozoic undivided." Rocks in this belt are a motley collection of sedimentary and volcanic rocks that defy a more specific classification than "upper Paleozoic, undivided." Unit X is the residue after Argillite B and Argillite C were split off from the State map category. Although Unit X resembles both Argillite B and Argillite C, it is sufficiently different to consider it a separate unit. It is possibly related, however, to the rocks near Kettle Falls symbolized on the State map as "Pm," an assemblage of Permian eugeosynclinal rocks that include graywacke, argillite, limestone, conglomerate, and greenstone (Mills and Davis, 1962, p. 41-42).

Unit X is predominately light to dark gray phyllite that encloses lenses and pods of quartzite, limestone, and greenstone. In some places the phyllite is thinly laminated in shades of gray, and in a few places is limy, especially where associated with lenses of dense dark gray limestone. The quartzite is a dark colored medium-grained rock; banded chert occurs in layers from 1.5 to 10 centimetres in thickness.

The age and stratigraphic significance of Unit X can only be inferred, nevertheless a few facts stand out. It belongs to the argillite facies that dominated Paleozoic sedimentation; it belongs to the Pacific Borderland province; it probably could logically be included with the rock units regarded as transitional to the eugeosynclinal facies, however, its close geographic and lithologic association with the fossiliferous Permian rocks of the eugeosyncline influenced placing it in this facies.

Rossland Group

Although Jurassic volcanic and sedimentary rocks extend southward into the United States, it is only in Canada that they can be assigned formations with adequate fossil control. Frebold and Little (1962) divided the Jurassic rocks of southern British Columbia into a lower unit, the Archibald Formation and its probable equivalent the Ymir Group (which may extend down into the Triassic), and the Rossland Group. The Rossland Group, overlying the Archibald Formation, is composed of the Elise Formation, predominantly andesitic to basaltic lave flows, flow breccias, agglomerates and tuffs; the Hall Formation, predominantly sedimentary rocks with local accumulations of volcanic material; and an upper unit informally called, the upper Rossland Group, predominantly volcanic and without diagnostic fossils. Time of accumulation extends from the lowermost Jurassic stage, the Hettangian(?), to fossil confirmed ages younger than middle Bajocian and possibly to the lowermost Upper Jurassic Stage, the Callovian. Meaningful estimates of the total thickness of Jurassic rocks in southern British Columbia are difficult to make; although local thicknesses exceed 5,500 metres, the average thicknesses are probably much less.

The Jurassic Period produced a sequence of four formational units, representing a cyclic alternation of sedimentation and volcanism. The sedimentary units, the Archibald and Hall Formations, have in addition to a few intercalated lava flows, a great abundance of volcanic detritus in the sedimentary beds. As for example, in describing a common rock type in the Hall Formation, Little (Frebold and Little, 1962, p. 7) says, "The rocks resembles much of the Archibald Formation. It consists of small angular fragments of feldspar (mainly saussuritized plagioclase) and quartz shards, in a groundmass of bladed tremolite, carbonaceous matter and finely divided, unidentified interstitial material."

The volcanic rocks of Jurassic age in southern British Columbia include lava flows, flow breccia, agglomerate, and tuff that range in composition from andesite to basalt. Some lavas are described as porphyritic and have phenocrysts of augite, hornblende and feldspar; all are altered, having saussurtized plagioclase, chlorite, and epidote as common minerals. They have been subjected to the same regional, greenschist, metamorphism as the older rocks. The only published analyses of these rocks are those of Daly (1912, p. 323-330). He described the Rossland Volcanic Group--which would include the Elise Formation--as including seven varieties of latite, as well as olivine basalt, olivine free basalt, and augite andesite. Four of these rocks were analyzed, augite latite, augite-biotite latite, augite olivine latite and hornblende-augite latite. The high potash content of these rocks, from 3.08 to 5.44 percent, and a mineralogy that includes biotite and orthoclase is anomalous for the Rossland lavas described by other workers Little (1960), Mulligan (1952), and Bowman (1950)as basalts and andesites. Possibly Daly's analyzed samples are not from Jurassic, but from Eocene lavas that cap the Jurassic rocks. The Eocene lavas belong to an alkaline suite and are petrographically similar to the described and analyzed samples.

The Jurassic rocks in the Northport quadrangle of Washington have not yielded fossils, therefore it is not possible to make more than the most general correlation with the Jurassic of Canada. The reconnaissance study on which this description is based yields little more than the distribution of the Jurassic volcanic rocks. West of the Northport quadrangle, in the Orient quadrangle, Bowman (1950, p. 54) describes the "Rossland volcanic group" as containing dark-gray argillite, graywacke, a minor amount of chert and a considerable amount of conglomerate interbedded with the volcanic rocks, but he was unable to arrange the rocks in a stratigraphic sequence nor to find fossils. Bowman recognized two volcanic rocks, an augite andesite, the most common, and a massive andesite.

Although reconnaissance mapping of the Northport quadrangle yielded no detailed knowledge of the Rossland rocks, it showed that the Rossland here has rock type rock compositions similar to those in the Orient quadrangle and those described by Little (1960, p. 65-70) in British Columbia. Flows, flow breccias, tuffs, and intercalated sedimentary rocks high in volcanic detritus are the rock types represented in all three areas.

In the Northport quadrangle pyroclastic rocks appear to be far more abundant than flow rocks. Of the flow rocks two varieties are represented among the specimens collected. The common variety is an augite-plagioclase rock of intergranular texture, which without chemical analysis is classified as a basalt. The other variety, of rare occurrence, is a porphyritic rock having phenocrysts of green euhedral hornblende in a matrix of plagioclase (An_{45}), chlorite and actinolite. This is called a hornblende andesite.

Post-Rossland limestones

Limestones identified as "Jl" on figure 14 occur in the northern part of the Northport quadrangle in five places. At three places the limestones, in part reworked into conglomerates, rest upon greenstone believed part of the Rossland Group and are overlain by Sophie Mountain Formation of probable Cretaceous age. The fourth occurrence, also conglomeratic, is in fault contact with rocks of the Mount Roberts Formation and is overlain by Sophie Mountain Formation. The fifth occurrence, about 3 kilometres northeast of Elbow Lake, is not conglomeratic and is surrounded by unconsolidated glacial deposits. The Upper Jurassic age selected for these limestones is based upon their position upon supposed Lower or Middle Jurassic greenstones and under late Cretaceous conglomerate (Sophie Mountain Formation).

Sophie Mountain Formation

The Sophie Mountain Formation is west of the Columbia River, where it occurs as isolated patches of conglomerate, which, in the aggregate, form a poorly defined east-west belt that nowhere extends more than 3 kilometres north and 10 kilometres south of the International Boundary. The patches of outcrop (fig. 20), remnants of a presumably once continuous mantle of Late Cretaceous gravel, are preserved in what is interpreted as a downwarp that trends a little north of west along the boundary.

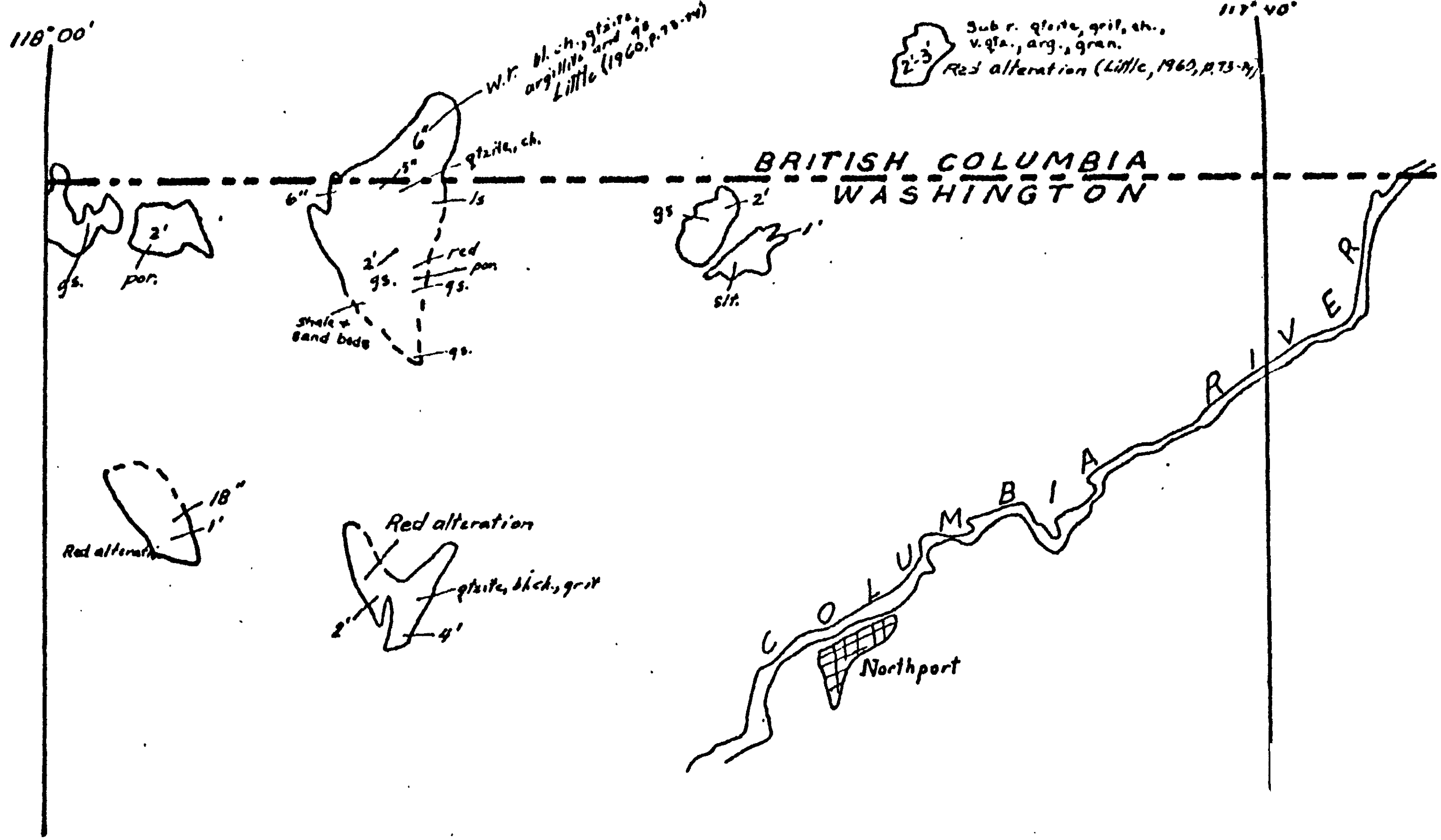


Figure 20.-- Map showing distribution of outcrop areas of Sophie Mountain Formation and scattered observations on clast compositions, maximum observed clast size, and presence of red alteration. Abbreviations used: arg. = argillite, bl. ch. = black chert, gran. = granitic rock, gs. = greenstone, ls. = limestone, por. = porphyry, sh. = shale, slt. = siltite, sub. r. = subrounded, v. qtz. = vein quartz, w. r. = well rounded, qtzite. = quartzite.

The Sophie Mountain Formation is conglomerate and little else. Although it contains interbeds of siltstone, sandstone, and shale, more than 95 percent of the formation is composed of well-cemented gravels, whose roundstones range from pebbles to boulders over one metre in length. It was not systematically studied, but enough field observations were made and enough specimens were collected and studied in thin section to arrive at a few valid generalizations on composition and character and, accordingly, to a reasonable interpretation of its origin.

The conglomerate is a rock of many colors. The color comes largely from that of the predominating clasts, but in places it is masked or was changed by hydrothermal alterations related to emplacement of the Coryell batholith. As a result of such alterations some rocks are green from epidotization or chloritization, whereas others are green from the abundance of greenstone clasts. Other outcrops, particularly those on the low ridge southeast of Elbow Lake, are red from the stain of oxidized iron sulfide minerals. Most outcrops, however, range from light to dark gray, the hue depending largely upon the ratio of dark colored chert and argillite to light colored chert and quartzite.

The conglomerate is a hard resistant rock, partly because of the high content of siliceous clasts, but perhaps more important, because of the siliceous cement that binds it so tightly that it is--where composition of clasts permits--a quartzite. The cement, nevertheless, is minor in quantity, commonly little more than a film welding together clasts or roundstones. Nor is there an abundance of fine-grained mud or silt matrixing the coarser material. Instead there is a seriate range of clasts, from boulders, through cobbles and pebbles, to sand size clasts, all packed together in a tight arrangement. The result is a completely unsorted rock.

The term roundstone, and those of the subclasses, boulders, cobbles, and pebbles, are convenient names for various sizes of clasts that compose the conglomerate, but the implication of roundness makes them highly misleading designations. Nor does "sharpstone" with its implication of pronounced angularity, describe more accurately the average shape of the clasts. In general, roundness is a function of size: the boulders are subrounded and the sand grains (quartz) are highly angular. Exceptions occur in the softer clasts, such as those of limestone and phyllite, which, regardless of size, are rounded, and in the occasional well-rounded cobbles of quartzite that are completely out of phase with their angular companions. These anomalously round cobbles invite the suggestion that their roundness was inherited from an earlier erosion cycle, an interpretation supported by rare clasts of conglomerate composed of well rounded clasts. On the other hand, the degree of roundness should not be equated against size, composition, or recycling, but may be influenced by the unknown factors of distance traveled and provenance. For example, the igneous clasts from porphyritic lavas or dikes of unknown provenance are much better rounded than clasts of contrasting lithologies representing quite different provenances.

The preceding descriptions of shape variations, illustrating the rounding characteristics of several rock types in the conglomerate, is an introduction to the great diversity of rocks represented by the clasts. Not all rock types are found in all outcrops: some outcrops are dominated by one rock type, other outcrops, at no great distance, are dominated by another. Some unusual types are found in only one area. Rarely do clasts of the underlying bedrock type dominate an outcrop area.

The clasts of the Sophie Mountain Formation can be divided into those derived from rock types indigenous to the outcrop area and those that are exotic. A third group of clasts falls in a negative category--those that are not present. Indigenous clasts include those from immediately underlying formations, the Rossland Group and Mount Roberts Formation. Also included are argillites, phyllites, and slates, similar to those in the Grass Mountain and Flagstaff Mountain sequences.

Most clasts, however, are of the group foreign to the map area (fig. 14). Principal rocks represented are quartzites, cherts, vein quartz, greenstone, and porphyritic, fine-grained felsitic igneous rocks; minor rock types are conglomerate, red limestone, phyllite, slate, etc. The quartzites are medium to coarse grained, white to gray, and range from those more than 99 percent silica to those having 10 percent of impurities, largely sericite. Some are mosaics of grains with sutured boundaries, others are closely packed spherical grains showning no recrystallization. The nearest quartzites that could be the source of the round grain varieties are the quartzites that occur in the unit considered as possible Ledbetter Slate found east of China Bend of the Columbia River in the Northport quadrangle (Yates, 1970); but this correlation is tentative. All the quartzite clasts are too coarse grained to be from rocks of the Belt Supergroup, nor do they resemble Cambrian quartzites in the Hooknose and Columbia anticlines. The chert range from white to black; no red or green varieties were observed. Clasts range from sand size to 4 inches and are round to very angular. A few clasts examined in thin section have circular areas suggestive of radiolaria.

Aside from the above described clasts and the rare clasts of fine-grained red limestone, recycled conglomerate, and locally abundant angular clasts of vein quartz, the remainder of the conglomerate is composed mostly of clasts of igneous rocks. The recycled conglomerate is of two varieties; one a sharpstone conglomerate of angular chert fragments and the other a conglomerate of rounded clasts of chert, quartzite, vein quartz, rhyolite porphyry, and rare graphic granite.

In Washington the clasts of igneous rocks, except greenstone, are most abundant on the south slope of Sophie Mountain, where locally they are the bulk of the conglomerate. Most are moderately well rounded and in the cobble size range. They range from feldspathic andesite to rhyodacite and almost all are porphyritic with feldspar phenocrysts. Others are granophyric and some are aplites. The textures are those commonly associated with dike rocks and small shallow intrusives, but probably some extrusive rocks are represented.

Rocks conspicuously unrepresented by clasts are the Cambrian quartzites, phyllites, and carbonate rocks. Coarse-grained granitic rocks that might have come from the Spirit pluton, Kaniksu or Nelson batholiths were not observed; however, Little (1960, p. 73) reports that on Lake Mountain in British Columbia at the eastern end of the conglomerate belt, "The largest and most abundant roundstone are of brownish purple, highly altered granitic rock, but many are of white to gray, fine- to medium-grained granite different from local phases of the Nelson plutonic rocks." The absence of clasts of schist and gneisses that might have been derived from the Shuswap terrane which lies to the west and north was noticeable.

The problem of source of the clasts is no greater than that of the structure of the formation, which also has puzzling aspects. Depositional contacts between the conglomerate and the rocks of the Rossland Group and Mount Roberts Formation were not observed, although they doubtless exist. Some, and possibly most, contacts are faults, although the map pattern and relations between outcrops and topography indicate that the conglomerate rests unconformably on the older rocks and is only gently folded. The measurements of structures of high angle believed to represent bedding attitudes are inconsistent with the interpretation that the formation is little deformed. Because of the great paucity of bedding, most measurements were made on the common orientation of the planar element of clasts, which admittedly does not necessarily represent a horizontal plane at the time of deposition. The relation between the "bedding" attitude and the topographic position of the contact might suggest that the conglomerate is a thrust sheet--a conclusion believed quite improbable.

The maximum age of the Sophie Mountain Formation is limited by its depositional relation to the Jurassic Rossland rocks and its minimum age by the middle Eocene satellitic dikes of the Coryell batholith (Yates and Engels, 1968, p. D242-274) that cut it. Paleontological evidence of an intermediate age is given by W. A. Bell of the Canadian Geological Survey (Little, 1960, p. 74), "Plant locality" Cat. No. 3890. From the top of Mount Sophie, 300 feet N. 20° E. of Boundary Monument No. 174, and an additional 660 feet north of this point, elevation 5,000 feet.

"In addition to imprints of rootlests and stems this collection has a number of fragments of _dicotyledonous_ leaves and several of a _sphenopterid_ fern. All the material is unidentifiable, although some of the leaves are apparently platanoid. Although a Tertiary age is not ruled out an Upper Cretaceous age seems more probable."

The assignment of a provisional Late Cretaceous age to the Sophie Mountain Formation leaves the time between Mid-Jurassic and Late Cretaceous without a depositional record. This is the time of the Mesozoic orogeny that produced the tight folds of the arc, the later cross fold and thrusts, and the Kaniksu and Nelson batholiths. The deposition of the Sophie conglomerate was during late stages of the orogeny, but not at the end. It was followed by middle Eocene plutonism and volcanism, accompanied by doming and extension faulting.

Volcanic and sedimentary rocks of the Cenozoic

The Cenozoic deposits of that part of northeastern Washington shown on figure 14, consist of Holocene and Pleistocene and semiconsolidated and consolidated Tertiary deposits which are largely remnants of downfaulted blocks. The semiconsolidated deposits of the Tiger Formation, which is conglomerate, sandstone, and clays of continental origin and unspecified Tertiary age occur in the low hills on the west side of the Pend Oreille Valley in the Metaline quadrangle. The consolidated deposits are conglomerates, tuffs, and lavas of Eocene age preserved at three places in the Northport quadrangle. These are described separately below. For the Cenozoic record in the British Columbia part of the map area the reader is referred to Little (1960, p. 74-76).

The unconsolidated deposits of the Cenozoic are dominantly Pleistocene in age and glacial in origin, the Holocene deposits, talus, stream gravels, etc., are inconsequential--neither group is described here. The Pleistocene deposits consist mostly of glacialfluvial gravels and erratic boulders that came to rest on all but the highest peaks, lake deposits of the Pend Oreille River Valley, and sand, silts, and gravels of the Columbia River terraces, the oldest and highest of which, pocked with kettle holes, records the final retreat of Pleistocene ice. I am not prepared to discuss the Pleistocene history and although recognizing the glacial deposits as a fascinating field of research, I'm inclined to regard them as the major handicap to deciphering the pre-Tertiary history.

The Tiger Formation as described by Park and Cannon (1943, p. 23) includes conglomerates, sandstones, and clays. The predominant conglomerate, a crudely sorted rock with subangular clasts and ill defined bedding contrasts with the sharply defined layers of sandstone and clay that are lignitic in places. The conglomerates locally are nonolithologic; that which lies east of underlying Ledbetter Slate is mostly slate clasts, that east of limestone or granite, mostly limestone or granite clasts, respectively. The poorly preserved leaves found in the fine-grained sediments give no conclusions as to age. On the other hand the 1 to 3 metre boulders of granite and the westward dips, locally as much as 20 degrees, suggests accumulation at the base of the Flume Creek fault scarp and renewed movement on the fault after the accumulation of the Tiger. The Tertiary age assigned by Park and Cannon seems fully justified.

The consolidated deposits of the Cenozoic are restricted to the Eocene. They occur at three places, all in the Northport quadrangle: (1) the Williams Lake area in the southwest quarter, (2) Grouse Mountain, in the northeast quarter, and (3) Belshazzar Mountain, in the northwest quarter of the quadrangle. Their distribution results from preservation by downfaulting and is independent of the distribution of the next oldest rocks, the Sophie Mountain Formation, which was seen in contact with the Eocene rocks in only one place. Rocks included are lava flows of several kinds, crystal tuffs of subaqueous deposition and associated nonvolcanic sediments. The assemblage in each area is different and is described separately.

Williams Lake area

Considerably more time was spent in studying the Eocene rocks in the Williams Lake area than the combined time spent in the other two areas, but even here the study was too brief to do little more than outline problems related to the origin and history of these rocks. Knowledge of the Belshazzar and Grouse Mountain rocks comes from two traverses across the areas.

The Williams Lake section in its two major stratigraphic units is comparable to and correlative with the Eocene rocks in the southern part of the Republic area (Muessig, 1967, p. 45-70), 55 kilometres to the west, but is not comparable to the Tertiary rocks Bowman (1950) describes in the Orient district, 24 kilometres to the northwest. The sequence at Belshazzar Mountain resembles the very uppermost part of the Williams Lake section; that at Grouse Mountain has rock types common to the others, but supposedly is too incomplete for correlation.

The roughly rectangular 20 square mile area of Eocene rocks centered about Williams Lake, is bounded on all but the south side by faults. In addition to the bounding faults, it is internally segmented so that it consists of three tilted fault blocks. It is shown on the geologic map of the Northport quadrangle (Yates, 1971) as the Williams Lake block, which includes in addition to the Eocene rocks, the underlying argillite of the "Cx" unit of probably Carboniferous age.

Volcanic rocks dominate the Eocene section and occur above an erratically distributed nonvolcanic conglomerate in three stratigraphic units which are, a lower crystal tuff, the O'Brien Creek Formation; an intermediate unit of rhyodacite flows, the San Poil Volcanics, and an upper, poorly represented unit of biotite-bearing mafic lavas (shown on figure 14 as "Tj"). Intercalated in the two lower units and separating the middle from the upper unit are conglomerates of nonvolcanic origin. These layered rocks are intruded by dikes of rhyodacite and sills of shonkinite.

The O'Brien Creek Formation, where it is not separated from the older rocks by a fault, rests, in almost all places, on a conglomerate of coarse, unsorted, nonvolcanic detritus. The material in the conglomerate reflects the composition of the nearby pre-Tertiary bed rock: that near Williams Lake is mainly clasts of phyllite, argillite, dark quartzite and fine-grained limestone; that to the north towards the Columbia River is similar; whereas that at the east edge of the Williams Lake block is largely granitic material from the nearby Spirit pluton. This granitic phase is in part a boulder conglomerate, and in places a megabreccia composed entirely of granitic detritus, which includes blocks 10 metres or more in length. It intertongues with the lower part of the O'Brien Creek Formation, thus demonstrating contemporaneity, and it accumulated by gravity slides from a fault scarp that existed before the O'Brien Creek, thus indicating block faulting began before volcanism.

O'Brien Creek Formation and San Poil Volcanics are names given by Meussig (1962, p. D56-D58) to rock units in the Republic Mining District. They were extended to rocks in the Northport district by the writer (Yates, 1971) because of close similarities of lithology and age. The correlation is not meant to imply that the formations were continuous between the two areas, although this is not improbable.

The O'Brien Creek Formation in the Northport quadrangle is both directly and indirectly of pyroclastic origin: some beds are subaqueous crystal tuff, others are probably subareal; much, however, is volcanic detritus reworked by sedimentary processes. Interlayered with medium to coarse-grained crystals tuffs are beds of fine-grained tuffs along with beds too fine-grained to be certain of their composition without chemical analysis, but inferred to be composed likewise dominantly of volcanic material. Many layers contain only minor amounts--commonly 2 percent or less--of nonvolcanic material. These exotic grains and clasts have a source similar to that of pebbles and cobbles in the thin layers and lenses of conglomerate that occur throughout the formation. The clasts are of the underlying argillite, fine-grained quartzite, limestone and granitic rocks from the Spirit pluton. In places thin coal seams and carbonized plant fragments, largely reeds, are present.

From the above it is obvious that not all the rocks in the O'Brien Creek Formation can properly be called tuffs. The coal seams and conglomerates are obviously not, and many beds, particularly the finer-grained rocks, probably have gone through a sedimentary cycle, complete from transportation through diagenesis, but the derivation of the great bulk of their constituents from unconsolidated crystal falls of contemporaneous volcanism, ties them closely to a volcanic event. To call the formation a pyroclastic deposit doubtless stretches a point; nevertheless, it relates the rock more closely to the process that brought it into being, than would sedimentary terminology, such as the mineralogically correct term, arkose, for the coarser varieties.

The O'Brien Creek Formation, which is commonly thin bedded and laminated, but also massive and medium bedded, range from light gray tan through greenish gray to green brown, and the rarer green-black and black shaley layers. In general, the lighter the color the coarser the grain, which ranges from 2 mm to a fine silt. Most beds are well sorted, particularly among the material of volcanic origin; some are diamictites, having widely spaced pebbles in a matrix of sand. The sand-size grains are angular, and mostly quartz and feldspar fragments of crystals. The grains of nonvolcanic material, particularly those of argillite and phyllite are commonly rounded, flattened chips lying parallel to bedding planes. Other nonvolcanic clasts are chert, quartz, limestone and quartzite. The volcanic clasts which predominate are plagioclase, quartz, orthoclase, and biotite, and pumice and other volcanic rock fragments, listed in the order of abundance. Sedimentary structures include mud cracks, cross-bedding, channels, slump structures, and intraformational conglomerate.

From the above described features of the O'Brien Creek Formation, the general environment of deposition can be reconstructed. The topography at the beginning of O'Brien Creek time had considerable relief. Conglomerate did not blanket the area, but was irregular in thickness and distribution; the granitic conglomerate on the east side that interfingered with the tuffs could only come from a scarp to the east, hence we can infer that faulting began before the basal conglomerate was deposited. The Williams Lake block, however, does not appear to have been down-faulted to any considerable extent during the accumulation of the tuffs, although the intraformational breccias and channeling may indicate minor movements on the bounding faults. It seems probable that during the Eocene the Williams Lake area was a tectonic and topographic low in which a lake formed. Coarse cobbles in the conglomerate indicate that pre-Eocene bedrock was only a short distance away. Eruptions from some unknown, but probably distant volcano showered both lake and adjacent highlands with ejecta, largely crystals from which most of the fine glass dust had been winnowed. Crystals and glassy ash that fell in the lake became aquagene tuffs; that which fell on the land was eroded and washed into the lake basin. Because very little nonvolcanic material was washed into the lake it is inferred that the land surface must have been blanketed by early crystal falls and remained nearly covered during the accumulation of the tuffs in the basin. An alternate explanation is that pyroclastic debris covered the terrain, but that this unconsolidated material soon washed off the slopes, choked the streams and filled the valleys. The loose bedrock detritus in the gullies, canyons, and valleys was now buried by pyroclastic debris; that on the slopes, was all that was available for transportation and it moved into the stream channels at a much slower rate than the volcanic debris in the stream channes could move into the lake basin.

In the Northport quadrangle, the San Poil Volcanics are recognized only in the Williams Lake block. Here they occur in the eastern half of the block in three places; two occurrences are erosional remnants capping higher hills of O'Brien Creek and the third is in an elongate graben down-faulted into O'Brien Creek. The formation at the graben occurrence is about 460 metres thick; at the other two places only the lower few hundred feet escaped erosion. Basaltic flows overlie the rhyodacite.

Stratified in the formation is a 60 metres thick conglomerate composed of clasts from the Spirit pluton and underlying rhyodacite as well as those of phyllite and argillite. These are enclosed in a sandy matrix which resembles the O'Brien tuff.

The lava flows have none of the features commonly associated with flows. They are massive, completely free of stratification, vesicularity, sheeting, and other structures typical of extrusive rocks. Lower contacts with the underlying O'Brien Creek are nowhere exposed; lavas in the upper part of the formation are separated from the lower by the afore-ementioned conglomerate, which probably rests on an erosion surface. The intercalated conglomerate is the only convincing evidence of their extrusive character.

The rhyodacite is a gray-green porphyritic rock with phenocrysts of plagiocalse, hornblende and minor biotite. The phenocrysts are euhedral except for occasional rounded biotites in some thin sections. The hornblende is greenish brown and has dark brown rims. The plagioclase contains inclusions of altered glass.

According to the chemical analyses (fig. 20A) the rock has a composition between dacite and rhyodacite in the classification of Nockold (1954, p. 1014-1015). The ratio between Na_2O and K_2O places it closer to rhyodacite. Norms indicate about 15 percent quartz, which was not detected in thin section and about 17 percent orthoclase, which also was not identified. Both minerals occur in the groundmass.

Chemically the rock compares favorably with a sample of the San Poil Volcanics from the Republic district (fig. 20A) and less so with more mafic, more potassic lavas in the Newport district (Miller, 1974. The relations between the felsitic constituents is shown in figure 20A.

Figure 20A -- Analyses, in percent, of volcanic rocks of the San Poil Volcanics and related formations in northeastern Washington

	Northport quadrangle 1/			Republic Quad. 2/	Bald Knob quadrangle 3/			Tortle lake quad. 4/		Newport quad. 5/	
	1	2	3	4	5	6	7	8	9	10	11
Chemical analyses											
SiO_2	63.2	62.3	61.2	61.8	63.08	66.79	67.05	61.0	61.0	57.3	61.7
Al_2O_3	16.1	15.8	15.8	15.9	15.58	14.87	15.27	16.5	16.6	14.8	15.6
Fe_2O_3	3.5	3.0	1.8	3.3	2.81	1.29	1.45	2.4	4.6	3.8	4.8
FeO	1.4	1.5	2.7	2.4	1.61	1.26	1.07	3.0	.88	4.0	60
MgO	1.8	2.0	3.0	2.7	2.55	1.49	1.38	2.8	2.3	4.8	3.3
CaO	4.7	5.0	5.0	4.4	4.56	3.34	3.04	5.4	4.6	5.6	4.5
Na_2O	3.7	4.0	3.6	3.9	3.94	3.59	4.13	3.7	3.1	2.7	3.5
K_2O	2.8	2.8	2.7	2.7	2.65	3.10	3.11	1.7	3.3	3.3	3.5
H_2O-	.88	.83	.57	1.5	.98	.33	.75	2.0	1.8	1.5	.82
H_2O+	.72	.67	1.4	.26	.90	1.46	1.67			.50	.48
TiO_2	.70	.65	.72	.71	.67	.41	.40	.81	.82	1.0	.74
P_2O_5	.27	.25	.27	.24	.26	.15	.15	.31	.33	.48	.42
MnO	.11	.11	.09	.09	.08	.05	.05	.10	.10	.09	.03
CO_2	.06	.42	1.2	—	.06	1.63	.15	.06	.05	< .05	< .05
Total	100	99.	100.	99.9	99.73	99.76	99.67	99.78	99.48	100.	100.
CIPW norms											
qtz	19.60	17.38	17.66	16.24	18.16	29.28	23.37	17.49	18.22	11.27	15.23
or	16.56	16.66	15.95	15.97	15.70	18.35	18.44	10.07	19.60	19.53	20.69
ab	31.33	34.08	30.45	33.03	33.43	30.51	35.06	31.38	26.37	22.88	29.62
an	19.06	17.00	15.45	17.92	17.05	5.10	13.20	23.47	20.46	18.54	16.52
ne											
wo	.89	1.53		.99	1.48	—		4.15		2.60	1.22
en	4.49	5.02	7.47	6.73	6.37	3.74	3.45	7.00	5.76	11.97	8.22
fs			2.85	.68	......	0.73	.20	2.38		2.73	
fo											
fa						(C) 3.76	(C) .29		(C .47)		(Tn .10)
mt	2.84	3.33	~~2.61~~	4.79	3.52	1.89	2.11	3.49	0.79	5.52	
hm	1.54	0.72			.39				4.08		4.80
il	1.33	1.24	1.37	1.35	1.28	.78	.76	1.54	1.57	1.90	1.33
ap	.64	.56	.64	.57	.62	.36	.36	0.74	.79	1.14	1.00
cc	.14	.96	2.73		.14	3.71	.34	.14	.11		
Total	100.02	100.02	100.02	98.25	98.13	98.22	97.58	98.01	98.21	98.02	98.72

1/ Ryodacites from Williams Lake area: 1, N76½°E, 2.3 miles from east edge of Williams Lake; 2, N25°E, 2.3 miles from north edge of Williams Lake; 3, west .7 mile from Lookout Mountain (3533T - B.M.) All collected from near base of San Poil Volcanics. Rapid rock analyses by Paul Elmore, Sam Botts and Lowell Artis.

2/ San Poil Volcanics, Meussig (1967, table 3, p. 53)

3/ San Poil Volcanics, Staatz (1964, table 5, p. F41)

4/ Gerome andesite, Becraft and Weis (1963, table 3, nos. 6W-502 A and 6W-502 B)

5/ Pend Oreille andesite (Miller, written comm., 1976). Analysts: P. Elmore, S. Botts, and L. Artis.

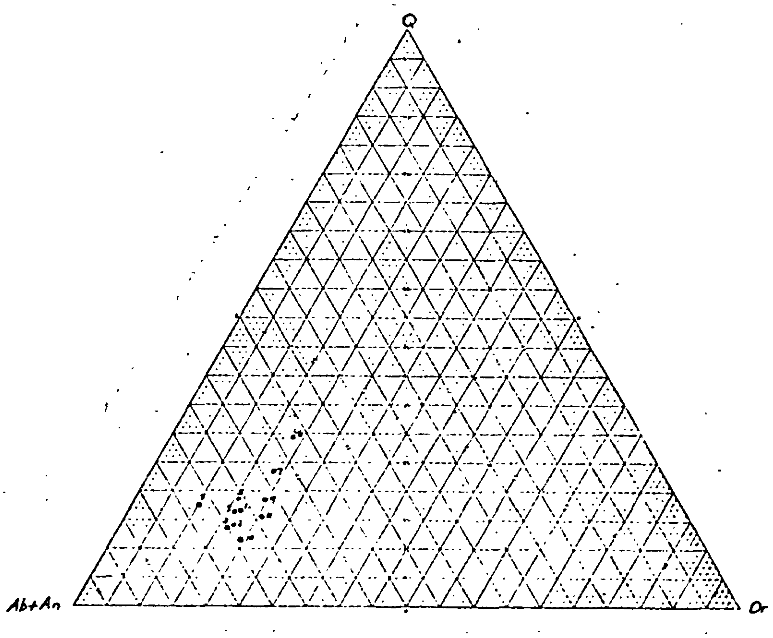


Figure 20A.-- Chemical analyses of the San Poil Volcanics and related formations in northeastern Washington and triangular diagram showing their chemical relations.

On the top of Lookout Mountain the San Poil Volcanics are overlain by about 90 metres of biotite-bearing basaltic lavas. The contact is probably an erosional unconformity because on the south side of the peak the two formations are separated by a sliver, about 30 metres thick, of coarse conglomerate composed of cobbles of San Poil rhyodacite. The rocks overlying the San Poil and the conglomerate are dark aphanitic to glassy porphyritic lavas. The phenocrysts are light colored feldspar and black biotite and hornblende.

In thin section, textures range from hyalopilitic to pilotaxitic. The phenocrysts are plagioclase (An_{45-50}), biotite and brown hornblende. The groundmass that is not glassy consists of plagioclase crystallites (An_{45}) and euhedral and rounded monoclinic pyroxene about 3 times the size of the crystallites. Glass of the matrix is included in the plagioclase phenocrysts. Most hornblende and biotite phenocrysts are oxidized to opaque black crystals, recognized largely by crystal outline and by comparison with the much rarer crystals that are only oxidized at the outer margins. The biotite occurs both as rounded prisms and as inclusions in the hornblende.

Belshazzar Mountain area

The Belshazzar Mountain area of Tertiary rocks begins where the Williams Lake section ends--with the mafic flow rocks. Only two traverses were made across this area, which is not enough to establish a stratigraphic sequence. In the northern part of Belshazzar Mountain and in the Grouse Mountain area as well, alteration of the rocks to chlorite and carbonate minerals is intense and in some places complete--only texture remains.

The base of the volcanics is exposed only at the south end of the mountain, where flow rocks rest on conglomerate. This conglomerate at the base of the volcanic pile, differs from that in the Williams Lake area by including detritus from older Tertiary volcanic rock. Also included are angular clasts of black chert, phyllite, and argillite and well rounded quartzite cobbles. The clasts, from 1 1/2 to 30 centrimetres in diameter, are set in a light brown sandy matrix. The conglomerate includes beds of fine sand and shale.

The flow rocks are interlayered with tuff breccias and agglomerate. The lavas are dark, dense, commonly porphyritic rocks that are moderately diverse in mineralogy, and probably include both basaltic and andesitic varieties. Chemical analyses were not made and alteration of plagioclase prevents accurate An determinations, consequently any specific name applied to the rocks would be subject to revision. All are mafic rocks with a felty or pilotaxitic groundmass of plagioclase microlites. Phenocrysts of monoclinic pyroxene and olivine are abundant, those of hypersthene sparse. Oxidized hornblende phenocrysts were seen in many thin sections, a few rounded crystals of red brown, highly pleochroic biotite were seen in section of all rocks. Inclusions of quartz from fist- to microscopic size, recrystallized and rimmed by reaction products are common megascopically or in thin section.

Grouse Mountain area

The Tertiary volcanic rocks in a small area on the south slope of Grouse Mountain (fig. 14) although very highly altered appear to be closely related to the less altered rocks on Belshazzar Mountain. On Grouse Mountain they also rest upon an irregularly distributed basal conglomerate composed of mixed volcanic and pre-Tertiary clasts. The volcanic rocks include flows, tuffs, and flow breccias. The flows are characterized by resorbed biotite phenocrysts and inclusions of quartz. Textures and phenocryst minerals are comparable to those in the rocks at Belshazzar Mountain.

Age

Other than a few poorly preserved plant impressions in the O'Brien Creek tuffaceous rocks, the Cenozoic rocks of the Northport quadrangle are unfossiliferous. Isotopic dating of representative rocks has compensated for this lack of fossils. Joan C. Engels (Yates and Engels, 1968, D242-D247) dated by the potassium-argon method samples of the San Poil volcanics of the Williams Lake area and a "hornblende andesite" from the Belshazzar Mountain area. The attempted dating of the biotite in the crystal tuffs of the O'Brien Creek Formation was unsuccessful; the ages measured are anomalous, being younger than the rocks that intrude the tuffs. According to the ages of both extrusive and intrusive rocks the Cenozoic igneous event occurred about 50 million years ago. Ages determined for the various rock samples are shown in Table 4.

Table 4

	Age (m.y.) ±Zo
Hornblende andesite, Belshazzar Mountain	50.5±1.5
Biotite "andesite, Belshazzar Mountain	50.4±1.5
Rhyodacite, Williams Lake area, San Poil volcanics	49.8±1.5
Rocks intrusive into O'Brien Creek volcanics	
Skonkinite	49.7±1.5
Lamprophyre dike	49.9±1.5

 (page 273 follows)

Regional structure

In the introduction to a preceding section, where regional stratigraphy is described in chronologic sequence, it is pointed out that the grouping of rocks is geographic as well as chronologic and--in a very general sense--defines three geologic provinces having contrasting environments and tectonic histories. It is also pointed out that boundaries between provinces are not precise, but arbitrary, because of structural and stratigraphic gradations.

Of the three provinces, the Belt-Purcell anticlinorium, the Kootenay Arc, and the Pacific borderland, only the Kootenay Arc lies entirely within the area of figure 14; only fringes of the other two provinces are represented. Consequently, the purpose of this section on regional structure is to give a structural cross section across the arc and relations between the arc and neighboring provinces.

Although the structure of the map area can be described within the framework of these three geologic provinces, it can be described more coherently under divisions based solely upon structure, and to these we turn. The area of the regional map (fig. 14) is divisable into four northeasterly trending structural belts (fig. 21) which, from southeast to northwest are: the homocline belt, fold belt, thrust belt and Jurassic volcanic belt. The homocline belt bridges the Belt-Purcell anticlinorium and the Kootenay Arc, the northeast trending fold and thrust belts fall within the Kootenay Arc, and the Jurassic volcanic belt, which as arc structures in its eastern part, corresponds to the eugeosynclinal province of the Pacific borderland. Although faults serve as arbitrary boundaries between belts--which are accordingly more precisely separated than are the depositional provinces--there is an overlapping of the features that characterize the belts. Seemingly the intensity of deformation gradually increases from southeast to northwest reaching a peak in the thrust belt.

(follows page 271

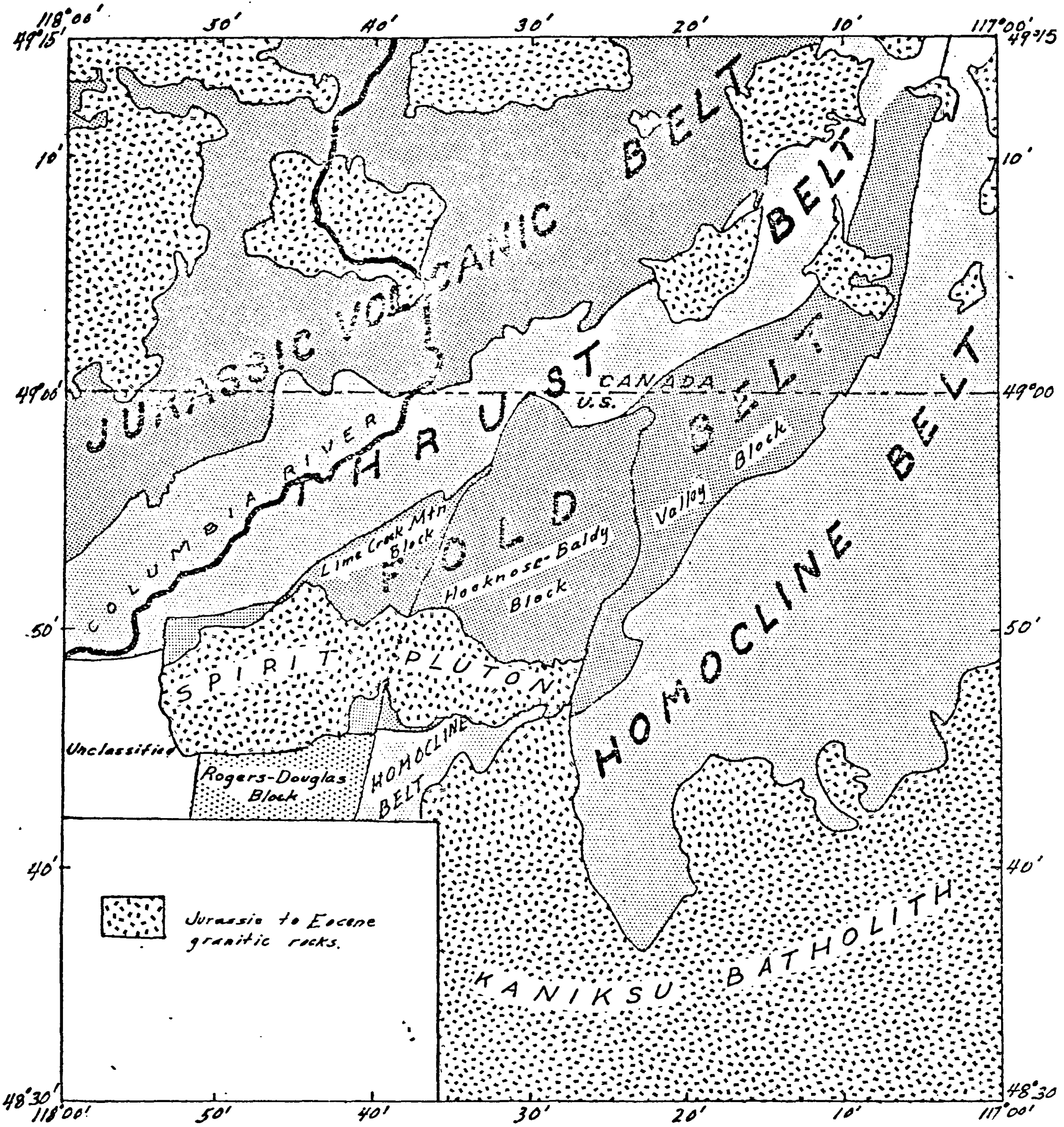


Figure 21. -- Index map showing structural subdivisions of the regional geologic map.

The homocline belt

The homocline belt is a northwestward dipping sequence of Precambrian and Cambrian rocks. The rocks strike N.30°E., and mostly dip northwest between 30° and 60°, probably averaging about 45°. The belt is bounded on the west by the Slate Creek fault and its southern extension, the Blue Bucket fault, on the east probably by an undiscovered fault that separates the homocline from the southeastward dipping rocks of the Prichard Formation (Belt Supergroup) (not indicated on fig. 17). The homocline appears to be the northwest limb of a faulted anticline, the most westerly of the open folds that compose the Belt-Purcell anticlinorium.

Northeast of the Metaline quadrangle the homocline extends into the East Nelson map area of British Columbia (Rice, 1941, map 603); to the southwest it terminates against the Flume Creek fault and the Kaniksu Batholith; however, what is believed to be the faulted southwest segment of the homocline--called here the Colville segment--occupies the southeast quarter of the Colville 30 minute quadrangle, figure 1. The Colville segment ends against what may be the south projection of the Leadpoint fault and against the Spirit pluton on the north and a lobe of the Kaniksu Batholith on the south.

The regularity of the homocline is interrupted by the high angle Flume Creek fault that segments it, by northwest trending cross faults that offset its stratigraphic units, and by the Brodie-Sullivan kink fold, a cross buckle that extends S. 80°E. from Brodie Mountain on the east side of the Columbia River to east of Hall Mountain, a traceable distance of 48 kilometres across the three structural belts. The stratigraphic sequence of the homocline belt is broken by a northwest dipping strike fault, named the Harvey fault by Park and Cannon (1943, pl. 1).

The northwestern boundary of the homocline is the Slate Creek fault, which continues as a fault, as far north as the International Boundary but becomes a synclinal fold a short distance further on. This fault, originally interpreted by Park and Cannon (1943, p. 30) as a normal fault; has been reinterpreted by Dings and Whitebread (1965, p. 41) and McConnel and Anderson (1968, p. 1470) as a southeast dipping reverse fault. Interpretation as a reverse fault is more compatible with the regional structural pattern, as it permits the northeasterly trending Slate Creek fault to be a product of the same compressional forces that produced the northeast folds of the Kootenay arc. In this time context the Slate Creek fault becomes eligible for later deformation, that of the cross-folding event. It is, therefore, necessary to reinterpret the covered southwest extension of the Slate Creek fault, the Blue Bucket fault, as bent into a southeast hook where it crosses the Brodie-Sullivan kink fold. The continuation of the Slate Creek fault southwesterly beyond the kink fold is unidentified. Unless it dies out a short distance beyond the kink fold, it should reappear on the west side of the much younger Flume Creek fault, but is has not been recognized here.

Having lost the Slate Creek fault to the south in the confusion of the kink fold zone, a new western boundary for the homocline belt must be selected. For expediency, we are forced to accept the Flume Creek fault as the boundary, but only locally, because west of the fault and south of the east end of the Spirit pluton, what is believed to be the faulted segment of the homocline reappears, the Colville segment.

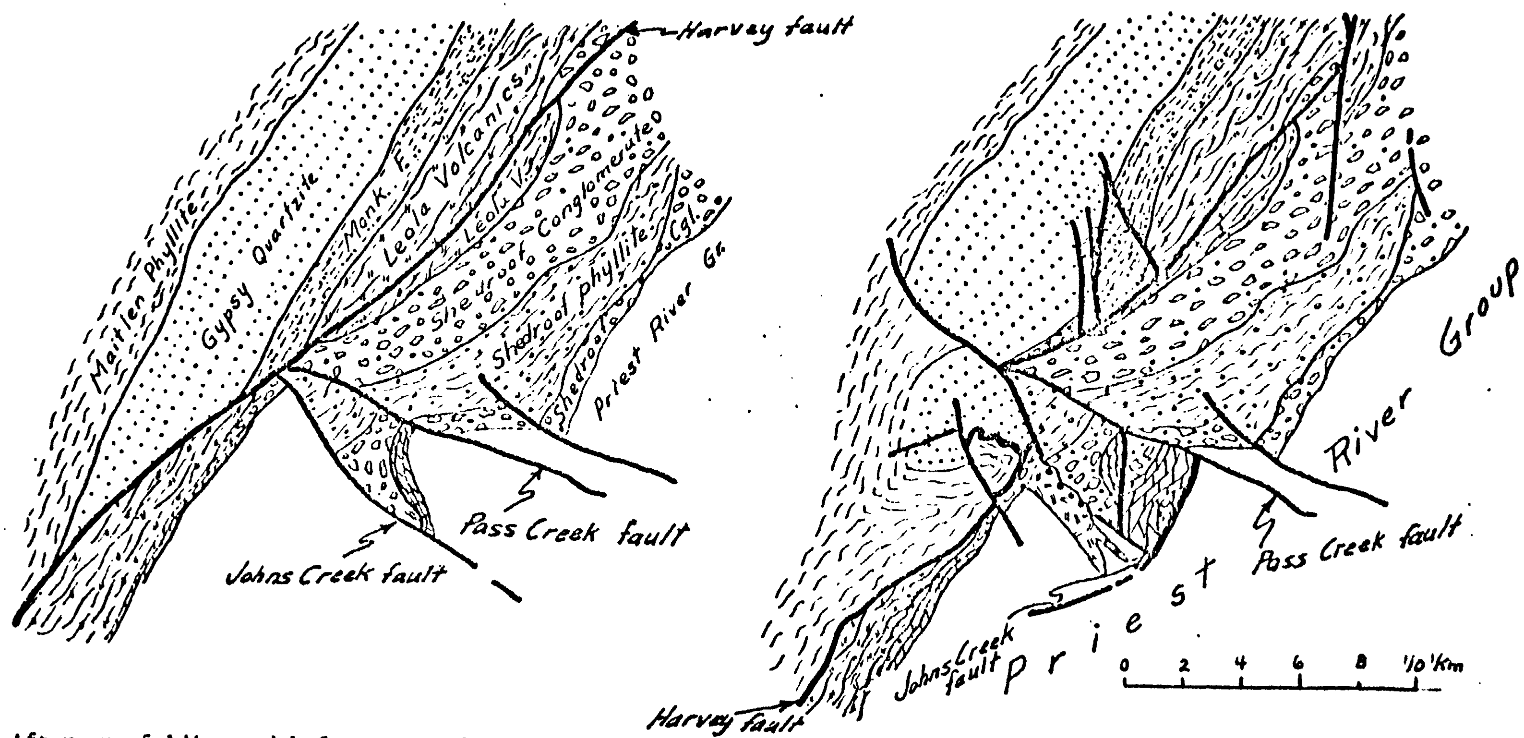
The rocks of the homocline belt have been subjected to at least four separate and distinct episodes of faulting, which are: (1) a Precambrian (Windermere) volcanic period, (2) a Late Triassic or Late Jurassic period of reverse faulting associated with the arc folds, (3) the episode of thrust and tear faulting that produced the kink fold, and (4) a period of extension faulting in the late Cretaceous and early Tertiary that produced faults of large vertical throw. Each of these episodes has left its impression on the homocline.

Precambrian faults are difficult to identify. In the area of Precambrian rocks in the Metaline quadrangle, only one fault can be demonstrated to have been active during the Precambrian--in this case, during the accumulation of the Shedroof Conglomerate. This is the Johns Creek fault. The Pass Creek fault is probably also Precambrian, but unqualified proof is not available. The Johns Creek fault, however, moved not only during the Precambrian but again during the late Cretaceous. The Precambrian movement produced apparent right lateral separation and the Cretaceous movement left lateral separation [1/]. Sometime between these two movements, the Harvey fault was formed and this fault was offset on the Johns Creek fault in a left lateral direction by the Cretaceous movement.

[1/] The terms, right and left lateral, are used without inferring strike-slip movement.

Before considering the character of the Harvey fault, the case for Precambrian faulting should be introduced. From a study of Park and Cannon's geologic map of the Metaline quadrangle (1943, pl. 1) I reinterpreted the Pass Creek fault as a Precambrian fault (Yates, 1970, p. 31) that had an apparent right lateral offset of about 6 1/2 kilometres, an offset measured on the contact between the Priest River Group and the Shedroof Conglomerate. I noted on Park and Cannon's map (1943, pl. 1) that the trend of the contact between the Shedroof Conglomerate and the younger Leola Volcanics projected across the fault zone without offset, which indicated that the Leola Volcanics were younger than the Pass Creek fault (fig. 22).

At the time this interpretation was made it was believed that the Johns Creek fault was strictly a postfold fault. Later I was able to examine numerous recently made roadcuts and found that the rocks shown as Monk Formation along Johns Creek belong to the phyllite phase of Shedroof Conglomerate. This correction in rock distribution and identification of the northeast extension of the Harvey Creek fault on Sullivan Creek, northeast of the Johns Creek fault, made it possible to reconstruct the following history of faulting in the Pass Creek area: (1) deposition of the Shedroof Conglomerate, (2) development of the Johns Creek-Pass Creek fault zone, with the Priest River-Shedroof contact on the southwest side of the zone moving upwards or northwestward, (3) development of an erosional unconformity in the Shedroof Conglomerate, at least on the southwest side of the fault zone, either after or contemporaneous with the fault movement, (4) deposition of the Leola Volcanics and pre-Mesozoic rocks on this unconformity, (5) folding of the Precambrian and Paleozoic rocks along a northeastly axis (arc folds), and production of the Harvey fault. At this stage the Shedroof-Priest



After arc folding and before cross folding

After cross folding

Figure 22. -- Development of the Pass Creek fault zone.

River contact is offset more than 10 kilometres in a right lateral direction and the Harvey fault is unbroken, (6) imposition of the regional stress that produced the kink fold, which reactivated the Johns Creek fault producing an apparent left lateral separation on both the Harvey fault and the Gypsy-Maitlen contact, and (7) emplacement of granitic magma along the Johns Creek fault.

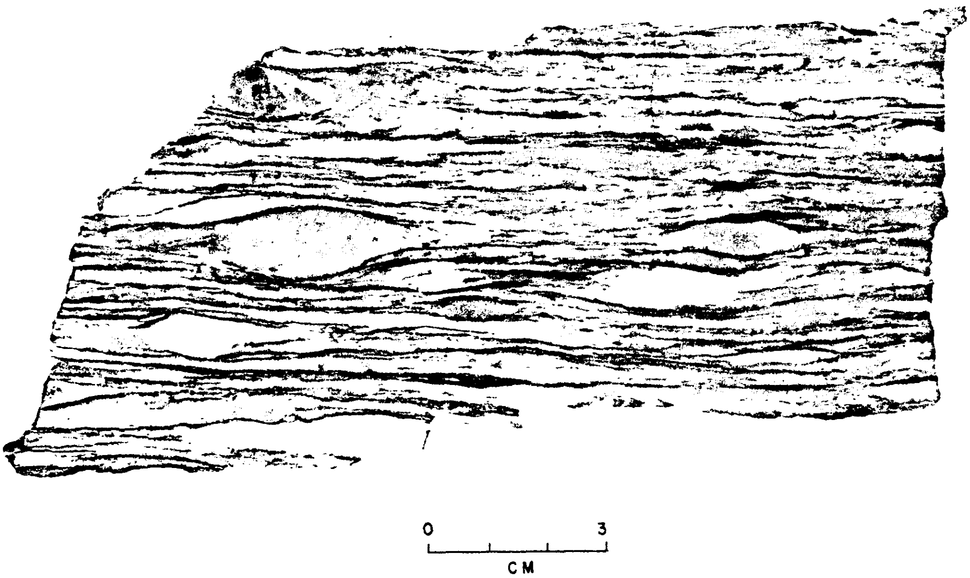
In the above structural history, the Harvey fault is considered a reverse fault related to the arc folding; this is a deduction and not a proven fact. On the basis of its northwest dip--determined from topographic expression--Park and Cannon (1943, p. 33) pronounced it a normal fault, down on the northwest. Because both beds and fault dip in the same direction in unknown angular relation, the movement direction of the fault is open to interpretation. With the angular relations unknown, the Harvey fault can be a reverse fault related to the folding, a prefold décollement thrust, or a normal fault older than the cross folding. If the dip of the fault is less than the dip of the beds, it is a reverse fault; if greater than the dip of the beds, a normal fault; if only a little greater or a little less, possibly a décollement fault. Although décollement movement is conceivable, no other relations support this interpretation. Because the trend of the fault parallels that of the Slate Creek fault, which is a reverse fault, and not that of any known major normal fault, the reverse fault interpretation is preferred and the Harvey fault is provisionally given contemporanity with the arc folding.

The age of the faults in the vicinity of Leola Peak is uncertain. They are surely post-Precambrian, probably post-arc folding, most probably of crossfold age, but not unlikely of Tertiary age (extension faulting).

Of yet greater uncertainty, is the degree of deformation the folding of the homocline represents. Although the homocline doubtless is the limb of a broken fold, the fold certainly is not a simple arcfold with flexural slip the dominant process; the strongly cleaved rocks, the boudinage, the distended clasts, all suggest that shear folding was important, particularly in the gross units where there is essentially no physical difference between adjacent beds and where beds of contrasting strength are of very limited continuity. In the Shedroof Conglomerate shears are prominent: they are the planar structure of the phyllite, which, however, is not a cataclastic product of shearing, but a mud whose grain size has increased through metamorphism instead of having decreased through cataclasis. In the diamictite, where clasts are widely spaced, shears wrap around clasts and the clasts are pulled apart but not ground up. But in closely packed conglomerate, shearing has produced granulation of clasts, as well as attenuation and boudinage (see fig. 23). Where clasts were buttressed against clasts they were granulated; where matrixed by fluid phyllite, they were extended and "pulled" apart--but not granulated.

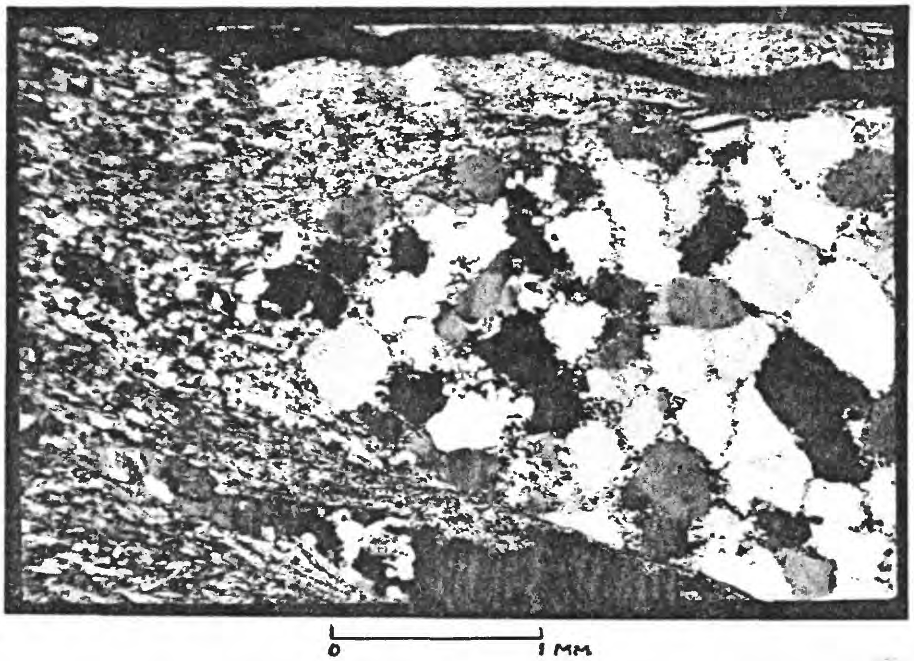
The penetrative deformation of the rocks of the homoclinal belt is clearly anomalous with the concept that the homocline is the limb of an open fold similar to those of the Belt-Purcell anticlinorium. A much more reasonable hypothesis is that it is the northwest limb of a faulted isoclinal anticline and that as a result the homocline has been thrust a considerable distance southeastward.

Figure 23. Photomicrograph and photograph of deformed Shedroof Conglomerate.



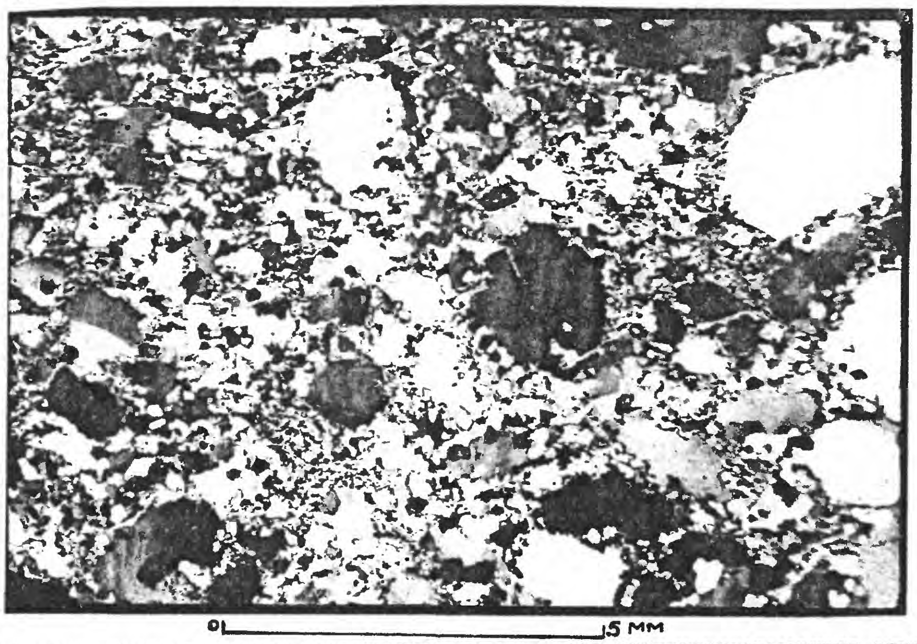
A

Figure 23. -- Photograph and photomicrographs of deformed Shedroof Conglomerate. A. Sawed slab of conglomerate showing stretched clasts of quartzite and limestone.



B

B. End of clast showing increase in granulation towards terminus. Note how crushed and recrystallized quartz surrounds each grain in the quartzite.



C

C. Detail of extremely crushed -- and recrystalized -- quartz.

(Figure 23, continuied)

The Fold Belt

The Fold Belt is the heart of the Kootenay arc: its northeastern fold trends define the arc. It is divided by transverse faults into several tectonic blocks, which are (fig. 21): (1) Valley block, (2) Hooknose-Baldy block, (3) Lime Creek Mountain block and (4) Rogers-Douglas block. North of the International Boundary in British Columbia, the Fold Belt (called the Eastern Belt by Fyles and Hewlett, 1959, p. 16) consists of multiple folds that coalesce at the boundary in the Sheep Creek anticline and east flanking Laib syncline. The anticline dominates the structure of the Valley, Hooknose-Baldy and Rogers-Douglas blocks. The Lime Creek Mountain block is the overturned limb of a syncline lying northwest of the anticline. Figure 21 shows the location and bounding faults of the several blocks.

Valley block

The Valley block, which includes the Metaline lead-zinc district, is the name--taken from the valley of the Pend Oreille River--Park and Cannon (1943, p. 28) gave to the structural unit that occupies the topographic low that extends from the International Boundary south to the Kanitsu Batholith. These writers bounded it on the west by the Flume Creek and Russian Creek faults and arbitrarily on the east by the top of the Maitlen Phyllite. The block is here expanded by extension into British Columbia and rebounded as follows: (1) on the west by the Flume Creek-Day-Black Bluff faults and (2) on the east, by the Slate Creek fault (in Washington), and by the axis of the Laib syncline in British Columbia.

The Sheep Creek anticline extends the entire length of the Valley block and is the longest, best preserved segment of a fold in the southern Kootenay arc. In British Columbia the anticline is a tight fold and according to Little (1960, p. 105), "is overturned to the west, and the axial plane strikes N. 10° to 20°E. and dips 65° to 70°E. The axis plunges gently, locally north but mainly south. In the vicinity of Lost Mountain (6 1/2 kilometres [4 miles] north of the Boundary) the Sheep Creek anticline is still overturned to the west, and the axis plunges about 10°S; the strike of the axial plane is N.30°E. South of the Ripple Creek fault the plunge increases to 25° and the anticline there is neither isoclinal nor overturned." As the fold continues southwestward, it broadens in width and increases its westward swing so that it crosses the International Boundary with a N.50°E. axial trend. South of the boundary the fold, called the Boundary anticline by Dings and Whitebread (1965, p. 34), continues this trend for a few miles then bends to S.20°W. a trend that extends to the vicinity of Metaline Falls, south of here it is lost in the mosaic of faulted blocks of Metaline Formation and Ledbetter Slate. From the boundary south to near the junction of Slate Creek and the Pend Oreille River the plunge is southwestward; from the junction to Metaline Falls the plunge reverses northeastward and this part of the fold is known as the Grandview anticline, the ore structure of the Pend Oreille and Grandview mines.

The greatest change in the Sheep Creek anticline occurs a few miles north of the International Boundary at the Ripple Creek fault, which strikes N.70°W. and bisects the anticline. South of the fault the fold is open; north of the fault it is tight, overturned on its west limb, and accompanied by several other folds. The exact nature of the movement on the Ripple Creek fault is controversial (Fyles and Hewlett, 1959, p. 62); however, regardless of the movement pattern, the exposed fold north of the fault is from a deeper level than the part to the south; the increase in depth by plunge of the structure was accented by movement on the fault. North of the fault the Sheep Creek anticline is flanked on the west by tight syncline and the western anticline (Little, 1960, p. 105) and on the east by the Laib Creek syncline, whose eastern limb belongs to the homocline belt.

Detailed mapping in the Metaline quadrangle by Dings and Whitebread (1964, pl. 1) shows the Boundary Sheep Creek anticline as a badly broken structure, fragmented by faults, most northeasterly in trend. The northeasterly faults are high angle faults; a few have displacements more than 300 metres, most are a hundred or more. Dips--observed or interpreted from relations between trace and topography--of these high angle faults are poor criteria of whether the faults are extension or contraction structures or whether they are related to the arc folding of the Mesozoic or to the extension faulting of the Tertiary. The northeast faults offset the Lee thrust fault (near the junction of Slate Creek and the Pend Oreille River) but the relative age of the thrust faults in the Valley block is debatable: Park and Cannon (1943, p. 30) relate the thrusts to the graben faulting (Tertiary extension faulting of this report), Dings and Whitebread (1965, p. 45-46) relate them to the arc folding, and the writer relates them to the cross folding, which comes after the arc folding and before the Tertiary extension faulting. Although the northeast faults offset the Lee thrust the displacement on the thrust is much less than that on the contact between the middle and upper members of the Metaline Formation. I interpret this to indicate that the northeast faults predate the cross folding and are contemporaneous with the arc folding, have been only mildly rejuvinated by the Tertiary extension faulting.

A few faults more northerly in trend that cross the northeasterly faults were most likely formed during the Tertiary extension. These, also high angle faults, clearly cross-cut the northeast fold trends, and approximate the trend of what might be their controlling structure, the Flume Creek fault. The Styx Creek fault, the most prominent of the north trending faults, has a displacement of about 1500 metres and is down on the east; displacement is the same movement sense as that of the Flume Creek fault.

In the Metaline quadrangle the western boundary of the Valley block is the Flume Creek fault. This fault, which has an overall strike of about N.10°E. is nowhere exposed, either on the surface or in mine workings. The only information on its dip is from three drill holes on Flume Creek northwest of Metaline Falls. From these holes the dip is determined as falling somewhere between 40° and 85° to the east. Dings and Whitebread (1965, p. 40) favor the steeper dip. The course of the fault as originally portrayed by Park and Cannon (1943, pl. 1), is moderately sinuous, but the sparcity of outcrops projecting through the glacial cover permits considerable freedom in interpretation. With very little shifting of the location of the fault under the cover, Dings and Whitebread (1963, pl. 1) changed the arcuate pattern of Park and Cannon to an angular pattern. Some changes were necessitated by the finding of new outcrops.

Although the angle of dip of the Flume Creek fault may be in question, the direction of displacement is not. Relative displacement is down to the east; or perhaps more properly, the west side moved up farther than the east side. In this movement sense, both structure and rocks can be matched across the fault. The amount of vertical displacement has been estimated by Park and Cannon (1943, p. 29) to be at least 3000 metres, by Dings and Whitebread (1965, p. 40) to be at least 3600 metres and interpreted by the writer to be at least 6000 metres and possibly 7500 metres.

The age of the movement on the Flume Creek fault is less than 100 million years, the age of the Spirit pluton. Possibly the post-pluton movement was preceded by pre-pluton movement, possibly during the cross fold event.

The Flume Creek fault makes an abrupt turn to the left at its northern end and becomes what is known as the Russian Creek fault. In this part of the Valley block fault movements are complicated and not thoroughly understood. The Russian Creek fault is a complex, poorly exposed structure that can be traced a long distance to the southwest. It has a history of two and possibly three different periods and kinds of fault movement, which are discussed in the description of the Hooknose-Baldy block.

The west boundary of the Valley block continues about 2 1/2 kilometres along the Russian Creek fault to the junction with the Day fault, a fault of northeasterly trend that continues into Canada to join the Black Bluff fault of Fyles and Hewlett (1959, p. 58-59-B). To join the Day fault to the Black Bluff fault, I took the liberty of projecting it through the Active Formation where no fault was shown by Fyles and Hewlett (1959, fig. 3, p. 162). The license for this projection comes from Fyles and Hewlett's statement (1959, p. 61) "Extreme contortion and shearing of the Active argillite on the north side of the river and along the Nelway-Waneta Highway make it difficult to locate a single fault zone even though the rocks are well exposed." As a corollary to this reinterpretation, the prong of faulted Active argillite that lies along the Pend Oreille River just north of the Boundary (see fig. 14), is interpreted as belonging to the lower plate of a thrust fault that lies under the Nelway (Metaline) Formation, the upper plate. The Day fault, so projected joins the Argillite fault 1 kilometre west of

Shenango Canyon; the Argillite fault joins the Black Bluff fault at Shenago Canyon in poorly understood relations (Fyles and Hewlett, 1959, p. 58-59). The west boundary of the Valley block continues along the Black Bluff fault to the north end of the map area (fig. 14). From the junction of the Day fault with the Argillite fault the western boundary of the Valley block is also the boundary between the miogeosynclinal facies and the Argillite facies.

In British Columbia, faults that form the western boundary of the Valley block are of two kinds: the Day fault is classified by Fyles and Hewlett as a transverse fault and the Black Bluff fault as an east dipping thrust fault. South of the boundary, the Day fault is regarded by Dings and Whitebread (1965, p. 43) as a near vertical fault of at least 1500 feet of throw with the northwest side down, however, the part of the fault that extends north of the Boundary is considered by Fyles and Hewlett (1959, p. 61) as "probably dipping steeply" and shown in cross-section (Sec. CC, fig. 2) as up on the northwest side. They also say (p. 58) that the Day fault is possibly the southwest continuation of the Salmo Valley fault and that the Black Bluff fault (a southeast dipping thrust) is under the Nelway Formation southeast of the Day fault.

The Black Bluff fault according to Fyles and Hewlett (1959, p. 55) is one of three regional thrust faults within the Salmo map area that are "old structures which probably originated during the primary folding and on which movement continued during the period of secondary folding." They describe it on page 59 as a fault whose "fault plane is irregular, locally nearly vertical, and elsewhere dips more gently to the southwest." I accept Fyles and Hewlett's interpretation of the Day fault as the southern end of the south Salmo fault, being a high angle fault, up on the northwest side and the Black Bluff fault as a southeast dipping thrust fault. However, I do not, as they do, relate the Black Bluff thrust, in time and origin, to the arc folding, but prefer to consider it a pre-fold thrust, reactivated during the folding. I would also consider it terminating a little west of the Black Bluff fault and would interpret the thrust lying beneath the Nelway Formation postulated by Fyles and Hewlett (1959, p. 58) as not the continuation of the Black Bluff fault, but an independent thrust related to the cross fold event.

Hooknose-Baldy block

The Hooknose-Baldy block lies west of the Flume Creek fault, south of the fault complex in the Russian Creek area, east of the Leadpoint fault, and north of the Magma and Meadow Creek faults and Spirit pluton. The rocks in the block range in age from the Precambrian Monk Formation to the Ordovician Ledbetter Slate. They are folded into the Hooknose anticline, which dominates the block, a subsidiary syncline to the southeast, and a badly broken northwest flank, complicated by a prefold detachment thrust fault. The Brodie-Sullivan kinkfold crosses the block just north of the Spirit pluton.

The Hooknose-Baldy anticline is believed to be an upfaulted segment of the Sheep Creek anticline (Boundary Anticline of Dings and Whitebread) and the southeast flanking syncline is believed to be the faulted syncline in Slate Creek. Park and Cannon (1943, p. 29) describe the anticline as follows: "The axial plane of this fold and the axial planes of the small satellitic folds dip about 80° SE." The writer suspects that the axial plane may have a dip of less than 80° and greater asymmetry than indicated. The anticline is broken by the northeast trending Ridge fault, which crosses the fold axis at a small angle, resulting in what appears to be two en echelon fold axes, the southwest axis stepped to the south. The Gypsy

Quartzite and Monk Formation that outcrops along the Flume Creek fault was regarded by Park and Cannon (1943, p. 32) as forming a faulted, southwest plunging anticline. This structure is now recognized as the Brodie-Sullivan kinkfold. On Park and Cannon's geologic map (1943, pl. 1) the flanking syncline is decidedly asymmetric, which suggests that the Ridge fault has either a large displacement, down on the southeast, or the syncline is a crenulated fold. It is most probably the latter.

The Spirit pluton effectively terminates the Hooknose-Baldy block as well as the Lime Creek Mountain block to the west. Before emplacement of the pluton, an east-west fault was the south boundary of these two blocks. The east end of this fault was engulfed by the pluton; the west end is the Magma fault, which is offset northward into the pluton by the northeast trending Meadow fault. The Magma fault is a north dipping fault of uncertain movement direction. Because it separates two different stratigraphic assemblages, it probably is a fault of large displacement.

The northern boundary of the Hooknose-Baldy block is the Russian Creek fault. The only stratigraphic unit found on both sides of the fault is the intraformational dolomite breccia unit of the Metaline Formation. The argillites that are the predominant rocks north of the fault are absent to the south; they belong to the Black Argillite belt of Fyles and Hewlett (1959, p. 48).

The Russian Creek fault has long been an enigma: the direction of dip, the extension or contractional nature of the fault, and even the reality of the hook connection with the Flume Creek fault are points of disagreement. The disagreement comes largely from the high degree of freedom of interpretation that results from the paucity of natural exposures in critical areas; consequently as artificial exposures increase, interpretations change and a

Figure 25.--Geologic cross section showing development of structure in the Kootenay arc.

A. Palinspastic section across the Kootenay arc as interpreted for the late Pennsylvanian. Dotted line that locates the décollement infers post-Pennsylvanian thrusting. To accomodate late Paleozoic thrusting the section can be adjusted by removing the appropriate younger rocks.

B. Gentle folding and westward tilting of section A. Dotted line indicates location of the décollement.

C. Section B after westward movement of the décollement.

D. Section A-B-C-D-E-F as indicated on figure 14. This section C after the Mesozoic folding and Tertiary faulting. The pre-Jurassic unconformity shown in the rocks at the west end of the section would have removed much of the rocks shown above the topographic profile to the east.

more complex tectonic history is postulated. The single straight-coursed Russian Creek fault of Park and Cannon (1943, pl. 1) is shown on figure 14 of this report as broken by three northwest striking high angle faults and as having a northeast striking high angle fault and a thrust fault terminating against it on the north side (see fig. 24). The current interpretation is that all postfold deformations that are recognized took part in producing this combination of faults. The sequence of events identified by the faults is partly inferential, and comes mainly from fault offsets and terminations.

The history of faulting in the Russian Creek area begins with the cross fold event and terminates with the extension faulting that postdates the 100 m.y. old Spirit pluton. The youngest faults belong to the set of northwest striking high angle faults that offset all other faults, including both the Russian Creek and Flume Creek faults. They probably belong to the extension faulting of the Eocene. In terms of offset relations the Russian Creek fault is next youngest; it is only offset by the northwest faults.

The Russian Creek fault is a high angle fault that has been interpreted in two different ways. Park and Cannon (1943, p. 28-30) believed it dips steeply to the north but Dings and Whitebread (1965, pl. 1, sec. BB') show it as a steep, south-dipping reverse fault, which flattens to a thrust at depth. They interpret it as produced in two stages (Ding and Whitebread, 1965, p. 47-48): (1) Flume Creek and Russian Creek faults, originated as a single, steeply-dipping normal fault, down to the east and north, and (2) later modification, "compressive stresses acting in a general north-south directions moved the block west of the Flume Creek fault northward and upward. The original steep northerly dip was overturned and the fault broken into a series of slice faults as postulated in Sec. BB', plate 1." The compressive

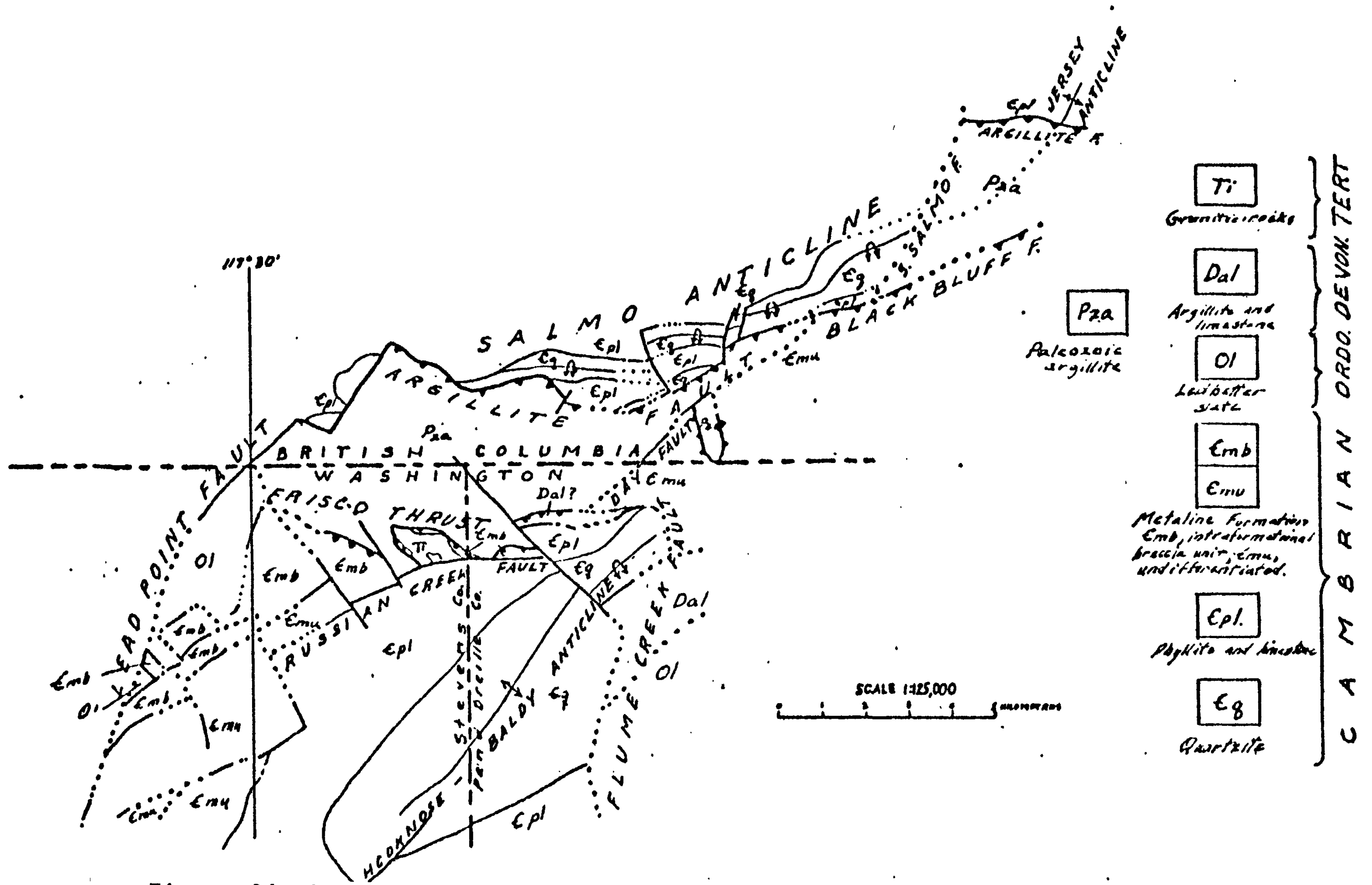


Figure 24--Structural setting of the Russian Creek fault. Formational boundaries and faults projected through areas covered by unconsolidated deposits

stresses were introduced to explain the folded, sheared, and recrystallized upper limestone member of the Metaline Formation that occurs north of the eastern segment of the fault which they did not believe could have been produced by normal fault.

The writer agrees with Dings and Whitebread's conclusion that the limestone is cataclastically metamorphosed, and that this could have resulted from northward movement of the Hooknose-Baldy block, but he believes the cataclasis more probably results from compression during the cross fold event--and before the extensional movement. The reason for this belief, is that structures in the Northport quadrangle (Yates, 1971), particularly in the Columbia anticline, strongly resemble, those at the north end of the Hooknose-Baldy block: the Northport structues are type examples of the cross fold event. By this comparison, the Russian Creek fault becomes a contractional or reverse, fault that dips southward, and the ancestral Flume Creek fault a tear fault, more or less vertical. This is mechanically the same model that Dings and Whitebread envision, but to make it operate they infer post-Spirit pluton compression after the high angle normal (extensional) faulting--an event the writer fails to recognize.

On the other hand, to produce the structure solely as a product of cross fold time is attended by difficulties. The timing is wrong; the Flume Creek fault cuts the Spirit pluton, which is clearly younger than the kink fold, the outstanding product of the cross fold event. Post-Spirit pluton movement on the Flume Creek fault is a necessity. The question is, "how much, if any, movement occurred before the extension, high angle, 'normal' faulting?" However, it is not essential, or even desirable to postulate compressional movement on the Russian Creek fault to account for the recrystallization of the upper member of the Metaline Formation. Deformation of the Metaline rocks could be related to movement on the Argillite or Frisco thrust faults, low angle faults, that belong to the cross fold event, and entirely unrelated to the Flume Creek-Russina Creek fault.

The favored interpretation is that the Frisco thrust fault and the deformation of the Metaline limestone were produced in the cross fold event and that the uplift of the Hooknose-Baldy block was considerably later, during the time of Cenozoic extensional faulting. This is a return to the interpretation of Park and Cannon (1943, p. 28-30) who favored both Flume Creek and Russian Creek faults as normal faults down on the east and north, respectively.

From the above, it is concluded that the Russian Creek fault dips at a high angle as does the north part of the Flume Creek fault. There is little doubt that these two faults are in reality a single structure and that the change in trend is not the bending of a once straight fault. Accordingly, any interpretation concerning the form or movement of one part must provide for that of the other part.

The Russian Creek fault terminates the Frisco thrust fault and the Day fault. The Frisco thrust fault, a thrust formed during the cross fold event moved the intraformational breccia unit of the Leadpoint section of the Metaline Formation over the argillite of Russian Creek. The displacement on this fault is unknown; it could be great. The Day fault, a high angle fault, is discussed above.

The Russian Creek fault, following a course offset by several northwest trending faults, extends southwestward to end against the Leadpoint fault, the western boundary of the Hooknose-Baldy block. The Leadpoint fault is a high angle fault that is exposed in only one place. The fault, as it is in the Deep Creek area, is described on pages 112-114, without mention of its possible northern and southern extensions. As shown on map I-412 (Yates, 1964), the fault is projected to the Canadian Boundary at the northeast corner of the Deep Creek quadrangle. It is possible, however, that it does not cross the boundary, but joins with the Columbia fault--strands of which cross the north slope of Red Top Mountain. The implementation of this suggestion requires later extension on the Leadpoint fault to explain the anamalous drag of Gypsy Quartzite near the junction with the Columbia fault.

The northward extension of the Leadpoint fault from the above-mentioned point of junction, would intersect the Argillite fault in British Columbia at its west end where it would separate the Active Formation (Pza on fig. 14) from the "sedimentary rocks, correlation unknown" (Ca on fig. 14) of Fyles and Hewlett (1959, fig. 3 sheet A). It is conceivable, but unlikely, that the Argillite fault and the Columbia fault are segments of the same fault offset by the Leadpoint fault.

The presence of a large décollement thrust in northeastern Washington is demanded by the contrasts in stratigraphic sections between adjacent tectonic blocks. Most demanding are the contrasts between the Hooknose-Baldy and Lime Creek Mountain blocks.

Structurally, the Lime Creek Mountain block is the overturned limb of a fold. The significant features of the block are not the internal structures, however, but the relations of the rocks in the overturned limb to those in neighboring blocks. That these relations are anomalous is recorded in the stratigraphic descriptions of this report but it has not been emphasized that the stratigraphic sections permitting comparison are of the Cambrian miogeosynclinal facies. Because of restriction to one facies, differences between sections are neither extreme nor everywhere obvious; eugeosynclinal rocks do not rest against miogeosynclinal rocks as they do in the thrust sheets in the Great Basin, nor is it necessary to invoke tectonic transport measured in scores of kilometres to explain the juxtapositions. Differences between the lower Paleozoic part of the stratigraphic sections in the Hooknose-Baldy, Lime Creek Mountain, and Columbia anticline--which represent a cross-section of the Kootenay arc--are presented in table 5.

The distinctions are in the substitution of one lithologic unit for another, modification of lithologies by addition of subunits, the absence of units, and the differences in boundaries between units.

A careful analysis of differences and similarities between these sections, lead to the conclusion that no reasonable geologic model would explain the rock distribution unless it was constructed under the premise that not all the rocks were deposited in their present locations. Once this is accepted, it becomes a question of deciding which section or sections of rocks are allochthonous and by what structural process they arrived at their present positions. In constructing such a model, the possibilities of variation of stratigraphy by gradation were examined, and although found to be capable of explaining some differences, were unable to explain others because of the lack of space needed for the changes in lithologies.

		Hooknose-Baldy block Leadpoint section (below thrust fault)	Lime Creek Mountain block Metaline section	Columbia anticline Northport section
Relation to black shale facies		Conformable with black slate that contains Ordovician graptolites	Conformable with black slate that contains Ordovician graptolites	Fault contact with black argillite of uncertain age.
Metaline Formation	Upper unit	Mixed unit of calcareous argillite, intraformational breccia, and limestone similar to that of lower unit.	limestone 760 metres	limestone 600 metres
	Middle unit	Intraformational dolomite breccia ------ 400 metres bedded dolomite- 300 metres	bedded dolomite 1430 metres	thick bedded dolomite 600 metres
	Lower unit	Fine grained dark gray limestone, contains subunits of black dolomite 450 metres (fault)	Fine grained dark gray limestone ------------460 metres Phyllite subunit that has intercalations of the dark gray limestone ---300 metres	Medium to coarse-grained sugary limestone. 150 metres

Table 5. -- Contrasts in stratigraphy of the lower Paleozoic rocks in adjacent tectonic blocks.

The Leadpoint section in the Hooknose-Baldy block is obviously the most exotic, partly because it is located between two like sections, the Metaline and Lime Creek Mountain. This, of course, does not make it allochthonous; it only indicates that the Leadpoint section, or its neighbors, were not deposited in their present locations.

On the other hand, if we make the a priori conclusion that the Leadpoint section is allochthonous, we can examine the means by which it could have arrived at its present position. Juxtaposition could have resulted from transport by nappe folds, by strike-slip faults, or by thrust faults. The nappes would have to form during the arc folding; the strike-slip faults would have to be faults dividing blocks (Flume Creek and Leadpoint faults) and transport would be during the cross fold event; the thrust faults would have to be independent of the folding and would be décollement thrusts formed before the folding.

Nappe folding as a possible mechanism of transport can be readily eliminated. Nappe folds and associated thrusts are flat structures with a high proportion of overturned beds, which are not present in the southern part of the arc. Dips of both beds and faults average more than 45° and the axial planes of folds are steep. The possibility that nappe folds and thrusts were produced before the arc folding is without support. The re-folding of nappes would produce synforms and antiforms as well as anticlines and synclines.

Strike-slip faults as transport structures would have to be faults such as the Flume Creek and Leadpoint faults. No other faults transverse to the strike of the rocks could bring the sections together. The hypothesis can be tested by postulating that the Hooknose-Baldy block, containing the breccia unit of the Metaline Formation, was moved into position between the

Valley and Limestone Mountain blocks. The only other place the breccia unit is known is south of the Spirit pluton on the east side of South Fork fault. To juxtapose these units requires moving the Hooknose-Baldy block at least 32 kilometres southward and even with this displacement there is no real correlation of stratigraphy or structure across the fault. The geometry of the structure of this part of the arc is opposed to transport by strike slip faults.

With two of the three methods of tectonic transport eliminated, thrust faulting wins by default. The thrusts, however, must be folded thrusts, as there are no known post-fold thrusts capable of making the necessary redistribution of the rocks. A pre-fold décollement thrust fault will satisfy the demands prescribed the stratigraphic contrasts existing between adjacent fault bounded blocks.

A décollement thrust fault is difficult to recognize unless it juxtaposes extremely contrasting facies, such as miogeosynclinal rocks against eugeosynclinal. If the fault has been folded and later broken by high angle faults it is even less obvious. On the other hand, where such a fault is a hypothetical necessity, it is equally difficult to determine which mapped faults are the fragmented parts of a folded and faulted thrust. Any fault that is nearly parallel to the strike and dip of bedding becomes suspect; but unfortunately in this area such faults also could have formed as high angle reverse faults during the arc folding. A further complication is that strike faults that do not bound map units are almost impossible to trace on the ground. The décollement thrust indicated on figure 14 was not mapped on the ground as such, but was later identified by selecting those faults that parallel or are nearly parallel to bedding and permissively bound unlike rock assemblage. Post-thrust folding and faulting has exposed most of the fault to erosion and burial.

A model to explain the reorganization of the Cambrian sequence of the Deep Creek area by thrust faulting begins with a reconstruction of the sedimentary basin in which the Cambrian sediments were deposited. This reconstruction follows the concept that the early Paleozoic miogeosynclinal rocks were deposited in a basin located at, or near, the continental margin and that no western land mass contributed sediment to the basin. This basin was open towards the sea and any land to the west would be an island arc, lying too far away to supply sediments. If such a basin existed, there should have been--according to geosynclinal theory--a eugeosynclinal facies between the miogeosyncline and the island arc. This facies has not been recognized, but could lie buried beneath the younger Permian and Mesozoic rocks, which are of the eugeosynclinal facies.

Within a miogeosynclinal basin, most changes between contemporaneous lithologic facies are progressive and transitional across the basin and not repeated. Using this principle, we can assume there existed a systematic depositional pattern within the basin. It follows that the two most unlike "facies" or assemblages would have been deposited at parts of the basin nearest and farthest from shore.

Guided by the above restrictions, let us consider the arrangement of stratigraphic sections occurring in the Valley, Hooknose-Baldy, Lime Creek Mountain, and Columbia anticline blocks. The Leadpoint section, which occurs only in the Hooknose-Baldy block has three distinguishing features found in none of the other sections (see fig. 3 and table 5). These are: (1) the intraformational breccia unit that proxies for the upper limestone member of the Metaline Formation at its type section in the Metaline quadrangle, (2) beds of black dolomite in the lower limestone member of the Metaline Formation, and (3) a unit transitional between the carbonate rocks of the Metaline Formation and the black shale facies of the Ledbetter Slate. The section in the Columbia anticline to the west has a section of Metaline Formation that consists of the three basic units, limestone, dolomite, and limestone, the sequence common to all but the section of the Hooknose-Baldy block; however, the lower unit of this block has little resemblance to comparable units in the other sections. In addition, the phyllite of the Maitlen is much less abundant in the Columbia anticline. In contrast to the sections is the Hooknose-Baldy block and Columbia anticline the sections in the Valley (Metaline quadrangle) and Lime Creek Mountain blocks closely resemble each other; stratigraphically equivalent units are very similar in both sections. Following the premise of non-repetitive lithology, we can rearrange the sections so that the most different, the Columbia and Leadpoint, are the known "end members" of the basin samples. This gives us two choices: (1) the Columbia anticline section was either deposited as the seaward (western) section or (2) as the shoreward section. In either case, the Leadpoint section of the Hooknose-Baldy block must have been deposited in the other outside position.

If we regard the Columbia anticline section as authochthonous, and postulate that the Leadpoint section was deposited in the easternmost position, the rearrangement would be accomplished by thrusting the Leadpoint section westward and downfaulting it into position along the Leadpoint fault. Conversely, we can interpret the sections in the Columbia anticline, Lime Creek Mountain block, and Valley block as belonging to an allochthonous plate thrust eastward over the Leadpoint section that was later faulted up along the Flume Creek and Leadpoint faults. If we disregard the movement directions on the block-bounding faults, both movements are theoretically possible rearrangements.

If the Columbia anticline section is considered as deposited in the eastern part of the basin and the Leadpoint section to the west, a theoretical rearrangement can be made in two ways; one by thrusting eastward and the other by thrusting westward. However, both methods require two thrust faults in very illogical relations. It is, furthermore, highly illogical to place the Columbia section in the most eastern position next to the Metaline section, because by doing this, two unlike sections are placed side by side. The most striking and most important difference between these two sections is in the Reeves-Maitlen. In the Metaline type section, the Reeves is a single unit of limestone that separates quartzite from a great thickness of phyllite; in the Columbia section, the thick section of phyllite is absent, the quartzite is separated from the Metaline Formation by an alternation of phyllite and limestone units, with limestone far in excess of the phyllite. The postulate that the Columbia section came from the east beyond the Metaline section, is at odds with the easternmost Reeves in British Columbia, which is a single limestone units, and it is also at odds with the contrast between Columbia and Metaline sections. This objection is further expressed by the contrast between the lower members of the Metaline Formation. The distinctive lower member, shaly dolomitic limestone, which is present in the Valley, Hooknose-Baldy, and Lime Creek Mountain blocks is absent in the Columbia anticline section, being substituted for by a white bedded limestone. Because of these objections it is concluded that it is reasonable to postulate the most eastward location as the place of deposition of the section of the Columbia anticline. The Leadpoint section of the Hooknose-Baldy block belongs in this shoreward position.

The depositional model as now proposed has the rocks of the Columbia anticline deposited in the westward position, the Leadpoint section of the Hooknose-Baldy block deposited in an eastward position and the section in the Lime Creek Mountain block and that in the Metaline quadrangle deposited in the central positions. The question as to whether the Leadpoint section is autochthonous or allochthonous can now be answered. If the Leadpoint section is allochthonous, it would belong to the upper plate, thrust westward and downfaulted along the Leadpoint fault; if autochthonous, it would belong to the lower plate, and upfaulted along the Leadpoint fault into the eastward thrust upper plate. The eastward-thrust solution, however, is unworkable because relative displacement on the Leadpoint fault is down on the east--the Leadpoint section side--and up on the west, a displacement that satisfies only a westward thrust.

An interpretation of the structural character of the Kootenay arc is shown in the cross section (fig. 25). It would be incredible luck if this section was a facsimile of the actual structure; at best, it can show little more than the style of deformation. Because certain interpretations enter into its construction, as, for example, the direction of thinning and of facies changes and the correlation and age relations of some of the younger rocks, it is open to challenge, to modification, to reinterpretation; nevertheless, when prefold thrusting is accepted as essential to an explanation of the stratigraphic inconsistencies, the choice of a structural model is severely limited--once the course of the thrust is determined.

(page 305 follows)

The cross sections of figure 25 were constructed to explain the stratigraphic anomalies by the use of only one detachment fault. Another, and in some ways simpler, explanation makes use of two thrust faults. This is illustrated diagrammatically in figure 26. In this diagram the junction between the thrust fault of figure 25 and a second fault, the "sole" thrust, is just west of the Black Canyon fault. The Lime Creek Mountain block in this model is only autochthonous in respect to the overlying plate, which consists of the rocks of the Leadpoint section. The sole thrust moved rocks of the Metaline section eastward over the facies of the Columbia anticline, which are the only truly autochthonous rocks in the cross section.

Thrust belt

The introduction of the Columbia anticline into the preceding discussion of the detachment fault was, in a sense, an introduction to the Thrust Belt because the Columbia anticline is the Thrust Belt (fig. 14): the anticline dominates the structure of the belt. The Thrust Belt is a modified part of the Fold Belt; it likewise, is composed of rocks folded along arc trends. The thrust faults in the Thrust Belt are not as the name suggests, thrust that broke from over-extended folds of the arc; instead they cut across the arc folds, striking nearly east-west (between N.80°E and N.80°W). The Fold Belt and the Columbia anticline extends S.60°W. from the International Boundary to the western boundary of the map (fig. 14) and probably as far beyond as the Kettle River. To the northeast the Columbia anticline crosses the International Boundary trending N.75°E. to the South Salmo River and from there, N.20°E., following the trend of the arc and including the Mine Belt and Black Argillite Belt of Fyles and Hewlett (1959, fig. 2).

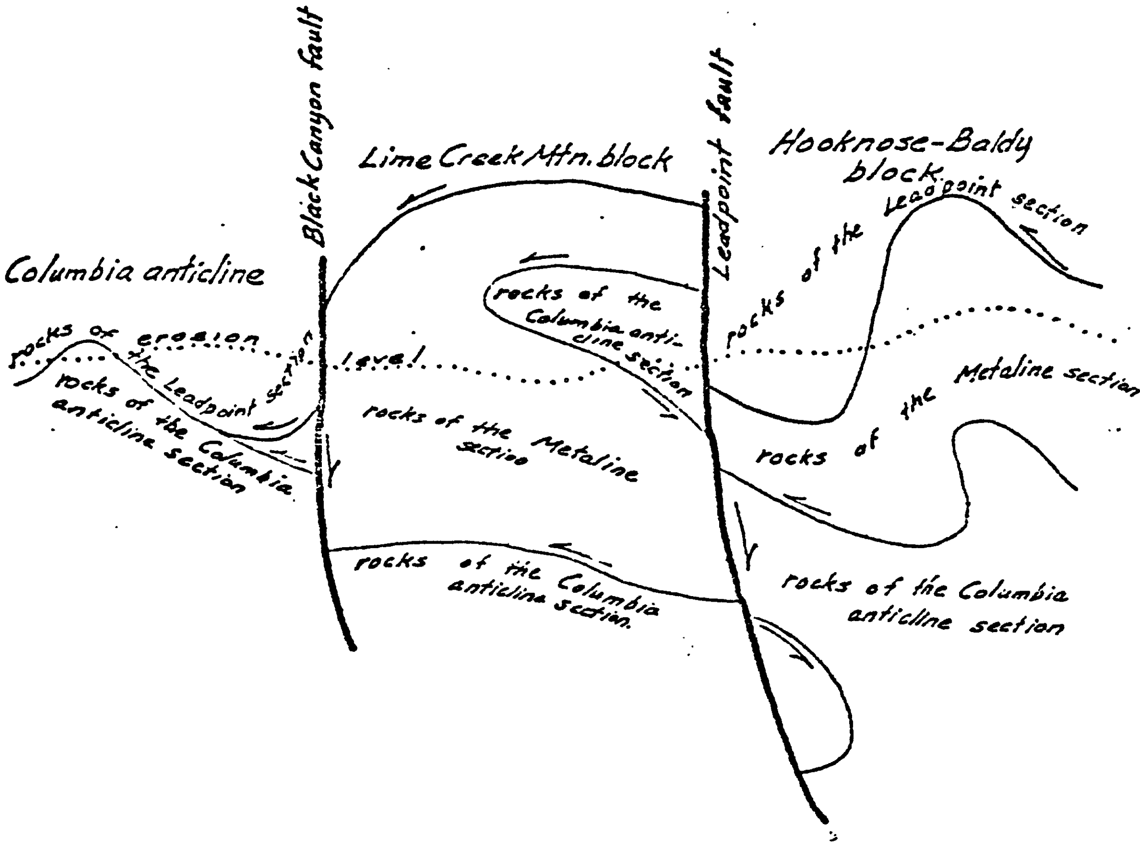


Figure 26.--Alternate interpretation of the pre-fold thrust fault of the Deep Creek area. Not to scale.

Two contrasting assemblages of rocks compose the thrust belt: the older, ranging from Gypsy Quartzite to the Metaline Formation, forms the core of the Columbia anticline; the younger assemblage, mainly unfossiliferous argillite of the transitional facies, forms the outer envelope of the fold. The two assemblages are separated almost everywhere by faults, either prefold thrusts or crossfold thrusts. Only on Red Top Mountain were the two groups seen in depositional contact. The assemblage of rocks in the Thrust Belt differs largely from those in the foldbelt by having much more argillite--in addition to differences between the Reeves-Metaline sequences.

Boundaries assigned to the thrust belt are more influenced by the distribution of argillite and Jurassic volcanic rocks than they are by the limits of thrust faults, mainly because the limits of lithologies can be fixed with greater precision than can the traces of thrusts in the argillites. The southeastern (and eastern) boundary of the thrust belt extends southwestward from the north edge of the map area of figure 14, following along the Black Bluff fault, Day fault, Russian Creek fault, Black Canyon fault, Ideal fault, and China Bend fault. Two general trends are represented by these bounding faults: (1) the northeasterly trend of the arc folds, and (2) the more easterly trend of the cross fold thrusts. The northwesten boundary follows the Waneta fault and back and forth along a jig-saw of unnamed faults that block off the Jurassic volcanics from the thrust belt.

The Columbia anticline is a very complex structure, complex because of post-fold deformation and not because of primary irregularity. It is only the core rocks of the anticline that bring out the anticlinal structure; the argillites, the envelope rocks, are too lacking in reliable stratigraphic units to permit the recognition of either folds or faults. In fact, neither in British Columbia nor in Washington has there been any correlation of argillites or phyllites on either side of the anticline. Fortunately, the core is composed of moderately thin, easily recognized, limestone, phyllite, and quartzite units.

The core rocks, however, are not continuously exposed along the length of the anticline, but appear on the geologic map (fig. 14) as islands in a sea of undifferentiated argillite. The gaps in the continuity of the core are not from simple alternations in the direction of plunge of the fold, but from cross folds and associated thrust faults transverse to the anticlinal trend. As a result, a once continuous anticline is now a series of discrete, disconnected folds.

Columbia anticline in British Columbia

Sufficiently great are the separations between the segments of the Columbia anticline in British Columbia to have been given separate names and treated as independent structures by Fyles and Hewlett (1959, p. 63-76) from whose work the following description is largely drawn. The core rocks of the anticline occur in what these authors term, the "mine belt," which contains all the major lead-zinc deposits of the Salmo area. The mine belt is one of four structural belts that compose the Salmo district. The envelope, or argillite rocks, of the Columbia anticline are in what they term "argillite belt." The division of the thrust belt into these two belts in the Salmo area is a logical and very useful division, but unfortunately it does not bring out the geologic relations south of the International Boundary, as well as the presentation of the structure as a faulted anticline in a belt characterized by thrust faults.

The segments of the Columbia anticline, Fyles and Hewlett (1959, p. 64, 68, and 73) identified by name are the Salmo River anticline, the Jersey anticline, and the Jack Pot anticline.

The Salmo River anticline is British Columbia's most southern segment of the Columbia anticline. It is in that part of the Kootenay arc having the most pronounced curvature and consequently the most easterly trend. The Salmo River anticline, trending N.75°E., has been traced for more than 11 kilometres, being terminated at its west end by northeast trending cross faults that bring the dark, fine grained clastics (ca of fig. 14) against the core rocks. At its east end it butts against the northerly trending Salmo River fault. Although an extremely complex structure, it is basically an overturned isoclinal fold whose axial plane dips 40° to 60° south. According to Fyles and Hewlett (1959, p. 65-66), "the symmetry of the Salmo River anticline is broken by bedding faults, some of which have been traced for more than two miles .....it is thought that the faults originated during the primary folding."

The Jersey anticline, which lies northeast of the Salmo River anticline, is also a complex isoclinal fold, but its axial trend is different, being N.15°E. The "axial plane dips steeply east on the east side of the Mine Belt and flattens to the west until the anticline is recumbent. The right side up eastern limb appears to have ridden over the overturned limb on a series of bedding faults," (Fyles and Hewlett, p. 68). The western part of the core is cut out by the Argillite fault, which separates the Argillite belt from the Mine Belt. The fold is bounded on its south end by an east trending segment of the Argillite fault and on its north end by the Hidden Creek stock.

The Jack Pot anticline is the most northerly segment of the Columbia anticline that is shown on figure 14. It lies in the Porcupine Creek area. This fold is a complex isoclinal anticline similar to the two described above. The axial plane of the anticline strikes a little east of north and dips from a little less than 30° to a little more than 45°.

The three core segments of the Columbia anticline in British Columbia are separated and surrounded by the envelope rocks of argillite and phyllite. The east and southeast flank of the Columbia anticline belongs to the Argillite belt of Fyles and Hewlett, where the structure is summarized as, "eastward and southeastward dipping cleavage and bedding." The west and northwest flank of the Columbia anticline is enveloped by phyllites and schists of the upper Laib Formation (the phyllite member of the Maitlen Formation). These rocks have small isoclinal folds, but no major structures have been defined. The rocks apparently reacted to folding the same as the structurally similar rocks of the Argillite belt.

Fyles and Hewlett (1959, p. 75) writing of the Argillite belt east of the Jersey and Jack Pot anticlines say, "regional considerations suggest that the Black Argillite Belt occupies the trough of a major syncline." Such a structure is now unrecognizable because of the internal deformation and lack of key stratigraphic units. In pursuance of this synclinal concept and in harmony with the isoclinal fold structure of the area, it would be consistent to interpret the two faults; that bound the belt, the Black Bluff fault and the Argillite fault, as one and the same, being an isoclinally folded pre-arc fold thrust fault.

The Columbia anticline in Washington

South of the International Boundary the core rocks of the Columbia anticline are exposed in three places: (1) Red Top Mountain-Cedar Lake area north of Leadpoint, (2) Northport area, on the Columbia River near Northport, and (3) China Bend area, on the Columbia River, southwest of the Northport area. The Red Top Mountain-Cedar Lake area is isolated by 8 km of argillite terrane from the Salmo anticline and isolated by similar rocks over a similar distance from the Northport area. The Northport and China Bend areas are almost continuous, being separated by only a cover of terrace gravel of no great extent.

The structures in the Red Top-Cedar Lake area are similar to the structures in the thrust belt in British Columbia, but those in the Northport and China Bend areas differ in two important respects. In contrast to the isoclinal character of the Salmo River and the Jersey anticlines, the Columbia anticline in the Northport and China Bend areas is an open structure, which is only locally overturned. The second difference is the absence of phase 2 folds (folds produced by refolding primary, phase 1 folds along the same axis). Refolding did take place along the Columbia River, but transverse to and not along, the axes of the arc folds.

The Red Top-Cedar Mountain area, is described on p. 100-103, where it is suggested that the open synclinal fold on Red Top Mountain could be a refolded isoclinal anticline like that west of Cedar Lake.

The Columbia anticline in the Northport area is difficult to recognize as an anticline, because it is so badly broken and distorted by cross-folding. From broken segments a stratigraphic sequence similar to that of the Red Top area can be roughly arranged. The oldest rock, Gypsy Quartzite, forms the interrupted core of the anticline. It crops out in three places: (1) on Deep Creek, about 2 1/2 km above the mouth where it is isolated from all other units by a thick cover of Pleistocene deposits; (2) on the east side of the Columbia River about 3 km southwest of Northport as an east-west and a N.35°E. segment of the Columbia anticline; and (3) on the west side of the river at China Bend. These three exposures were once part of an unbroken anticline, but during the cross fold event they were reoriented by movement along thrust faults and associated tear faults. In the process, some beds were overturned. The trace of the anticlinal axis and the faults that have segmented it are shown in figure 27.

The China Bend segment of the anticline is partly covered by terrace deposits, which makes structural interpretation difficult. The exposed Gypsy Quartzite, supposedly representing the crest and axis of the Columbia anticline, can be connected to the axis of the Northport segment by a N.60°E. projection, but this inferred change in axial trend may not be a bend but the result of an offset and rotation along the N.15°E. trending tear fault that separates the two segments and along which the China Bend segment moved upwards and northward. The axial region of the Columbia anticline shifts from a medial position east of Northport to northwest of median southwest of Northport, and to southeast of median in the China Bend segment.

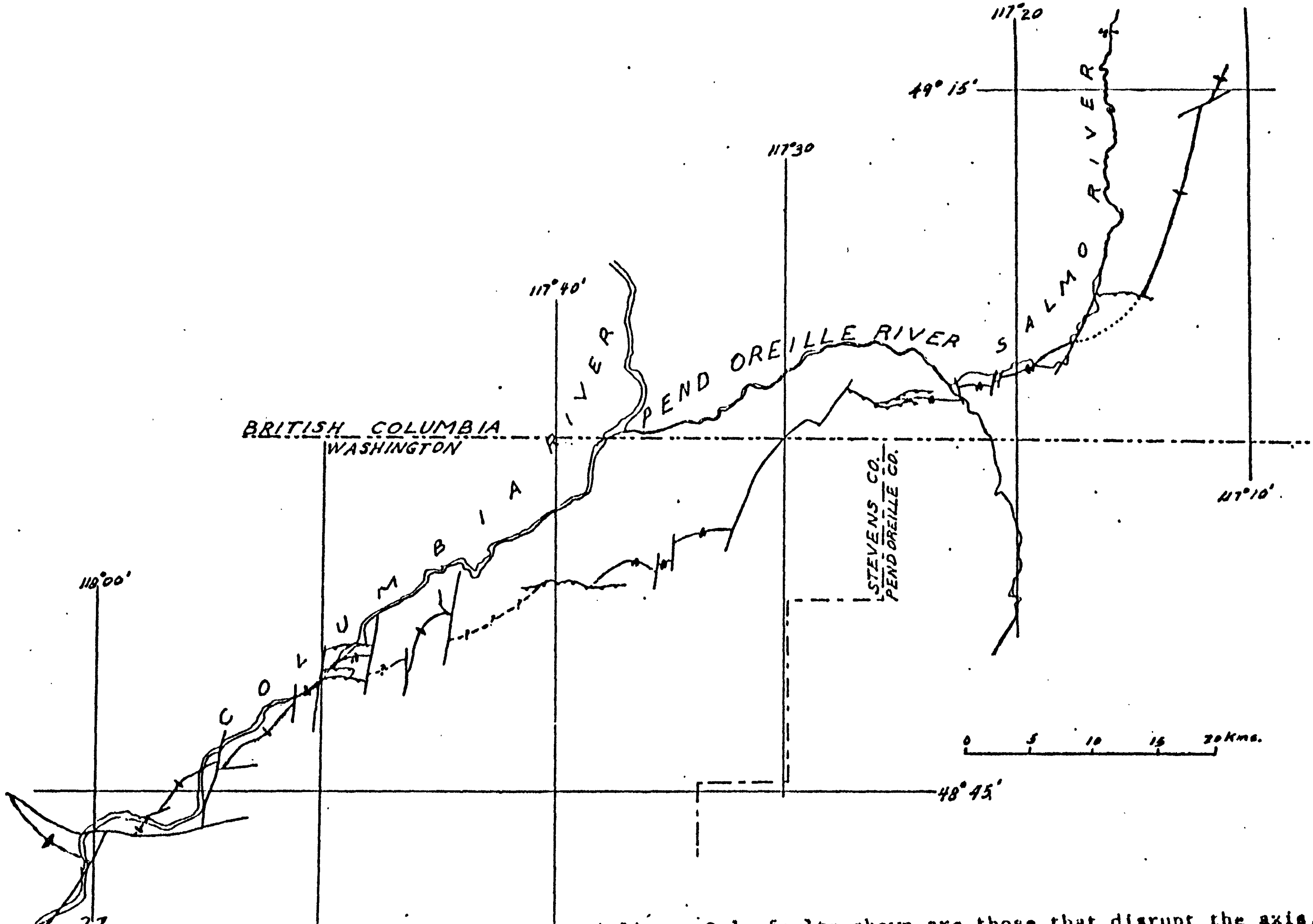


Figure 27 -- Trace of the axis of the Columbia anticline. Only faults shown are those that disrupt the axis.

The China Bend segment of the anticline is a mini-anticlinorium, being composed of numerous small shallow anticlines and synclines. The structure indicated on the map of the Northport quadrangle (Yates, 1971) and on figure 14 is a generalization of a even more complex structure. Of particular importance are the relations between the northeasterly strike of the stratigraphic units and the east-west trending faults. The China Bend segment terminates to the north against a south dipping east-west thrust fault. The northeast strike of the beds continues to the thrust without deviation--drag is absent. The only distorted rocks near the thrust are in a thrust slice of Flagstaff Mountain sequence rocks that lies north of the bounding thrust. Here an east-west striking, north dipping cleavage has developed. The reaction here to cross fold stress contrasts with that at the southern end of the segment, which is bounded by a thrust fault dipping south at a high angle. Here beds are bent or faulted into parallelism with the faults, and nearby structures become increasingly complex.

By the end of arc folding the Washington part of the Columbia anticline is interpreted to have been a moderately symmetrical anticline whose horizontal axis trended about N.30°E. During the cross-fold event, which affected all rocks from the arc at least as far west as the Cascade Range, the arc was shortened in a north-south direction. The shortening was along east-west trending (N.80°E. to N.80°W.) thrust faults and N.10°E. tear faults and to a lesser extent by folding along trends of the thrusts. Compression across the trend of the fold broke the anticline into a mosaic of fault blocks. The thrust and tear faults shown on the geologic map of the Northport quadrangle (Yates, 1971) are only the major faults; doubtless, detailed mapping would reveal many lesser faults with the same structural style. The most prominent example of folding is the bending of the anticlinal core southwest of Northport into an east-west trend. Folding is further expressed by kink bands in the fine-grained phyllites and schists and by a giant kink, the Brodie-Sullivan kink fold.

In describing structure of the cross fold event in the Columbia anticline, I have used terminology that may be confusing. Not all the faults of north to N.15°E. trend called tear faults are tear faults in the usual sense, faults restricted to the upper plate of a thrust fault; some faults cut both upper and lower plates. My usage is that of Hills (1963, p. 208) who applies the term to faults that tear across fold structures in a late or intermediate stage of the folding. As an example, he uses faults described Jean Goguel (1947, p. 31-33) from the region of the Vanige in the Subalpine Chain of southeastern France. The north striking faults in the Northport and Deep Creek quadrangles are of this type except here the rocks were more brittle than those in the Vanige and the folds have broken into thrusts.

Most post-arc fold structures are in the core rocks of the anticline, but enough extend into the argillite envelope to eliminate the possibility that the cross folds are older than the faults that separate the core rocks from the envelope rocks.

I relate the structures that deform the Columbia anticline to the event that produced the Waneta thrust fault of the Salmo lead-zinc district and the post-fold thrust faults near Metaline Falls. Folds and compressional faults of east-west trend are common among the more northwesterly fold trends west of the arc, as can be seen in figure 28, where the major fold trends are indicated in an area that extends from the arc to the Cascade Range. Few of the trend lines shown on figure 28 represent folds of two limbs; nevertheless, they do indicate the trend of folded rock units. Probably the most continuous fold structures are included on this diagram. Those that might be related to the buttressing or deforming effects of plutons are not included. Included are trends of folds in the complex of Monashee Gneiss of the Shuswap Terrane in the Vernon map area (Jones, 1959), the trend of the belt of Permian Cache Creek Group rocks that extends across the Shuswap terrane as the western extension of the Slocan fold, and that of the overturned synclinorium north of the Nelson batholith. The Spirit Pluton also follow this trend.

The Brodie-Sullivan kink fold

The east-west structures obviously are not restricted to the thrust belt, at least one extends eastward across the foldbelt and well into the homoclinal belt. This the Brodie-Sullivan kink fold--although a different type of structure than those described above--seems clearly a product of the cross fold event. The Brodie-Sullivan kink fold is a short limbed fold that resembles the minor structures known as kink bands, chevron-shaped folds produced across the grain of previously folded rocks. It, however, is not an angular but an arcuate deviation from the arc trend and is certainly not a minor structure because it extends from Brodie Mountain 5 km south of Northport S.80°E. along the north border of the Spirit pluton, past Sullivan Lake and Hall Mountain for a traceable distance of almost 50 km. West of Brodie Mountain it is lost among the confusion of thrust and tear faults in the core of the Columbia anticline. All arc structures, bedding planes, faults, and fold axes, are bent from a northeasterly trend to a southeasterly trend. The axial plane of the fold appears to have a steep southerly dip. Minor folds within the kink fold have vertical and near vertical plunges which contrast with the low angle folds north of the kink fold.

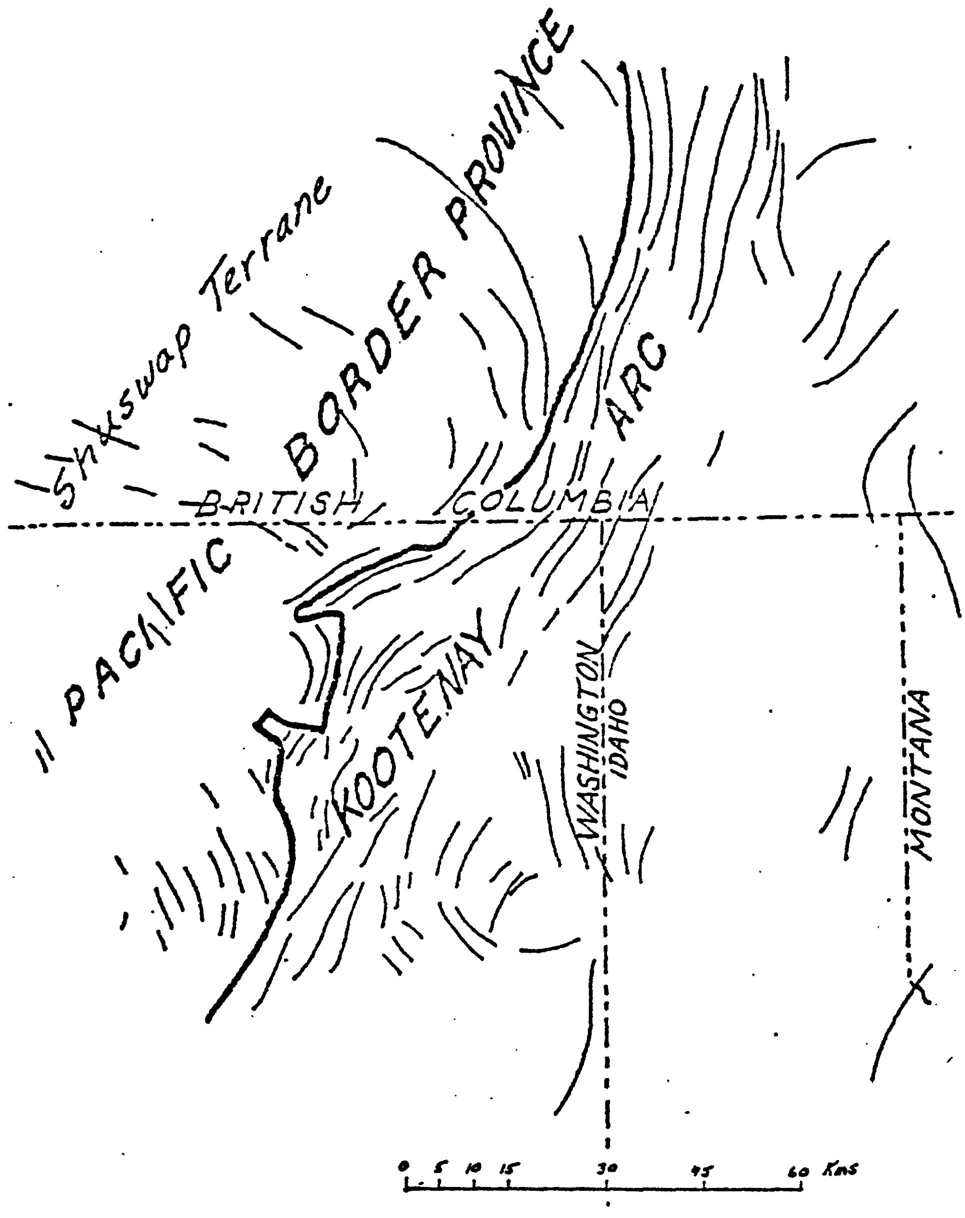


Figure 28.-- Map showing pattern of fold trends in Kootenay Arc and Pacific Border province. Heavy line marks province boundary.

An early impression of this structure was that it was restricted to the north margin of the Spirit pluton and resulted from forceful emplacement of the pluton. Later when mapping had been expanded, it was realized that the fold extended several kilometres beyond the pluton and projected to the west into thrust faults that indicated shortening along the same trend as that of the fold. It was then realized that the kink fold did not produce the pluton, but that the pluton's emplacement was either controlled by the fold or was contemporaneous and controlled by the same stress as the fold. The frozen contact between granitic rocks and folded beds clearly demonstrates that the pluton is later than the fold or contemporaneous with it.

The Brodie-Sullivan kink fold is obviously younger than the arc folds that it deforms; is older than the high angle north trending faults, such as the Flume Creek fault, which are not bent by the kink fold; and is older than, or contemporaneous with, the 100 million year old Spirit pluton, which is older than the Flume Creek fault. The kink fold therefore enables us to establish the following post-Middle Jurassic sequence of events: (1) folding along northeasterly trending axes, (2) thrusting and kinking along east-west axes, accompanied or followed by, (3) emplacement of the Spirit pluton, and (4) high angle extension faulting in the Late Cretaceous and Early Eocene.

Jurassic volcanic belt

The Jurassic volcanic belt is structurally the least understood of the three belts, mainly because only reconnaissance mapping was done in this belt, but partly because the ligholologic units are gross, and lack preciseness of stratigraphic position. Although the belt is named for the volcanic rocks of Jurassic age, an important part of the belt are the nonvolcanic rocks of Pennsylvanian and Permian age.

In British Columbia, the belt is bounded on the southeast by the Waneta thrust fault; in Washington it is bounded by a combination of faults of northeasterly, northwesterly, and north-south trends, which in the aggregate from a northeast trending boundary. Only one fold was mapped in Washington the southeast nose of the northwest trending anticline mapped by Bowman (1955) in the Orient quadrangle. In the West Nelson map area, of British Columbia, several folds were mapped by Little (1960), the most prominent of which are the anticline and syncline south of Silver Creek in the Bonnington Range. These are open folds whoses axes trend about N.30°E. and plunge gently to the southwest. North of Silver Creek, the fold trends swing to a northwesterly trend. As these folds involve Middle Jurassic rocks they are Late Jurassic or younger.

The folds in the Jurassic volcanic belt of British Columbia are cut by northerly, northeasterly, and northwesterly trending faults; the same fault trends are in Washington. These are high angle faults; low angle faults may be present but cannot be easily identified without detailed mapping. Some faults are younger than the Eocene volcanic rocks, others may be as old as Triassic.

The discrepancy between bedding attitudes in the Sophie Mountain Formation and those in the older rocks is, as mentioned in another section of this report, puzzling, because the dips measured in the conglomerate are steeper than those in adjacent older rocks. For example, the rocks in the Roberts Mountain Formation west of Belshazzar Mountain dip at moderate to low angles, whereas those in the younger Sophie Mountain Formation immediately to the northeast dip from 45° to vertical. Accompanying this apparently irrational relation between rock units and bedding dips is an equally puzzling relation between strike of bedding of adjacent rock units. The pattern of erratic attitudes suggests deformation by faulting, instead of by folding. Because both the Mount Roberts and Rossland Formations are older than the cross fold event it is probable that some folds are compressional, however, if the deformation dies out upwards--as is believed--it may never have reached the level of the Mesozoic rocks.

Sequence of events and interpretation of regional geology

This section might have been titled, Historical Geology, but so much of the history is recounted piecemeal in preceding sections that its systematized inclusion here would be redundant. A need exists, however, to summarize all events, so that some can be amplified, others interpreted, and still others correlated with those in distant provinces. Again we proceed chronologically, beginning with the Precambrian and continuing on through the Cenozoic. If the reader looses the thread of the account in what might appear to be digressions into analyses, speculations, multiple interpretations, and reconstructions he can refresh his memory by reference to the condensed arrangement in the chart (fig. 29) where the geologic history of the area is outlined. Before proceeding further, however, the reader is cautioned that it is important to remember--and easy to forget--that any interpretations, such as those that are to be made, are built upon a very limited number of hard unshakeable facts and that, as interpretations, they are subject to constant revision as new facts are found. If truth admits comparison, one has to be satisfied with interpretations that can never be more than relatively true.

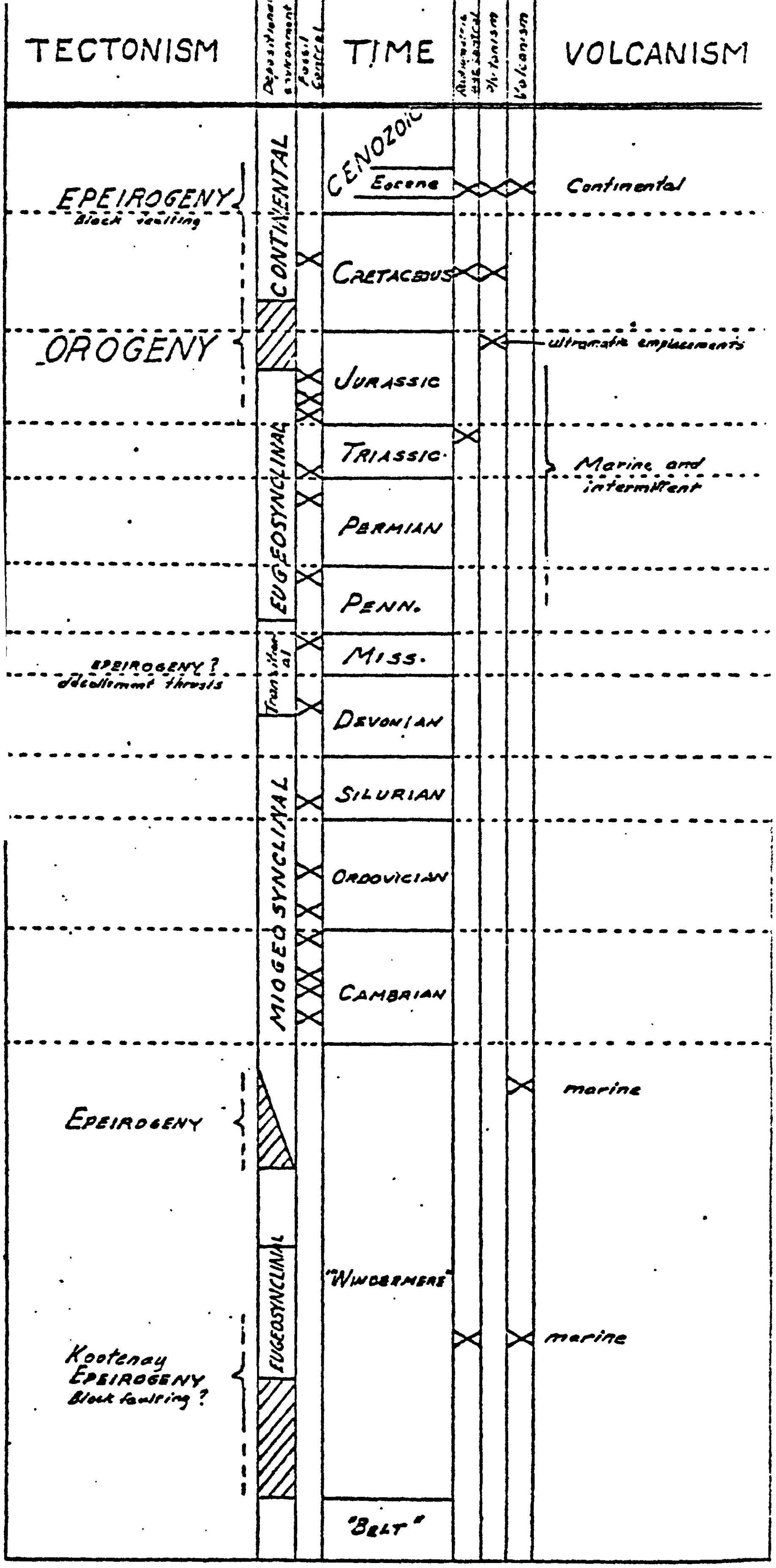


Figure 29.-- Chart illustrating geologic history of Northeastern Washington and adjacent Idaho and British Columbia.

Temporal control on the geologic history of the southern half of the Kootenay arc is generally imperfect. Fossils diagnostic of all periods, from Cambrian to Tertiary, have been found, but they are few and far apart both in space and time. Faunal control is supplemented by radiometric ages of igneous rocks of the late Precambrian, Triassic, Cretaceous, and Eocene. Sequences in the Precambrian and lower Paleozoic also can be established by superposition, but representatives of the upper Paleozoic and Mesozoic are for the most part, structurally isolated from each other, thus compelling the establishment of the stratigraphic sequence by more indirect and less exact methods. The period of time best understood is the lower Paleozoic where the superposition of strata is supported by fossils of Cambrian, Ordovician, Silurian, and Devonian ages. But even here, fossil horizons are too sparse to control precisely the age of individual formations or detect hiatuses between or within formations.

Sedimentation in the southern Kootenay arc was interrupted in the Precambrian by at least three demonstrable epeirogenic events, in the Mesozoic, by one orogenic event--perhaps intermittently long lived--and in the late Paleozoic by an epeirogenic event that can only be established by inference. The distinction made here between epeiorogenic and orogenic events is based mainly upon the angularity of the unconformity that established the tectonism, a criterion, which at best, is only of local applicability. In interpreting and correlating tectonic events, it must be remembered that the site of an orogenic event is linear, whereas the site of an epeirogeny is planar. For example, if only the dying edge of a foldbelt is exposed it could easily be mistaken for an epeirogenic event.

At the risk of circular reasoning, one can equate tectonic movement with the type of associated basement. The stable cratonic areas, unless softened by thermal mobilization of the basement, are the homes of the epeirogenies whereas the crustally weak areas, the continental borders, are the homes of the orogenies. On this basis, the Pacific Border province was a site of more or less continuous unrest until that province was welded by plutonism into a rigid basement.

Events and problems of the Precambrian

Belt-Purcell time

The histry of the Kootenay arc begins in the Proterozoic with deposition of the Belt-Purcell sedimentary rocks. For 600 million years (Harrison, 1972) fine- to medium-grained clastic sediments, along with minor carbonate muds, were deposited in a deltaic environment. Although a western source has been suggested for some of the 15,000 m to 23,000 m (50,000 to 75,000 ft) of sedimentary rocks, and eastern and southern provenance seems more logical.

The Belt rocks are in open folds and are cut by a multitude of faults. In the Coeur d' Arlene district, the folds are considered by Hobbs and others (1965, p. 113) to be older than the 1250 m.y. old uranium bearing veins, however this conclusion is at odds with the regional structure of the Belt rocks in the Kootenay arc, where the unconformities between Belt or Windermere rocks and Cambrian rocks are only very mildly angular. The assumption that all Belt rocks were gently folded during the Precambrian and only those in the arc refolded during the Mesozoic, is unwarranted, because the mind angular discordance between Precambrian and Cambrian rocks may be largely the result of tilting along high angle faults with only broad warping of the crystalline basement. Harrison (1972, pl. 10) dates movement on the Hope fault, a strike-slip fault belonging to the same system as the Osburn fault of the Coeur d'Alene district, as being Precambrian in origin.

The Belt terrane is an outstanding example of a depositional basin underlain by a stable crust, both during the 600 million years it took to accumulate the 15,000 m to 23,000 m (50,000 to 75,000 ft) of sediments and during the intervening period before the Mesozoic folding. Along the eastern and southeastern borders of the anticlinorium, Belt rocks rest unconformably upon the crystalline rocks of the craton, a product of the Hudsonian orogeny (1640 to 1820 m.y.). It is highly probable that these basement rocks extend under the entire Belt basin.

The state of the basement on which the Belt rocks were deposited can also be evaluated indirectly. Areas of high heat flow (steep geothermal gradients) indicate that zones of melting in the crust are closer to the surface than those of low heat flow. Orogenic belts have high heat flow; shield areas, low. If it can be assumed that the highest grade of metamorphism of Belt rocks unassociated with granitic rocks is related to depth of burial, one should be able to approximate the geothermal gradient at the time of maximum load and reach some conclusion on the stability of the underlying crust. It is further assumed that the internal stability of a portion of cratonic crust depends upon its thickness and depth to zones of melting, which is a function of the geothermal gradient; the lower the gradient the deeper the zone of melting and the greater the stability.

The accumulation of 15 kilometres to 23 kilometres (50,000 to 75,000 ft) of sediments without metamorphosim of the lowermost unit above biotite grade (Hobbs and others, 1965, p. 60) indicates that the geothermal gradient must have been low, because the commonly used gradient of 30°C/Km would place the lower part of the Prichard Formation, the oldest Belt unit, well above the melting point of granite. A gradient of 8.5°C/Km with an increase of .6°C/Km is that of the Gulf Coastal Plain (Clark, 1966, p. 488), probably more comparable to conditions in the Belt basin at the time of deposition. Using the diagram prepared by Hietanen (1967, fig. 1) to indicate the metamorphic stability fields (modified with a depth scale and reproduced as fig. 30) the Prichard rocks under 23 kilometres (75,000 ft) of overburden would have been in the muscovite-biotite field, those with 15 kilometres (50,000 ft) of overburden would have been in the lower part of the greenschist facies.

This projection, however, has not been verified; the one age determination of metamorphic biotite considered to have formed by load metamorphism is inconsistent with such an origin. In the Alberton area of Montana metamorphic biotite from the upper Prichard was determined by Obradovich and Peterman (1968) to be 1330±45 m.y. old. According to Harrison (1972, p. 1238), "such a metamorphism results from depth of burial and (steep) thermal gradient.' If the metamorphism was 1330 m.y. b.p., which also is the age of the St. Regis, according to Harrison (1972, p. 1236), it was metamorphosed under a load of 10,000 m (33,000 ft) of sediments and before the remaining 6,000 m (20,000 ft) of Belt sediments had been added to the load. If this was load metamorphism, one would not expect the metamorphic climax to be reached until after the total load had been added, which was 400 million years later. It would be preferable to correlate this metamorphism with a heat source other than that generated by depth of burial. Accordingly, we turn to the generation of granitic magma at depths well below the base of the Prichard. Only one example of a granitic pluton of the required age is known in the Belt-Purcell anticlinorium. This is the Hell Roaring Creek Stock of British Columbia, about 64 kilometres (40 mi) north of the International Boundary, which intruded the Aldrich Formation (upper Prichard) at least 1300 m.y. ago (Ryan and Blenkinsop, 1971). Until age determinations of biotites from

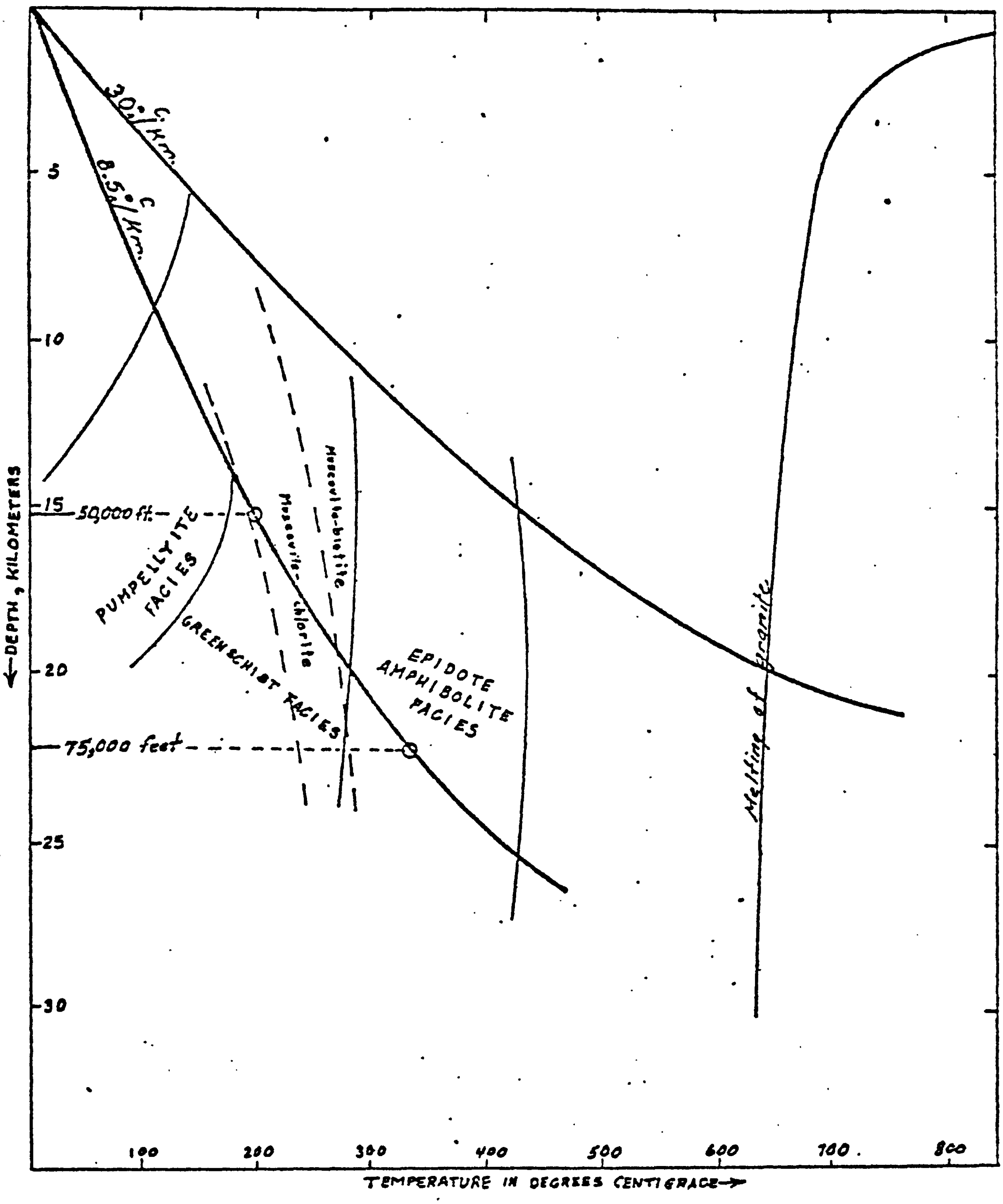


Figure 30.-- Relation of depth to metamorphism. Metamorphic fields after Hietanen (1967, fig. 1) corrected to agree with Hietanen (1973, p. 34).

many localities in the Prichard are made, we will not know if the Alberton biotite is representative of a regional metamorphism associated with plutons intruded into the Hudsonian basement. If the basement softened at this time, the softening was not accompanied by the development of foldbelts.

The contrast between the open folds of the Belt-Purcell anticlinorium and the tight folds of the Kootenay arc suggests a difference in the underlying basement. There is also a contrast between lithology and thickness of Belt equivalent rocks, the Priest River and Deer Trail Groups, that were deposited in the arc, and Belt rocks that were deposited in Idaho. Although the two lithologies may be juxtaposed by thrust faults, it is unlikely that this is the sole explanation. One can infer that the changes in the lithology of Belt equivalent rocks, the occurrence of mafic volcanic rocks in the overlying Windermere rocks, and the great thickening of Cambrian sediments in the arc collectively point towards either a thinning of the cratonic crust or the proximity of a boundary between continental and oceanic crust.

The companion questions, "how far west did the Hudsonian basement extend during Belt time? and how far west did Belt sedimentation extend and what facies is represented?" can only be replied to by further speculation. In Washington, the only gneissic rocks west of the arc that have been assigned to the Precambrian are in the northern Cascade Range. A lead-uranium geochronological study made by Mattinson (1970) found zircons in the Yellow Aster Complex. Swakane Gneiss, and Skagit Gneiss to have ages between 1600 and 2000 million years. According to Misch (1966, p. 106) the Yellow Aster is a low-potassium complex, whose lowermost units, schistose hornblendites and amphibolites, are conceivably remnants of primary oceanic crust. The Skagit Gneiss (Misch, 1966, p. 112) is derived from a eugeosynclinal suite of clastics and volcanics. The Swakane Gneiss is a metamorphosed clastic assemblage with eugeosynclinal affinities. Clearly all these rocks were formed at or beyond the continental margin.

In British Columbia geologists disagree as to whether any rocks exposed in the Shuswap Complex are of pre-Belt age--or of Belt age. Ross (1970, p. 61) suggests that the gneiss in the Frenchmans Cap dome is stratigraphically equivalent to Hudsonian gneisses of the Canadian Shield. On the other hand, Reeser (1970, p. 85), in discussing possible equivalence of units says, "Within the Shuswap terrane the horizon marked by the change in sedimentation at the beginning of the Lower Cambrian may be tentatively identified within the Mantling gneisses." The Mantling gneiss is supposedly underlain by core zone rocks that are migmatitic, heterogenous biotite-quartz-feldspar gneisses derived from pelitic to semipelitic metasedimentary rocks. Assuming an orderly succession and no tectonic or erosional deletion of strata, the Core zone rocks should be equivalent to the Windermere rocks. In other words, no rocks of Belt age have been identified in the Shuswap complex. Preto (1970, p. 32-34) supports Reeser's conclusion by correlating the oldest rocks in the Grand Forks map area with the Windermere Horsethief Creek Series. The 15 kilometre section of Belt rocks in the Chewelah-Loon Lake area (Miller, 1975) suggests sedimentation continued westward during Belt time, but the differences between Belt rocks and Deer Trail rocks suggests that the facies deposited would not be that of the Belt rocks. But none of this is positive evidence and we remain in doubt as to the presence and facies of any Belt equivalent rocks that might underlie the Pacific Border Province.

It is emphasized later in this paper that all rocks west of the arc, from the Devonian of the Cascade Range and San Juan Islands to the Lower Cretaceous, belong to the eugeosynclinal suite. Therefore it seems reasonable to postulate the western limit of the Hudsonian basement as somewhere near the boundary between the arc and the Pacific Border province. This requires that the inferred lower Cambrian rocks of the Shuswap complex, which are classified lithologically with the miogeosynclinal or shelf facies, were deposited on oceanic crust that was welded to the continental margin.

Before proceeding to the mid-Proterozoic epeirogeny, let us examine the ancestral Kootenay arc, which is believed to have been born in late Beltian time. To show changes in shape of the Belt basin, Harrison (1972, figs. 9, 10, 11, and 12) isopached representative stratigraphic units. Because basement rocks are not exposed within the basin, the lower Belt could not be measured. Harrison's isopach maps are reproduced as figure 31. Figure 31A indicates the size and shape of the middle part of the Ravalli Group; the axis of the basin during Ravalli time was N.70°W. Figure 31B representing the upper part of the Ravalli Group shows a bifurcating basin, with the N.70°W. axial trend retained in one fork and a N.30°W. trend in the other fork. In the isopach map of the middle Belt carbonates (fig. 31C), the N.70°W. trend has disappeared and the basin axis is now N.45°W. with the thickest accumulation in the northwest corner of Montana. In figure 31D the lower part of the upper Belt Group, the Missoula, reverses all previous trends; a high is developing within the basin so that accumulations in northeast Montana and northern Idaho are much less than those to the southeast. This N.25°E. trending high, sparked the beginning of the Kootenay arc. It developed into the positive area that was the source for the Windermere conglomerates.

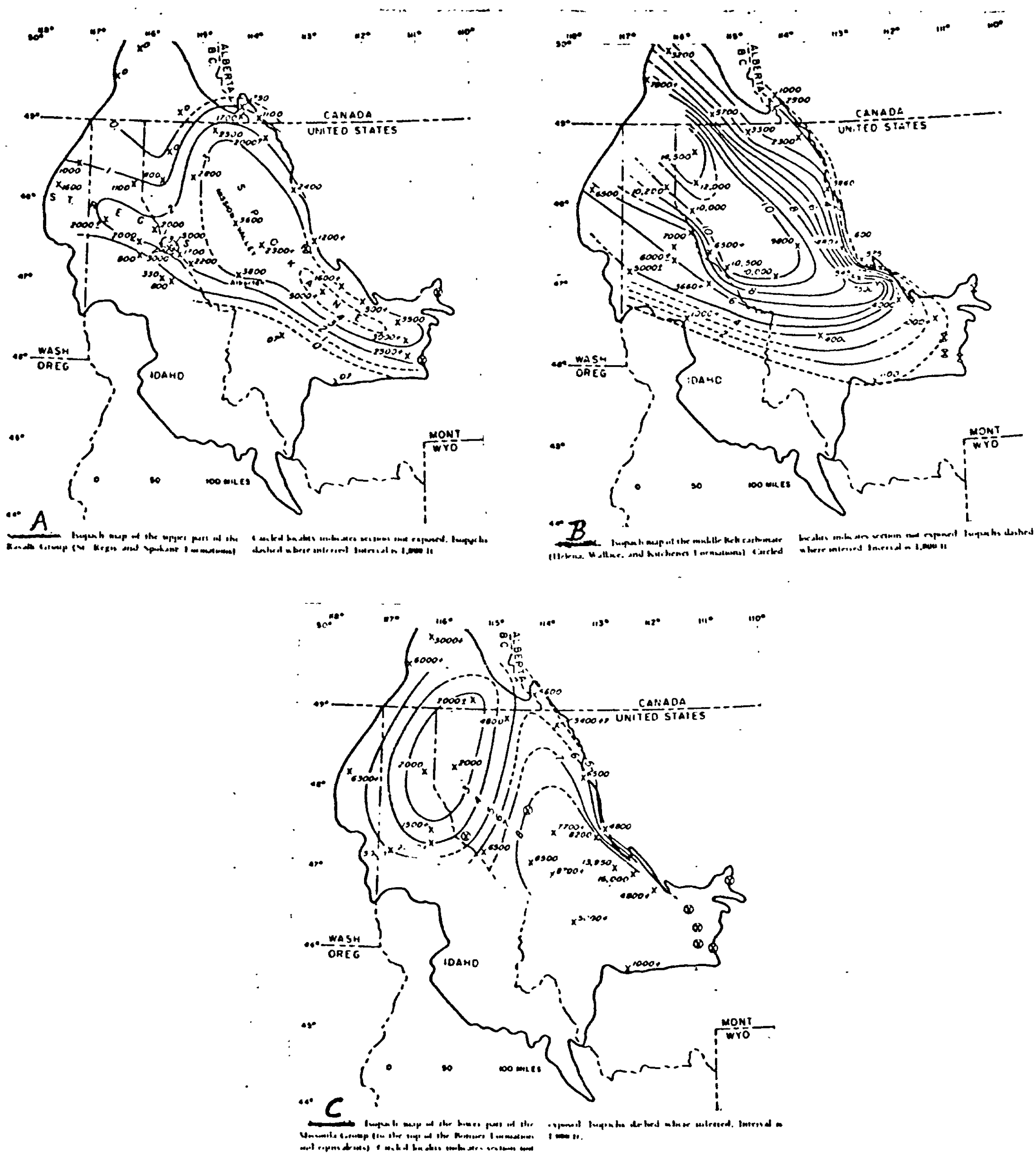


Figure 31.--Isopach maps of Belt-Purcell rocks. Taken from Harrison (1972, figures 10, 11, and 12).

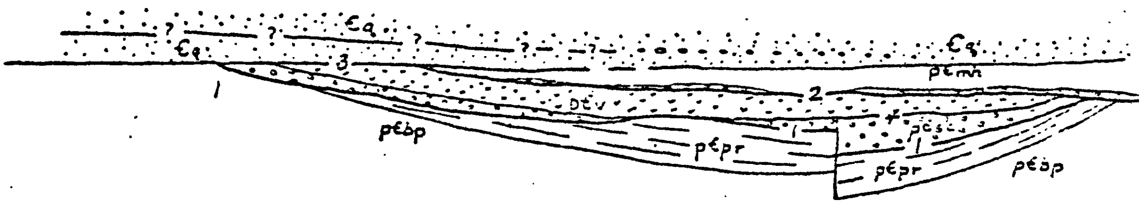


Figure 32.--Sketch showing relations between Precambrian and basal Cambrian units. Unconformities shown by heavy lines. Not drawn to scale. Angular relations between units exaggerated. Stratigraphic units as follows: p€bp, Belt-Purcell Supergroup; p€pr, Priest River Group; p€sc, Shedroof Conglomerate; p€v, Precambrian volcanic rocks (Irene, Leola, and Huckleberry); p€mh, Monk Formation and Horsethief Creek Group; Cq, Cambrian quartzites, includes Gypsy Quartzite, Addy Quartzite, and equivalent Canadian units. Numbers on unconformities correspond to those in text (p. 332).

Windermere time

Windermere time is defined (this report)as the time required to deposit the post-Belt Precambrian rocks that rest unconformably upon the Belt or Belt equivalent rocks (Purcell rocks). Belt and Windermere times are separated by the Kootenay epeirogeny, that period of time required to uplift and erode the Belt terrane.

Accepting the Holmes age of 570 m.y.b.p. as the end of the Precambrian, 930 m.y. as the youngest of the Belt rocks (Obradovich and Peterman, 1968) and 825 m.y. (age falls between 825-900 m.y.) (Miller and others, 1973, p. 3727) as the youngest age of the Windermere volcanics we can reconstruct, a crude calendar of the post-Belt Precambrian for northeastern Washington. The time available for deposition and epeirogeny during the Windermere is 360 m.y., which allows 100 m.y. to be alloted arbitrarily to the deposition of the Shedroof Conglomerate and epeirogeny and 260 m.y. to the deposition of the Monk Formation (Horsethief Creek) and lower part of the Gypsy. This distribution seems out of balance unless we interpret the unconformity below basal Monk conglomerate as representing a time loss much greater than the lack of angular discordance suggests. Also, the 100 m.y. before the Windermere volcanics were extruded is more than adequate to accomodate the deposition of 1500 m (5,000 ft) of shale and conglomerate, as well as the uplift and erosion of possibly as much as 3,000 m (10,000 ft) of Belt rocks. It would probably be closer to the truth, to reduce this period by half and add the deleted 50 m.y. to the duration of Belt sedimentation which would thus end at 880 m.y.b.p.

The Kootenay epeirogeny, as a Precambrian event, is recorded only in the rocks of the western and northern parts of the antilcinorium; in western Montana, where Middle Cambrian sandstone rests unconformably upon Belt rocks (Belt-Flathead unconformity), the epeirogeny extends into the Cambrian. In order to make the above statement, epeirogeny has to be defined as a complete cycle that includes the destruction of the uplifted area by erosion followed by downwarping to a position below sea level. Using this definition, the length of an epeirogeny, such as the one described, is less at the borders of the uplift than it is along the axis. This is because the depression of the positive area is accompanied by transgression across the Belt basement.

The Belt-Flathead unconformity of Montana supposedly represents erosion and nondeposition during all the Windermere and Lower Cambrian. If so, it is the summation of the several late Proterozoic unconformities observed in northeastern Washington (fig. 32). These are: (1) the unconformity between Belt (Belt equivalent) rocks and the Shedroof Conglomerate, (2) that at the base of the Monk and the top of the Windermere volcanics, and (3) that between the Belt rocks and the Lower Cambrian Addy (Gypsy) Quartzite, which is restricted to the Chewelah quadrangle. A fourth unconformity, (4) that between the Shedroof Conglomerate and the Leola Volcanics, which post dates the Precambrian movement on the Pass Creek fault zone, may be of local significance only. Two of these unconformities, no. 2 and no. 4 are between Windermere units, the other two are between Windermere and Belt rocks and Cambrian and Belt rocks. Because of the overlap relations (fig. 32) all unconformities are either part of, or merge with, the unconformity that separates the Belt basement from younger rocks. The unconformity and overlap relations between Belt equivalent and Windermere, and Lower Cambrian rocks is diagrammed by Becraft and Weis (1963, fig. 3) along a northeast trending section.

 p. 334 follows

The western limit of the Windermere volcanic rocks is unknown. If the correlations of Reeser (1970) and Preto (1970) are correct the equivalents of the Leona (Irene) Volcanics are not exposed in the Shuswap terrane. Both workers correlate the oldest exposed Shuswap rocks with the Horsethief Creek Series. Although Preto (1970) describes abundant amphibolites--which he interprets as basalt sills-in the gneiss section, he regards these rocks as no older than Cambrian and possibly as young as Permian.

The Windermere seas apparently did not extend far into Idaho and probably not at all into Montana; however, the possibility that the Lower Cambrian sea transgressed into Montana is not to be overlooked and is discussed in a later section. In summary, the Montana portion of the anticlinorium appears to have been a stable positive area during Windermere time, whereas northeastern Washington was oscillating between deposition and erosion.

The volcanic rocks and associated dikes and sills of the Windermere Group occur in a very narrow belt (fig. 33) that extends from a few miles north of the International Boundary southwesterly as far as the south edge of the Hunters quadrangle (lat 48°), where it is overlapped by Addy (Gypsy) Quartzite. This belt crosses two north trending regional faults, the Flume Creek fault in the Metaline quadrangle and the Jump-off Joe fault in the Chewelah quadrangle (Miller, 1975), without major offset (Geologic map of Washington, Huntting and others, 1961). Because the volcanic rocks everywhere along the belt have a westward dip, all associated dikes and sills are on the east side of the belt, intrusive into prevolcanic rocks.

The upper limit of the volcanic rocks of the Windermere is an erosional surface of unknown hiatal significance. The Monk Formation (Horsethief Creek), which lies on this surface, introduced the second major change in the

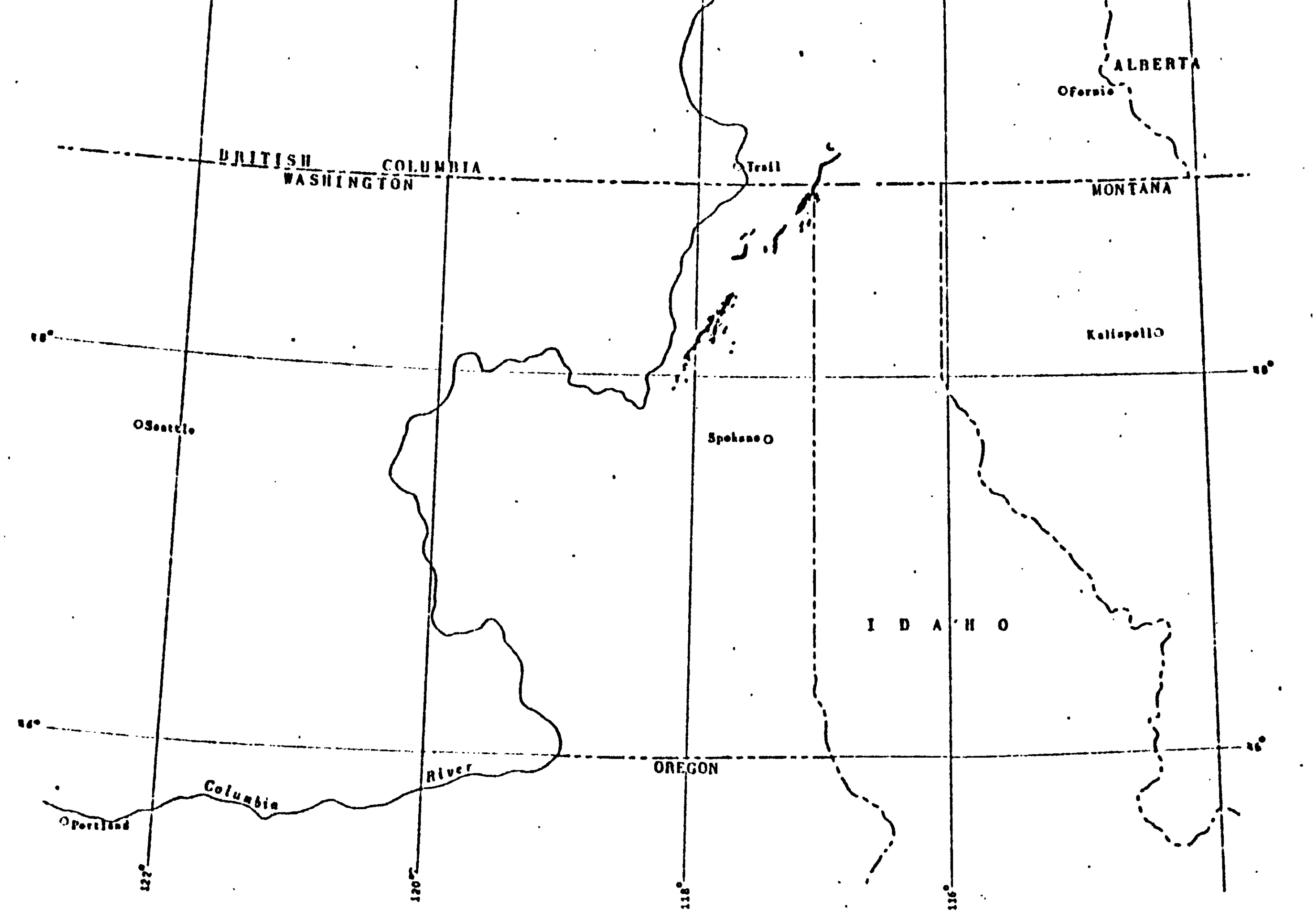


Figure 33. -- Distribution pattern of the Windermere volcanic and intrusive rocks.

depositional history of the Proterozoic--the first was caused by the Kootenay epeirogeny, which initiated the Windermere. This second change, which is within the Windermere, is a change from orogenic conglomerates and volcanics to miogeosynclinal quartzites, shales and limestones. The second stable period, which began with the deposition of the Monk, conceivably lasted for 400 million years.

We have seen that rocks representing these two long periods of crustal stability were brought together in Montana, where the Flathead Sandstone rests on Belt rocks, as well as in the eastern part of the Chewelah quadrangle, where the Addy Quartzite likewise rests on Belt rocks. This unconformity in the Chewelah quadrangle is the coalescence of two or more unconformities in the Metaline quadrangle. The tectonic nature of this interruption of the depositional cycle was clearly epeirogenic in Montana and, from the evidence available, no different in northeastern Washington, except that in the Metaline quadrangle, negative movement predominated over positive movement, allowing Windermere and Lower Cambrian rocks to accumulate to great thickness. In the area of the arc, a fold belt did not develop; the Precambria rocks are no more folded than the Cambrian rocks. The homoclinal structure of the eastern part of the arc includes both Windermere and Cambrian rocks, as well as, Priest River rocks, therefore any model of the Kootenay epeirogeny must consider the mildness of the unconformity between the two assemblages. A model proposed by Stewart (1972) is not in conflict with this relation--although it conflicts with other relations.

Stewart suggested that the North American continent was rifted during Windermere time in the area of the arc. He based his suggestion upon: (1) the highly irregular pattern of the late Proterozoic, pre-Windermere depositional basins, which contrast with the linear beltlike pattern of the Cambrian rocks, (2) occurrence of oceanic tholeiites in the Windermere strata, and (3) similarity of the model thus conceived to the theoretical concept that "thick sedimentary sequences accumulate along stable continental margins subsequent to a time of continental separation." The concept is attractive because, as a model, it is tensionally and not compressionally motivated, and hence does not produce a fold or thrust belt. As an additional attraction, it accommodates the initiation of the Cordilleran geosyncline along the course of the postulated separation.

Despite these attractive features the hypothesis is not supported by the geology of northeastern Washington, nor by that of Cordilleran North America. Stewart's first line of evidence, the contrast in depositional patterns between the pre-Windermere, late-Precambrian rocks and that of the Windermere-Lower Cambrian rocks (which Stewart groups together) can be explained by other than a rifting hypothesis. The pattern shown in Stewart's figure 4, reproduced here as part A of figure 34, is basically that of outcrop distribution, the balance between stratigraphic thickness and erosion. The patter, can be that of an embayed continental margin, where embayments received accumulations abnormally thick. Unless the sediments--which are regarded as marine--were deposited in a landlocked sea, which seems unlikely, there was more sea and more sediments to the west. To illustrate this I have added a dotted line to part B of figure 34 to indicate the continental margin as it may have been during Belt-Purcell time. Belt rocks deposited west of this line, if not subducted, would be buried under younger rocks. A depositional model such as this gives the continental margin of Belt time and overall trend very similar to that of the upper Precambrian and Cambrian, therefore the difference in sedimentation pattern is that the continent was embayed in Belt-Purcell time and was not in Windermere-Lower Cambrian time. An

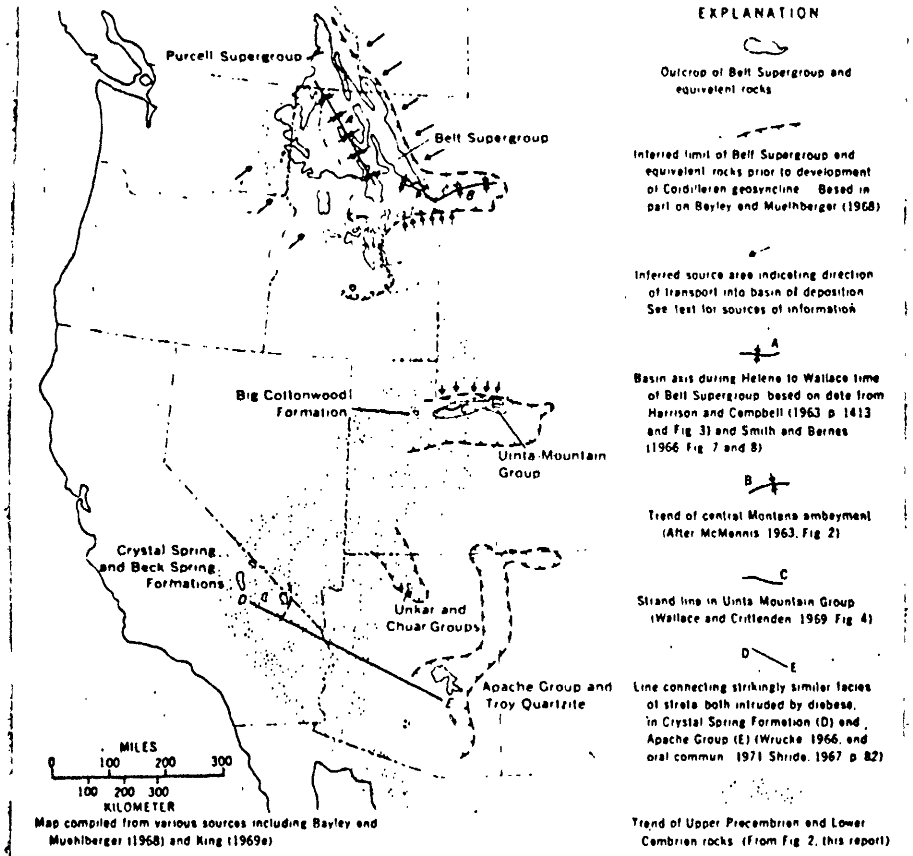
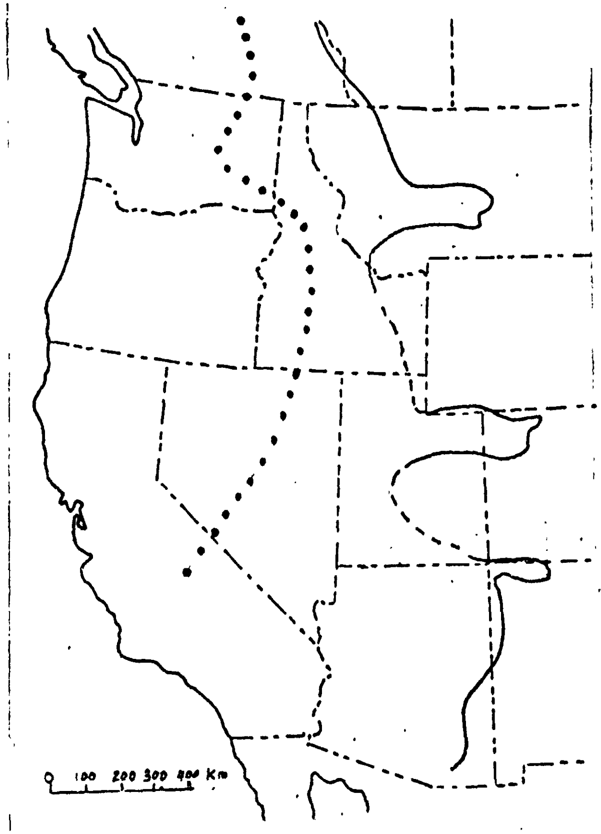


Figure 4. Comparison of distribution of upper Precambrian and Lower Cambrian strata with underlying supracrustal rocks.

A



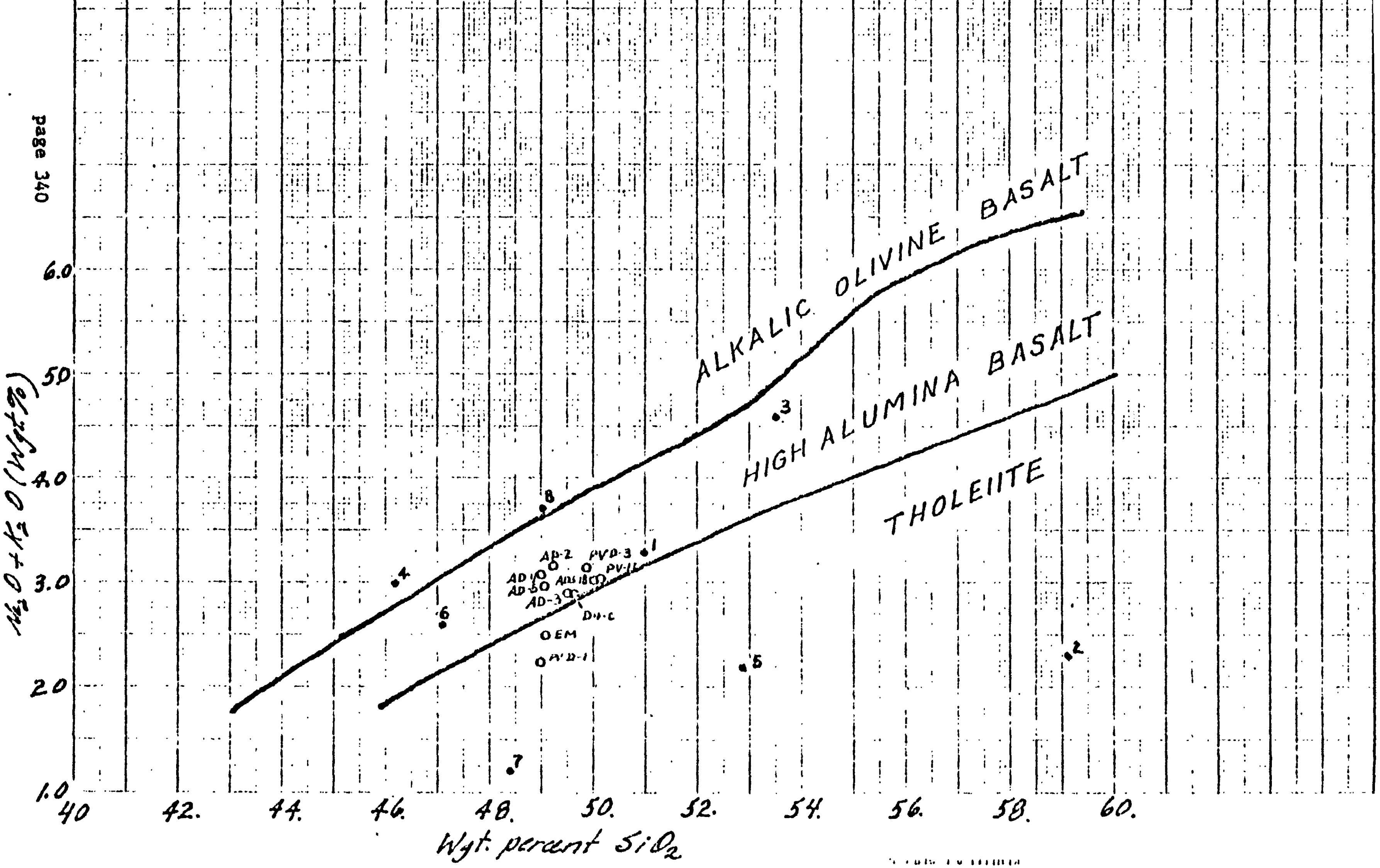
B

Figure 34. -- Distribution of Belt Supergroup and equivalent rocks. A, reproduction of map of J. H. Stewart (1972, fig. 4). B, reinterpretation of A: solid line represents eastward limit of deposition as determined by outcrop distribution (where dashed, the limit is inferred); heavy dotted line represents minimum westward extent of deposition.

embayed western North America is typical of much of the Paleozoic and Mesozoic. The presence or absence of embayments does not seem a valid criterion to be used for or against continental rifting.

The use of basaltic rocks to prove the tensional separation of continental blocks is a criterion that also must be used with caution. For example, three samples of Huckleberry Voclanics (Leola Volcanics) collected by Miller and Clark (1975) are cited by Stewart as being tholeiitic. Although correct, in general, the conclusion apeears unjustified because these rocks are chemically quite different from <u>oceanic</u> tholeiites and have an equal or greater affinity to plateau basalts. In potassium content, regarded as the most critical element in distinguishing continental from oceanic tholeiites, they most closely resemble the Archaen metatholeiite (Engel and others, 1965, p. 729), a continental basalt. In fact, if the analyses of basalts given by Stewart (1972, p. 1352) are corrected for water and carbon dioxide content and plotted on the diagram (fig. 35) introduced by Kuno (in Poldervaart, 1967, p. 627) to classify basalts, only three of the 8 examples classifies as an oceanic tholeiite, the high alumina basalt field of Kuno. Two are alkalic olivine basalts and three are non-oceanic tholeiites. To further clarify the point, I have plotted Engel's (1965) analyses of tholeiites collected from the Atlantic and Pacific Oceans and find that those from the Atlantic form a tight cluster within the high alumina basalt field and those from the Pacific bridge the high alumina and tholeiite boundary and the groups together form a field around which the Windermere samples are broadly and randomly scattered in a pattern that fails to show genetic kinship with any single class of basalt. If silica is plotted against potash, the scatter is even greater.

Figure 35.-- Alkali-silica diagram comparing Windermere basalts (Stewart, 1972, p.1352) shown by solid circles with Atlantic and Pacific samples (Engel, et al, 1965) shown by open circles. Sample numbers are those of authors quoted.



Although rifting during Windermere time cannot be established in the Kootenay arc on the evidence cited, the contrast in depositional pattern between Belt and Windermere rocks requires some explanation. The explanation lies in the provenances and depositional environments of the two groups of rocks. Belt rocks formed from, and accumulated on, cratonic crystalline rocks; the nonvolcanic Windermere rocks were derived in part from cratonic rocks and in part from Belt rocks and were deposited at or near the edge of the craton. The lower Windermere rocks are orogenic; Belt rocks are not. If the present distribution of Belt rocks in Idaho and Montana represents an estuarine embayment, the distribution reflects the drainage pattern, which in turn was probably controlled by tectonic weakness in the craton as Hobbs and others (1965, p. 125) first suggested and Harrison (1972, p. 1225) iterated seven yars later, "The thickest part of the prism (Revett) is parallel to the Lewis and Clark line...which suggests that the prism may be related to a troughlike depression along an ancient line of crustal weakness--the ancestral Lewis and Clark line." On the other hand, the distribution of Windermere rocks was controlled and initiated by the development of a welt, Montana, whose western edge was near and parallel to the cratonic margin. Presumably this welt was an effective barrier that prevented the rivers that crossed the Belt terrane from entering the sea at the site of the Kootenay arc; consequently, Windermere sediments, derived largely from the welt, could only accumulate along its western margin. It was not until the Cambrian that the welt was reduced by erosion and downwarped to a level where transgressing seas could bury it in sediments. Accompanying the rise of the welt a depression or trough developed to the west, which allowed the accumulation of the thick Windermere deposits.

As time went on, the trough was filled and the Windermere deposits spread both landward and seaward. On the landward side, the Toby Conglomerate (basal Monk Formation) represents the last of the orogenic deposits, except for conglomerate wedges that occur within the overlying Horsethief Creek (Monk) Formation in British Columbia. On the seaward side, are rocks of the Shuswap Complex that are believed to be equivalent to the Windermere.

How far into Idaho or Montana the Windermere conglomerate extended is highly speculative as there are no downfolded or downfaulted remnants to control our decision. To the north in Canada the Toby Conglomerate overlaps much of the anticlinorium (fig. 17). It is not unreasonable to assume that Windermere seas extended from Canada southeastward hardly more than a few miles into Montana.

The development of the postulated welt and offshore trough suggests the possibility that a subduction zone developed during the Windermere. This is what Monger and others (1972, fig. 4) proposed for the Canadian Cordillera, although they present the proposal as a questionable hypothesis. The concept of a Windermere subduction zone across northeastern Washington is as difficult to fit into the known geology as is the concept of a rift. The homoclinal section of Precambrian to Ordovician rocks in the Metaline quadrangle allows no freedom for subduction during the Windermere unless the subduction is accomplished without disturbance of the upper continental plate and the suture is buried under rocks to the west. Such a subduction would be a very passive subduction indeed. The model that is indicated--providing continental Proterozoic rocks are present in the Shuswap terrane--is a stable welding of continental and oceanic crust that permitted the extention of continental derived sediments onto an oceanic basement.

Precambrian structure

The thesis that the Belt-Purcell basin acted as a relatively stable continental area during the Precambrian is supported largely, and most directly by the mildness of the angularity between Belt and Cambrian beds. Indirect evidence, largely negative, is the general lack of tight folds. The accumulation of Belt sediments was uninterrupted by the extrusion of basaltic lavas and emplacement of sills during Belt time. Radiometric age measurements on sills in southwestern Alberta are interpreted (Hunt, 1962) as representing two different periods of emplacement, one about 1500 m.y. and the other about 1100 m.y. An even younger age of 870 m.y. is reported (Harrison, 1972, p. 11) for basaltic sills in the Prichard Formation, which could be correlative with the age of the Windermere volcanics. The 1300 m.y.b.p. biotite grade metamorphism of Prichard rocks suggests that plutonism was active south of the International Boundary as well as to the north.

Contrasting with the relative quiet of the igneous events, is the intense deformation that preceeded the emplacement of the uraninite veins in the Coeur d'Alene district of Idaho (Hobbs, and others, 1965, p. 125). This is the only reported indication of strong Precambrian deformation within the Belt Super Group and if the age measurement of 1250 million years is meaningful it must date a very local deformation.

Precambrian faulting has been demonstrated or inferred by several workers in Idaho and Washington (Hobbs, and others, 1965, p. 125; Miller, 1969, p. 5; Yates, 1970, p. 31; Harrison, 1972, p. 5). All the Precambrian faults have northwesterly trends and displacements that can be interpreted as right or left lateral or as normal movement. Faults that parallel the northeasterly axes of folds have not been recognized, probably because they are difficult to distinguish from later faults. Hobbs and others (1965,

p. 125) and Harrison (1972, p. 5) postulate movement on the Lewis and Clark line as having been initiated in the Precambrian. Northwesterly trending faults of unknown direction of displacement but with both right and left lateral apparent displacement terminate against Cambrian rocks in both the Loon Lake quadrangle (Miller, 1969, p. 5) and Metaline quadrangle (this paper).

None of the above Precambrian deformation can be referred to Belt time. It seems most likely that both warping and faulting occurred during the Windermere, as demonstrated for the Johns Creek fault in the Metaline quadrangle (this report).

Events and problems of the Cambrian

The exposed Cambrian of Washington is all of the miogeosynclinal facies, which is here, a tripartite lithology of sand, mud, and carbonate, deposited in that order. Changes between the lithologies are by interlayering; change to the overlying black shale facies is abrupt in most sections, transitional in one; change from Cambrian sand to the underlying rocks is an unconformity in the Chewelah quadrangle but conformable and in the pre-Cambrian in the Metaline quadrangle.

Rocks of the Cambrian are widespread. The sand or quartzite facies continues into British Columbia for hundreds of kilometres; the muds, as phyllites, for more than thirty kilometres. To the south, where the Paleozoic rocks disappear under the Tertiary lavas of the Columbia Plateau, sand and carbonates are present, but mud is limited to interbeds in the sand. To the west in the Northport quadrangle, the sand and carbonate continues and the mud is interlayered with carbonate. Still further west, in the Shuswap terrane of southern British Columbia thin representatives of sand and carbonate are provisionally correlated with Cambrian rocks to the east (Reeser, 1970, p. 85).

Diachrony and the sand facies

The Cambrian rocks extended eastward across the anticlinorium--recorded by downfaulted slivers resting on Belt rocks--to and beyond the thrust belt of Montana. The basal quartzite of the Cambrian looses its status as a lower Cambrian unit in Idaho and is regarded as Middle Cambrian, although the evidence for this is not conclusive.

At Pend Oreille Lake, Idaho, 80 kilometres east of the most eastern Cambrian rocks in Washington, the Cambrian section consists of the basal Gold Creek Quartzite, 120 metres thick, the Rennie Shale, 30 metres thick, and the Lakeview Limestone, a limestone dolomite unit, which is 600 metres thick with an eroded top. Fossils which occur near the base and top of the Rennie Shale belong to the Albertella zone (Lower-Middle Cambrian). Others from the base of the Lakeview are of the same zone. The shale and the lower part of the limestone are clearly of Middle Cambrian age and the quartzite, which contains no fossils is assumed to be Middle Cambrian because of its apparent conformable relation to the overlying fossiliferous Rennie Shale. About 130 kilometres (80 mi) further to the east in northwestern Montana at the Fishtrap Creek locality (Keim and Rector, 1964), a faulted syncline of Cambrian rocks overlies accordantly the Libby Formation of the Belt Supergroup. The section here is 7.6 metres of unfossiliferous sandstone overlain by 15 metres of shale that contains brachiopods and trilobites of the Albertella zone. These two sections of Iaho and Montana, deposited on the shelf, show a modest westward thickening, but nothing to compare with the thickening between the Idaho and Washington sections, where in 80 kilometres the quartzite increases from 120 metres to more than 2400 metres and the shale (phyllite) from 30 metres to over 1500 metres. In section, figure 36, the thickening is shown as a wedge with a apical angle of 3°, which, however, is not an abnormally rapid change from shelf to miogeosynclinal environment. There is certainly no need to telescope the section by thrust faulting. A time line dividing the Middle and Lower Cambrian is extended from an approximated point, 600 metres below the top of the Maitlen Phyllite, to the base of the Gold Creek Quartzite. As drawn, the section indicates a diachronous relation in the basal quartzite with the sea advancing eastward across the craton at the rate of 400,000 years per mile. This is a minimum rate because the base of the quartzite conceivably could be within and not above the Albertella zone. On the other hand, because of the absence of fossils in the Gold Creek Quartzite, the time line could be drawn to the top of just below the top of the Gold Creek Quartzite, which would place the quartzite in the lower Cambrian.

Until fossils are found in the Gold Creek Quartzite--or its Montana or Wyoming equivalent, the Flathead Sandstone--the age of the basal sand is uncertain. To assign it a Middle Cambrian age poses problems in sedimentation. Such an age assignment indicates that the lower Cambrian seas never transgressed far into northern Idaho, and the lower Middle Cambrian seas covered all of Idaho, Montana, and Wyoming. As the seas advanced eastward from the rapidly subsiding miogeosyncline, they covered the gently warped--and perhaps peneplained--Belt terrane with a thin blanket of sand that was derived, for the most part not from the underlying Belt rocks, but from the crystalline craton that lay beyond the Belt basin. This sand is remarkably free of detritus from Belt rocks, although Harrison (1965, p. 3) recognized a few cobbles of Belt rocks in the basal quartz pebble conglomerate of the Gold Creek Quartzite--which is composed predominantly of quartz pebbles. It is pertinent to note at this point, that scattered through the Gold Creek

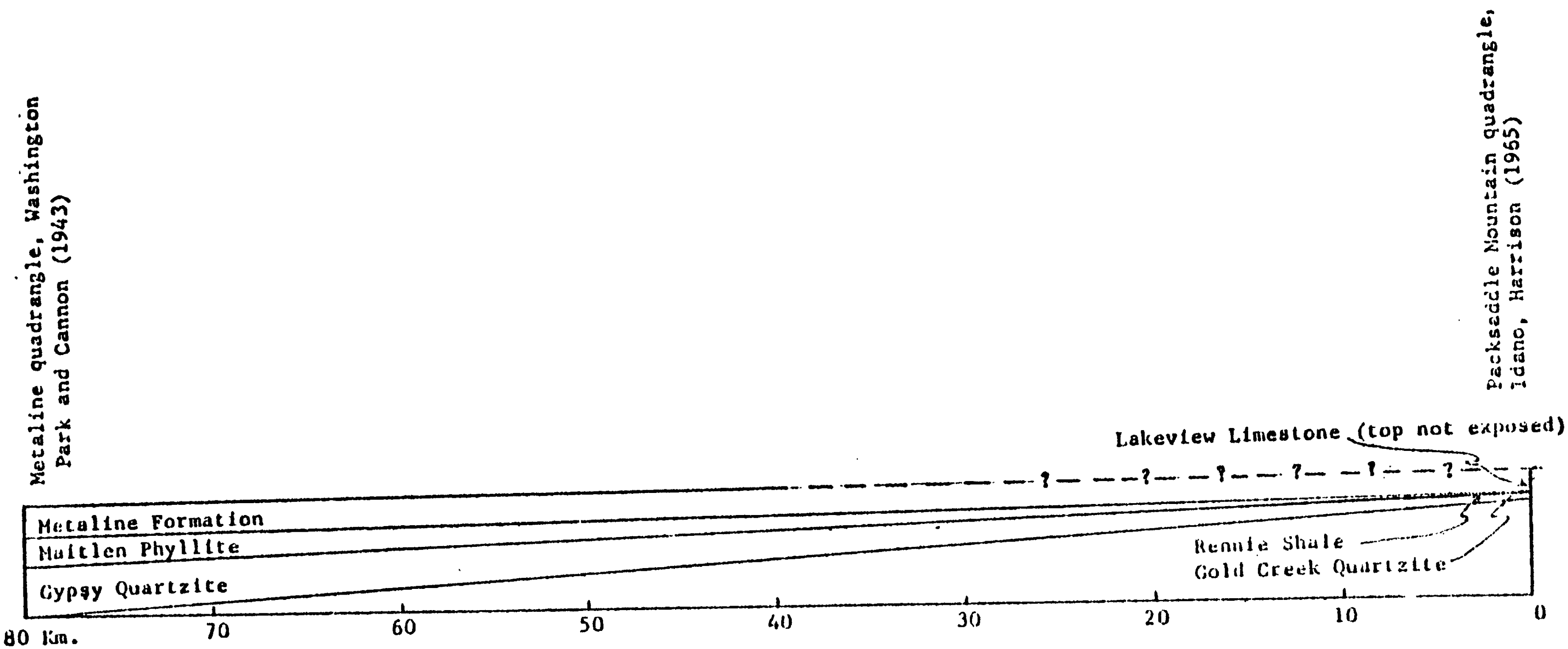


Figure 36.-- Figure showing eastward thinning of the Cambrian rocks between the Metaline quadrangle, Washington and the Packsaddle Mountain quadrangle, Idaho.

Quartzite are pebbles of jasper. Park and Cannon (1943, p. 14) also found similar pebbles of the Metaline quadrangle in a conglomerate 240 metres above the base of the Gypsy Quartzite and 2400 metres below the Lower Cambrian archaeocyathid limestone. These jasper pebbles and abundant quartzite clasts that could not be derived from Belt quartzites indicates considerable reworking and cannibalism of the Cambrian sands after cementation. The purple quartzite unit that occurs in the Addy Quartzite (Gypsy Quartzite) mapped by Miller and Clark in the Chewelah Loon Lake area (1975) may have been the source of the cobbles of purple quartzite in the conglomerate of the Gypsy Quartzite. These clasts indicate that deposition of Cambrian quartzite was not without interruption and that deposition occurred to the east of the Washington outcrops well before Middle Cambrian time.

It is tacitly assumed in the preceding paragraph that the Cambrian sands, pebbles, and cobbles, being derived almost exclusively from the North American craton, were transported across an eroded Belt terrane. The failure of the depositional process to erode the bedrock of Belt rocks and add the detritus to the load of the transporting streams suggests a very delicate balance between stream load and gradient. Such a balance requires a surface of transport that has no component of erosion or deposition. Streams capable of carrying cobbles and pebbles for several hundred miles do not have gradients that permit transport without erosion of the bedrock. This difficulty is eliminated, however, if a protective covering of craton derived material is deposited on the Belt bedrock by seaward filling, followed by partial erosion of this fill and its final redeposition as conglomerate and grits in northern Idaho and Washington. Such a history requires that at least the basal part of the Gold Creek Quartzite and the Flathead Sandstone be lower Cambrian, if not late Precambrian, in age, which in turn requires an unconformity representing a considerable time loss within the basal sandstone. The diachroninity of the basal sandstone should be questioned.

If both sand and mud lithofacies of the Cambrian are time transgressive the transgression should be reflected in the internal structure of the sand and in the relation between sand and mud facies. If both sand and mud transgress time, the boundary between them is also time transgressive and at any point in the encompassing time interval both sand and mud are being deposited, the sand shoreward and the mud seaward. The boundary between the two lithofacies is a gradation in grain size and composition if the transition is without reversals; if the elements that control deposition oscillate in effectiveness, the transition is interbedded mud and sand. All transitions will be present in both horizontal and vertical sections, therefore the sand (quartzite) sections in both miogeosyncline and shelf should show gradations upward from coarse to fine. Although conglomerates appear at or near the base there is no progressive downward coarsening from top to bottom in the sections that have been described--as there should be if the basal sands are truly diachronous.

There is even less evidence that the supposed diachronous relations of the sand lithofacies extended upward into the shale and carbonate lithofacies. The nature and distribution of the Reeves Limestone Member is not consistent with the interpretation of a time transgressing lithology. Although the upper boundary of the Gypsy Quartzite is taken at the base of the Reeves Limestone and the top of the Reeves as the base of the Maitlen Phyllite, the change from quartzite to phyllite occurs as an interlayering below the base of the Reeves. The deposition of the Reeves Limestone interrupted this change but did not initiate it. If the time transgression had continued into the Reeves one would expect a shoreward migration of the Reeves Limestone, but this apparently did not happen because the Reeves appears to be a linear deposit along the trend of the arc and not across it. The limestone extends to and beyond the northern end of the arc, is recognized in both Metaline and Leadpoint sections, and is probably the lowest limestone in the Red Top sequence in the Northport quadrangle. It has been traced as far as Colville to the south, but it has not been recognized in the Turtle Lake quadrangle (Becraft and Weiss, 1963). It is absent in the Pend Oreille Lake area of Northern Idaho and in all Cambrian exposures to the east in Montana and is not between the arenaceous (Cranbrook Formation) and argillaceous facies (Eager Formation) in the East Nelson and Cranbrook map area of British Columbia (Rice, 1937 and 1941). Its great lateral extent and restricted width suggest a reef deposit, a reef that did not migrate eastward. This suggestion is supported by the presence of reef forming Archeeocyathids (Okulich, 1948) that establish its lower Cambrian age. Its sudden appearance in, and disappearance from, the section suggests equally sudden changes in water depth or current directions. The lack of shoreward migration into Idaho and Montana is inconsistent with time transgressive relations.

The time-transgressive relations of the Cambrian--whatever they are--were doubtlessly influenced by the arcuate shaped depositional basin that had been established in the Precambrian. The northeasterly trend established by the birth of a high in northwest Montana during late Belt time persists into the Cambrian and is reflected in the shape and orientation of the sedimentary wedge formed by the sand and shale facies, which is shown by isopachs in figure 36A.

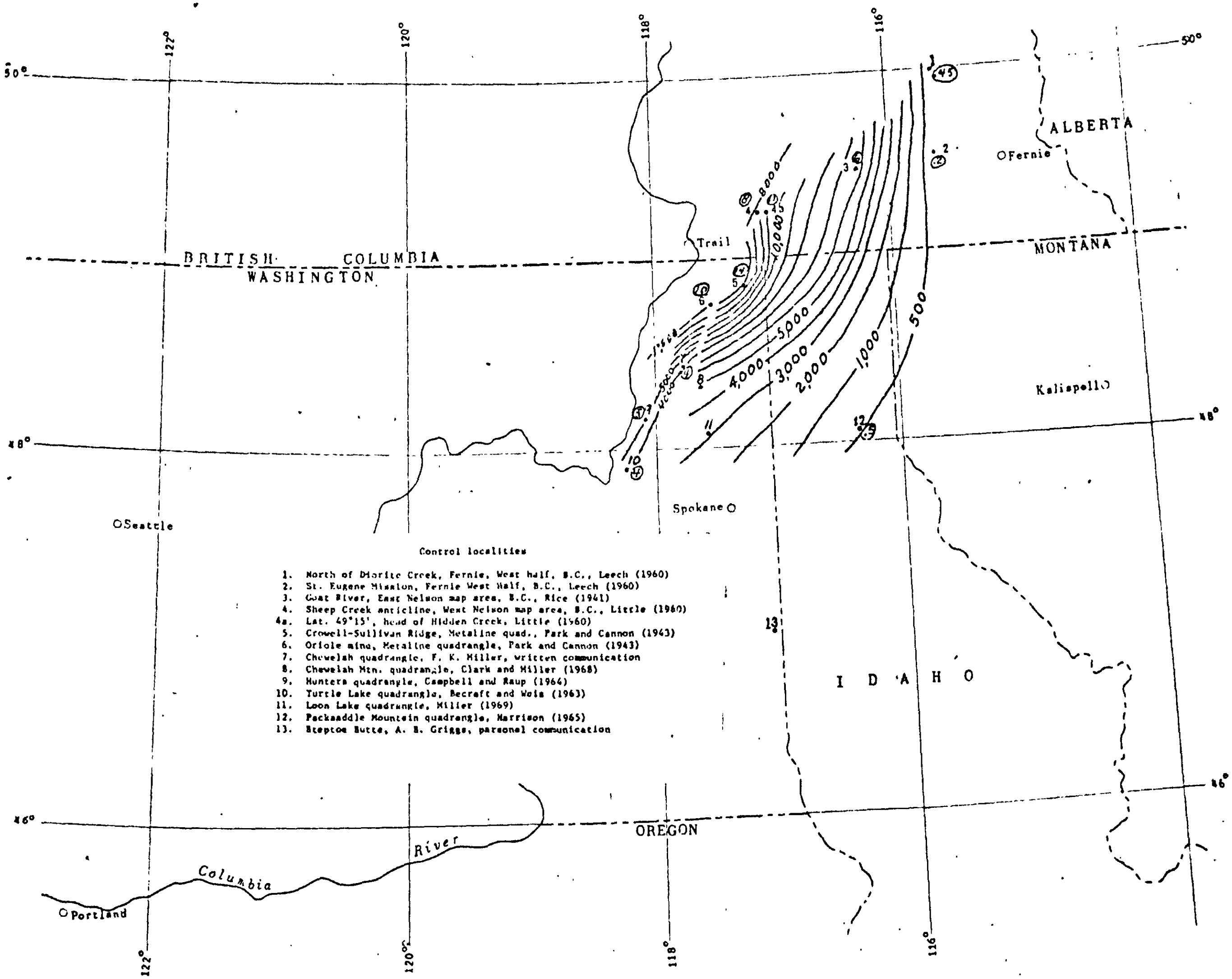


Figure 36A.-- Isopach map of Cambrian sand-shale facies, which consists of Gypsy Quartzite and Maitlen Phyllite and equivalents. Break in contours isolating control localities 4,6,7,9, and 10 may represent tectonic transport in the late Paleozoic of the rocks west of the disjunction. Thickness is indicated in thousands of feet by circled numbers at only those control localities where section is believed complete.

The argillaceous facies of the Cambrian is perhaps as widespread as the arenaceous facies. The Laib Formation (Maitlen Phyllite) can be traced only 30 kilometres north of the International Boundary, but it may have a possible equivalent in the Index Formation, which occurs as far north as the Ferguson area (Fyles and Eastwood, 1964, table p. 32). The equivalence is based upon the presence of the underlying Badshot Limestone (Reeves). To the south in the Turtle Lake quadrangle the argillaceous facies is restricted to phyllite units in the upper part of the Addy (Gypsy) Quartzite (Becraft and Weis, 1963, p. 12). To the west in the Columbia anticline it occurs interlayered with the calcareous facies. To the east in Idaho and Montana it separates the arenaceous from the calcareous facies and is considered Middle Cambrian in age.

The Maitlen Phyllite (Laib Formation) is probably not of deep water origin. Its close association, in both upper and lower parts, with limestone and its interlayering in the western arc with limestone suggest at least moderately shallow water and an eastern provenance.

Carbonate facies

The Cambrian carbonate facies, represented by the Metaline Formation in Washington and the Nelway Formation in the Salmo district of British Columbia, extends as far to the southwest as the rocks of the Kootenay arc can be traced, but only a few miles north of the International Boundary. These carbonate rocks probably are part of a great blanket that extended northeasterly across the Belt-Purcell anticlinorium in towards Canal Flats. Most likely the facies overlaid the Eager Formation in the Cranbrook map area (Rice, 1937, p. 18-19) and certainly extended across the panhandle of Idaho and Montana where limestones are interbedded with calcareous shale and described (Lochman, 1957) as having been deposited in a shallow sea over a stable shelf. The Middle Cambrian section of western Montana as described by Lochman (1957, p. 117) in terms of broad stratigraphic units is, (1) a basal transgressive sandstone, (2) a succeeding shale, (3) a limestone with interbeds of shale, and (4) a repressive shale and limestone pebble conglomerate. The conditions of sedimentation continue into the Late Cambrian; the pebble conglomerate is succeeded by a fragmental oolitic limestone that wedges out eastward. Above the limestone is a variable succession of sandy shales, thin limestone, limestone pebble conglomerates, and dolomitic siltstone which extends into the basal Ordovician. Whether or not any of the limestones in the Shuswap terrane are tongues of this carbonate blanket is unknown, but the correlation is a reasonable hypothesis.

Limestone deposition also continued into the Late Cambrian in the Kootenay arc. However, the stratigraphic position of the controlling fossils is widely different in different parts of the arc. In the Leadpoint section Late Cambrian(?) fossils occur in the unit transitional between carbonate rocks and black slates of the Ordovician. In contrast, the Late Cambrian brachiopods found by A. B. Campbell (personal commun.) in the Hunters quadrangle and figured by A. R. Palmer (1959, personal commun.) as Angulotreta and Dictyonina perforata of Early Late Cambrian age occur about 150 metres (500 ft) above the top of the Addy (Gypsy) Quartzite and the base of the limestone mapped as "Old Dominion Limestone of Weaver," which is the lithologic equivalent of the Metaline Formation. The several thousand feet of limestone and cherty limestone that lies above the fossils and below the Middle Ordovician slates suggests that the upper part of the Old Dominion in this area can be Early Ordovician. If the age assignments are correct, the Cambrian carbonate facies overlaps southward as well as eastward. Upper

Cambrian limestone on Lower Cambrian quartzite indicates that the limestone rests on a Middle Cambrian surface of disconformity. In the same general area the Addy Quartzite rests unconformably upon Windermere rocks and on lower Deertrail rocks. It seems likely that a positive area somewhere south of the arc was not only active during the very late Precambrian, but during the Middle Cambrian as well.

McConnel and Anderson (1968, p. 1476) speak of the uppermost beds of the Metaline Formation (Josephine horizon) as belonging to "an area of algal reefs, inter-reef, and off-reef basin." However, the term reef can hardly be applied with authority to the whole of the Cambrian carbonate facies. The term as now used requires that the carbonates be the product of reef forming organisms, a condition, although perhaps fulfilled in part in the Metaline Formation, is without faunal support. The sparse fossils that have been found are not reef formers. It is better described as a bank deposit, using this term for extensive shallow water, marine carbonate deposits located well beyond the reach of contamination by clastic sediments and at least locally exposed to the air and to wave action. As a depositional unit it crossed both miogeosyncline and shelf, differing mainly in stratigraphic thickness.

The lack of fossils capable of being used for zonation in the miogeosynclinal deposits makes it impossible to detect diachroninity between miogeosyncline and shelf. Certainly evidence is lacking of diachroninity similar to that reported by Lochman-Balk (1972, p. 41, fig. 11) in Montana, where a carbonate facies transgresses time eastward from the Early Middle Cambrian, through the Late Cambrian into the Early Ordovician. In fact, the reverse appears to be true; for example, the basal Lakeview Limestone of northern Idaho contains Albertella stage fossils of the Lower Middle Cambrian, whereas the basal carbonate beds of the Metaline Formation are in the Bathyuriscus

Elrathina zone of the Middle Cambrian. Another example is further southwest in the Hunters quadrangle where the basal Cambrian limestone is Early Late Cambrian in age. In both example the oldest Middle Cambrian carbonate rocks are in the east and not in the west. It appears that the miogeosyncline is truly a miogeosyncline and not a miogeocline (Dietz, 1966, p. 566).

Black shale facies

The change from carbonate sedimentation of the Cambrian to the euxinic suite of black muds of the Ordovician was abrupt and apparently without oscillation, except for the easternmost deposits--which are interpreted to have been thrust westward to become the Leadpoint section. This change probably was caused by either a change in water depth or a change in bottom currents. A change in water depth, resulting from an increased rate of subsidence that depressed the carbonate environment into a depth zone where muds were being transported by long shore currents, could explain the change in lithology but not the suddenness of the change. Nor will a change in water depth explain the sudden appearance of the black graptolitic shales of the Glenogle Formation--the equivalent in age and lithology of the Leadbetter Slate--that occurs in the Rocky Mountains of southeastern British Columbia as part of shallow water facies. Comparison of the black shale facies of northeastern Washington with the Ordovician black shale facies of southeastern British Columbia, however, is not realistic, because the Glenogle shale, little more than 300 metres thick, is immediately overlain by the Wonah Quartzite and followed by more limestones and dolomites, whereas the Ledbetter Slate is only the basal member of a black shale facies that extends at least into the Devonian and probably into the Mississippian and is many hundreds of metres thick; the shallow water carbonate facies never returns to the Kootenay arc. The two diverse sequences can be equated by interpreting the Glenogle Formation as a tonguelike extension of a much thicker deposit that has been eroded from the anticlinorium and the Ledbetter Slate as having had a tongue that extended into the panhandle of Idaho and southeastern British Columbia where it was interlayered with carbonate rocks and quartzite.

Whether or not the black shale facies was a deep water deposit cannot be resolved in this discussion, but the postulate of a deep water environment is a necessary premise to the explanation, this report, of the distributio of pods and lenses of Devonian bioclastic limestone in the black shale facies by the downslope sliding of blocks of limestone that had been deposited at a higher level on the shelf.

Black shales of the Ordovician, Silurian, and Devonian

To establish the eastward extent of the black shale facies is as difficult as estimating water depths. The eastern limits of Ordovician, Silurian, and Devonian sedimentation is not directly measurable, but it is commonly inferred from a Cambrian-Devonian unconformity that western Montana was upwarped in the lower Ordovician (Lochman, 1957, p. 158) thus ending marine deposition until the Devonian. The corrollary to this inference is that the uplifted area would shed detritus westward and that this detritus would accumulate in the Ordovician and younger rocks. The black muds of the Ordovician could not, however, have been derived from a highland that was covered by carbonate rocks of the Cambrian. The slices of Cambrian carbonate rock downfaulted during the Mesozoic(?) into the Belt-Purcell anticlinorium strongly suggests that the anticlinorium was not being eroded during the early Paleozoic unless the Cambrian-Devonian unconformity represents not Ordovician, but Late Silurian or Early Devonian, uplift and erosion.

Accordingly, it seems likely that the Ordovician seas extended eastward across the panhandle of Idaho into Montana. The black shales of northeastern Washington are not deposits of the littoral and certainly not of the strand line: deposits of these environments should have been deposited somewhere to the east. In Idaho and western Montana there could have been a shoreward sand facies, as quartzite beds in the Ledbetter of Slate Creek suggest, but if the above inferences are correct, the sand did not come from the anticlinorium nor is it likely that it reached the strand line that extended across northwestern Montana. It may have migrated northward from central Idaho and once formed a continuous deposit with the Kinnikinic Quartzite (Ross, 1962).

Provenance for the sand was probably the same as that for the Cambrian sand, the Canadian Shield, but the route of transport, from shield to northern Idaho, is difficult to establish. Because of the distance from exposed rocks of the Canadian Shield to the place of deposition in central and northern Idaho, it is doubtful if more than one river system delivered the sand to the sea. The mouth of the river may have been in central Idaho and from here the sand moved northward by long shore drift to northern Idaho and northwestern Montana. In southern Idaho during the Silurian and Devonian carbonate rocks were being deposited upon the sandstone that was to become the Kinnikinik Quartzite, therefore the sea was advancing northeastward as the mouth of the river retreated. By this time, the uplift of the anticlinorium had begun and the rivers gradient had been so reduced that it was no longer capable of carrying sand but only mud and fine carbonate detritus that was being eroded from the anticlinorium. The carbonate detritus was deposited in a bank off the mouth of the river and the muds moved through channels in the bank to open sea and thence northward to be added to the Ordovician deposits in northeastern Washington.

The sand of the Eureka Quartzite of northern Nevada, which was deposited at about the same time as the Kinnikinick Quartzite probably was not transported by the same river system. It belongs to a different assemblage, the Eastern Assemblage of Roberts and others (1958), a miogeosynclinal, basically carbonate facies that is separated by a thrust fault from the black shale facies (Vinini Formation) of the Western Assemblage (eugeosynclinal). The carbonate association of the Eureka Quartzite contrasts with that of the Kinnikinick, which occurs within a shale facies. The shale facies of Nevada, considered as eugeosynclinal, sharply contrasts with the black shales of Washington, which are deposited conformably upon miogeosynclinal carbonate rocks and are clearly part of the miogeosynclinal assemblage, unless one defines the eugeosyncline on the presence of black shales and permits the change from "mio" to "eu" to change in the vertical direction instead of in the horizontal.

The Mississippian and the problem of mid-Paleozoic orogeny

Miller and Clark (1975, p. 30-33) have mapped and described carbonate rocks of the Loon Lake quadrangle that contain Mississippian and possibly Devonian fossils. These rocks are of a distinctly different facies than that of the Ordovician to Devonian in the Metaline and Colville quadrangles. The environment defined is intertidal. Because the limestones are fault bounded and eroded at the top, their relation to other Paleozoic rocks is uncertain. Conodonts from one locality (p. 127) represent about the same fauna as that from the basal Banff of Alberta and is "one zone higher than the fauna described from the basal Lodgepole," which places it in the Early

Mississippian. The association of Early Mississippian and questionable Devonian limestones suggests that these, conceivably allochthonous, rocks may bridge the Devonian-Mississippian boundary. They may be part of an extensive carbonate bank that covered all the Rocky Mountain states and at least part of the eastern Great Basin and a bank whose western border was not far from the Idaho-Washington boundary. The postulated gravity slides of Devonian limestone in the Metaline quadrangle may represent the edge of the shelf, and the Loon Lake rocks may have been carried by gravity thrusts a score of kilometres or more westward.

Although the limestone of the Chewelah quadrangle is the only known fossiliferous Mississippian in northeastern Washington, most probably some of the dark gray to black unfossiliferous pelitic rocks in the Deep Creek area and Northport quadrangle are likewise Mississippian. In the Colville and Metaline 30-minute quadrangles, the fossil control for the black shale facies extends from the Cambrian into the Devonian; control for rocks possibly Mississippian is absent, but is present for eugeosynclinal Pennsylvanian and Permian rocks. Because the undated rocks of possible Mississippian age belong to a facies intermediate between the two fossiliferous assemblages, they are interpreted as being of intermediate age and a part of a depositional sequence that extends from the Cambrian to the Permian without a major break in deposition.

The transition in facies can be interpreted from a study of rocks in either the Deep Creek-Northport area or the southwest quarter of the Colville 30-minute quadrangle. In both areas the section is faulted, but there is no reason to believe that two depositionally unrelated assemblages have been superimposed by regional thrusts. No matter which fault or combination of faults one might select for such a divider, the rocks on either side of the fault would have common characteristics. This is not to suggest the section has not been telescoped; Permian rocks near Kettle Falls may have been thrust eastward--but not more than a few miles. The belief stated in the preceeding paragraph, that deposition was uninterrupted by a major break in deposition does not imply that minor breaks are absent. Conglomerates within the sequence speak eloquently of uplift and erosion in nearby areas, however the location of structural highs and dating of uplift is as yet, speculative; consequently the correlation of this postulated event with either the Antler orogeny of the Great Basin or a mid-Paleozoic disturbance in British Columbia is premature.

Conglomerates and their significance

Perhaps the key to an understanding of the depositional history of the upper Paleozoic is in the origin of the conglomerate that occurs in the pelite sequence that crops out near Kettle Falls southwest of Williams Lake. East of Kettle Falls on Rattlesnake Mountain is the largest outcrop of post-Windermere, pre-Cretaceous conglomerate in northeastern Washington. The conglomerate is either Permian or pre-Permian; to the northwest are rocks of the eugeosynclinal facies that have limestone pods with Permian fossils, to the northeast are dark argillites, graywackes, grits, and mafic volcanics and rare limestone pods--but no fossils--that extend eastward to the Ordovician graptolitic slate. The conglomerate appears to intertongue to the northeast with black slates and argillites interbedded with thin layers of coarse graywacke whose clasts are similar to but finer grained than those of the conglomerate.

The conglomerate is a massive rock, devoid of bedding and having a minimum of matrix. The texture is seriate, consisting of highly angular clasts as large as 6 inches welded into a hard compact rock by a quartz cement. Small cavities record the solution of limestone clasts; in one cavity W.A.G. Bennet (Resser, 1934) found silicified of Nisuia hyoliths and a small trilobite suggesting Agnostus, fossils that have been found in Middle Cambrian rocks of Washington and Idaho. Under the microscope, the larger clasts are mainly quartz, quartzite, siltite, and black chert and lesser amounts of phyllite, argillite, argillaceous limestone and grit. In the finer grained rocks of the conglomerate, which are coarse graywackes, a few grains of sodic plagioclase and microcline were seen. No clasts of volcanic rocks were noted. Clasts of argillite and phyllite, soft when incorporated, are squeezed into shape accommodating shapes.

Because the conglomerates were not mapped in detail and no contacts with the black argillites were directly observed, it is difficult to know how the conglomerate was introduced into the black pelite sequence. Clasts of the enclosing black argillite in the conglomerate indicates that erosion of the sequence was going on as the conglomerate was deposited. This indicates subarea erosion, a debris flow, or turbidity current deposits. The limited distribution of the conglomerate is interpreted to indicate deposition in channels which could be either by debris flows or turbidity current gravels.

Evidence of the source of the clasts is contradictory. A western source is favored by the eastward interlayering of the conglomerate with black slates, the apparent eastward decrease in grain size, the presence of black chert, and the close similarity to Permian sharpstone conglomerates of the Okanogan region (Rinehart and Fox, 1972, p. 8-9)—part of the eugeosynclinal suite. An eastern source is favored by the abundance of quartz and quartzite clasts, the presence of Cambrian fossils in limestone clasts, the absence of volcanic detritus, and the close spacial association with the black argillite facies, which conformably overlies the Cambrian limestones of the miogeosyncline.

The presence of Cambrian fossils in clasts in a conglomerate that is Permian or as old as late Devonian, indicates a pre-Permian uplift along with the erosion of all post-Cambrian rocks older than the late Paleozoic conglomerate. To accurately date this uplift requires a more precise age for the Kettle Falls conglomerate than late Paleozoic. If the uplift was west of the arc, the Cambrian carbonate facies would have extended at least as far west as the uplift and the quartzite clasts would have come from quartzites deposited in a volcanic free terrane.

According to descriptions by Reeser (1970, p. 85), a source of rocks of appropriate compositions and age occurs in the mantling gneiss of the gneiss domes of the Shuswap Complex to the northwest in British Columbia. Reeser provisionally recognizes the facies change from Windermere to Lower Cambrian rocks that occurs in the Kootenay arc as represented in the mantling gneiss zone, which "consists of quartzite, pelitic schist, quartzite paragneiss and much marble and calc silicate gneiss." He also recognizes lithologies of the Milford Group (Mississippian and later) in the Pinnacles dome of the complex. If Reeser is correct in his tentative correlations, and if the source of the carbonate clasts is in the west, the time of the uplift is pre-Mississippian and post-Cambrian. By comparing the overlap of this time bracket with the timespan within which the conglomerate falls, the time of uplift and erosion, Late Devonian and Early Mississippian, coincides remarkably well with the Antler orogeny. Nevertheless, there is no indication whatever that the rocks that form the mantling gneiss were exposed to erosion during the mid-Paleozoic.

A better and permissive correlation can be made if an uplift is postulated in the Idaho panhandle during the Late Devonian or Early Mississippian. The presence of reworked Cambrian fossils in rocks no younger than Permian indicates the pre-Permian post-Devonian erosion of the black shale facies to the depth of the Cambrian carbonate facies. The quartzite clasts could have come from the postulated sand facies of the Ordovician and the chert clasts from nodular chert in carbonate rocks such as the upper limestone unit of the Metaline Formation or from beds of black chert in the Ordovician-Silurian rocks. Dark gray to black cherts interlayered with limestones of Cambrian or Ordovician age occur in the Hunters quadrangle (Campbell and Raup, 1964). These beds may represent the postulated chert facies of the Ledbetter. The presence of conglomerates in the black slates that lie above the Devonian slumped blocks indicates that whereas northeastern Washington was a depositional province, somewhere to the east a sandy facies of the Lower Paleozoic was being eroded.

Clasts in the Kettle Falls conglomerate could not have come from both east and west. The conglomerate occurs as a discrete isolated body, therefore it requires a remarkable chain of coincidences to supply it with detritus from both east and west. The evidence favors an eastern source, but this conclusion is influenced by the need for an eastern uplift to explain the westward thrusting of the Leadpoint section of the Cambrian rocks.

In a preceding section, Rocks of the Pacific Borderland, the rock assemblages that are predominantly fine clastics cratonically derived but contain volcanic units are classified as rocks transitional to the eugeosynclinal facies and are considered most likely of Late Paleozoic age. However, for the sake of discussion, it is admitted that a part of this argillic suite could be allochthonous and correlative to the Lardeau Group in British Columbia. The Index Formation, the lowermost formation of the Lardeau Group, rests on the Lower Cambrian Badshot limestone. The Index Formation, which is probably also of Cambrian age, has in its upper part volcanic rocks and if these are correlative with volcanic rocks in the transitional facies, the transitional facies is also lower Paleozoic. However, the northward dying out of the Metaline Formation, indicates that the Lardeau is more probably a facies of the Lower Paleozoic rocks of the Metaline district, which reappears somewhat further changed in the Shuswap Complex.

The Flagstone Mountain sequence of the Northport quadrangle also contain volcanic rocks and accordingly was included in the transitional facies, but these rocks are not older than Devonian and possibly as young as Triassic. The predominance of fine-grained clastic rocks of dark color associates the sequence with the black lutite assemblages and the addition of limestone and volcanic rocks indicates a changing environment. Unfortunately the sequence is in fault contact with Permian and Jurassic rocks of the eugeosynclinal facies, so it is not possible to physically trace the transition.

The rocks in the southwestern part of the Colville-30 minute quadrangle, in particular those in the Echo Valley 7 1/2-minute quadrangle, perhaps best demonstrate the change in facies accomplished by the addition of "foreign" rocks into the persistent background of dark colored pelites. The only fossils in this area are Middle Ordovician graptolites and these occur near the bottom of the section near the Cambrian limestone. Westward and up section from the black slate that contains the graptolites, the section gradually takes on a new character until it is dominated by volcanic rocks, which include at the west side of the quadrangle, limestone lenses containing Permian fossils. Within the section no persistent monolithologic subunits can be traced more than 100 metres. The one exception is a greenstone unit, perhaps 6-0 metres thick. The section is subdivided by separating units on the basis of lithologic variation from the dark lutite background: the same method that was used in subdividing the Grass Mountain sequence. Three major map units are separated, but boundaries are not precise and depend upon exposure of the small lenselike variants that characterize the particular map unit. The basal black slate unit, about 1500 metres thick, has, besides the graptolites in its lower part, small lenses and pods of white to black, medium- to fine-grained quartzite, a little sooty limestone, and tan weathering siltite, an assemblage very similar to that on the south side of Black Rock Canyon in the Deep Creek area, which contains in addition to Middle Ordovician fossils, a questionable Devonian coral. The upper three fourths of the unit is black slate and argillite. Thin beds of graywacke and grit are the variants that set off the overlying unit. The black slate and argillite which forms over 95 percent of this unit closely resembles that in the lower unit. The upper greenstone unit includes both lava flows and fragmental volcanic rocks, as well as minor amounts of thin graywacke beds and black argillite, which also are similar to that of the underlying units.

Admittedly this section is faulted, accordingly, not all the time from Ordovician to Permian may be represented. If any fault or combination of faults represents large scale horizontal displacement, the fault lies within a sequence of rocks deposited in the same environment and does not superimpose rocks of two contrasting environments. Any thurst faults of this nature would have to be older than the folding, because any low angle thrust faults that are younger than the steeply dipping beds would be readily recognized.

Rocks of Pennsylvanian age may be included in the black argillite sequence of the Echo Valley quadrangle, or in the Flagstaff Mountain sequence, but the only fossiliferous Pennsylvanian in the map area, the Mount Roberts Formation (fig. 14) is just north of the International Boundary at Paterson, British Columbia. The upper part of this formation extends westward from Paterson and is what Bowman (1950) mapped as the Churchill Formation in the Orient quadrangle. The Mount Roberts Formation is representative of the eugeosynclinal facies, being replete with mafic flows, volcanic sediments, cherts, and chert pebble conglomerates. The lower part of the formation has, in limestone bioherms, a marine shallow water fauna, and has in dark argillite, plant remains that indicate a near shore environment. Similar conditions of volcanism and shallow water, near shore environments are characteristic of the Pennsylvanian of northwestern Washington (Danner, 1966, p. 67) and central Oregon (Merriam and Berthiaume, 1943, p. 152-153). The environment indicated is back arc, which in Washington would extend across the state. The position of the Rossland Jurassic Formation upon these rocks is the only direct evidence of an Early Mesozoic disturbance in this part of the state.

The Permian rocks of the Kettle Falls area contain the same lithologic elements as the Mount Roberts rocks, both sedimentary and igneous. Mills and Davis (1962, p. 41-43) describe them as including siltstone, argillite, metavolcanics, graywacke, limestone, and quartzite. Fossils of Late Leonardian age and Early Guadalupian time occur in both siltstone and limestone bioherms. The assemblage, except for the quartzite is typically eugoesynclinal and finds its time-facies equivalents westward across the state; examples have been described from the Curlew quadrangle (Parker and Calkins, 1964, p. 27-35), Republic quadrangle (Muessig, 1967, p. 12-28), in the Okanogan River area (Waters and Krauskoff, 1941, p. 1358-1364 and Rinehart and Fox, 1972, p. 19) and from northwestern Washington (Danner, 1966, p. 68-75).

Mid-Paleozoic orogeny

To evaluate the likelihood of a mid-Paleozoic orogeny in northeastern Washington--for which there is no direct evidence--one must examine areas that are structurally related to the area covered by figure 14. The nearest, most closely related, and tectonically most relevant area, is the northwestern extension of the Kootenay arc, the Eastern Fold Belt of British Columbia, where Wheeler (1966, p. 34) and others have reported an Early Mississippian orogeny. Little is known of the type of deformation this orogeny represents. According to Wheeler, the orogeny is not closely dated except in the Cassiar district, 1100 kilometres northwest of the Washington part of the arc, where Lower Mississippian rocks lie with angular unconformity upon Upper Devonian rocks [1/]. In the Cariboo Mountains, 500 kilometres northwest of the International Boundary, early Middle Mississippian rocks unconformably overlie early Paleozoic rocks of the Cariboo Group, thus giving support to the Cassiar event. However, in the Ferguson area, in the northwest end of the Kootenay arc, Fyles and Eastwood (1962, p. 30-33) were uncertain if the conglomerate at the base of the Upper Mississippian Milford Group was lying unconformably upon the Broadview Formation (possibly post-Ordovician), which they considered the upper unit of the Lardeau Group. Wheeler (1968, p. 56-57), having found "variously oriented foliated boulders of Broadview grit" in the Milford conglomerate, supported the interpretation of an unconformity, and questioned Fyles stratigraphic position of the Broadview as the upper member of the Lardeau Group. He suggested that the Broadview, on the

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1/A somewhat different view is expressed by Monger, Souther, and Gabrielse (1972, p. 586-588) who have reinterpreted the Upper Devonian and Mississippian volcanic and ultramafic rocks of the Ominica Crystalline belt (which includes the Kootenay arc) as representing oceanic crust obducted upon a "Proterozoic to Middle Devonian shelf-like assemblage of carbonates and shale." They see the eugeosynclinal rocks of the Great Basin as a similar mass of oceanic crust and draw a parallel in time and style with the Antler orogeny of Nevada. Although a parallel in time exists, a parallel in style is lacking. They present for the Canadian Cordillera a plate tectonic model with tectonic shortening, whereas the current model of the Antler orogeny implies vertical uplift and gravity thrusting.

basis of its lithology, was a western variant of the Windermere, Horsethief Creek Group that had been thrust into its present position. Read (1973, p. 127) who mapped the adjacent Poplar Creek Map area concurs with Wheeler's interpretation of the correlation of the Broadview, the existence of the unconformity, the pre-Milford deformation and the décollement. But since then, Read (1974, p. 29-30) has carefully studied the conglomerate and has extended his mapping into the Ferguson area, where he was able to disprove the thrust concept. Read (personal commun.) would interpret the unconformity and foliated cobbles as products of a mid-Paleozoic deformation that also produced nappes and greenschist metamorphism. He would correlate the phase I structures of Fyles (1964) with this deformation.

In the northern end of the arc the length of the depositional record lost through pre-Mississippian erosion is unknown, as is the intensity of deformation. If the Broadview is indeed the Horsethief Creek, the unconformity represents a loss of time from the late Precambrian (Windermere) to the Mississippian; but if the Broadview is, as Fyles and Read believe, the upper unit of the Lardeau Group--which is free of fossils--the gap begins no earlier than the Late Cambrian and conceivably as late as the Devonian.

Such an orogeny at the present state of our knowledge cannot be extended into Washington; although, as mentioned in a preceeding section, an epeirogenic uplift in the panhandle of Idaho is considered likely. The most likely time of this uplift, late Devonian-Early Mississippian, coincides with the time of the Antler orogeny and accordingly invites the extenison of the "orogeny" and the structures of Nevada into northeastern Washington and southern British Columbia.

In Nevada, the Antler "orogeny" is, in effect the uplift that produced the Roberts Mountain thrust fault, a gravity thrust (Roberts, 1968, p. 106) that slid uplifted eugeosynclinal rocks (the western assemblage) 160 kilometres eastward over miogeosynclinal carbonate rocks (the eastern assemblage). Regional fold belts did not develop, nor was these associated plutonism or even extensive volcanism. The model proposed by Roberts requires a linear uplift considerably greater than 100 miles wide, clastic sediments, both coarse and fine, shed from the uplift, and a fault boundary between the eugeosynclinal and miogeosynclinal facies. The horizontal displacement on the fault is measured from windows in the upper plate and from outliers.

The identification of a thrust fault in Washington similar to the Roberts Mountain thrust fault would have to be on the recognition of juxtaposed contrasting sedimentary facies. Because of the strong Mesozoic folding, however, once horizontal faults are now inclined at high angles and are not readily recognized as former flat faults. The only juxtaposed, contrasting sedimentary assemblages (within the same facies) that have been identified are the Belt and Deer Trail assemblages in the Chewlah-Loon Lake area (Miller and Clark, 1975) and the Leadpoint and Metaline sections in the Deep Creek area (this report). Neither example is a contrast between facies, but a contrast within a facies: both sections in the Deep Creek area are in the miogeosynclinal facies and those in the Chewelah-Loon Lake area are within the distinctive facies of the Belt Supergroup. Outliers and windows in a thrust plate that measure the displacement between eugeosynclinal and miogeosynclinal facies are nonexistent. Both postulated faults are undated, but conceivably could have occurred during the time of the Antler orogeny. The upper plate of the Leadpoint-Metaline thrust is interpreted to have moved westward, opposite to the movement on the Roberts Mountain thrust. The fault in the Chewelah-Loon Lake area is of unknown movement driection, but Miller (personal commun.) favors west over east movement and some movement after the emplacement of the Cretaceous granitic rocks.

One can say without qualification that evidence of major thrust movement of eugeosynclinal rocks eastward over miogeosynclinal rocks is lacking in northeastern Washington; but one cannot say that the two facies have not been telescoped to a minor degree by possible Mississippian thrust faults. This is not inconsistent with the concept that the upper Paleozoic eugeosynclinal deposits are gradational to the black shale facies, which are so closely related to the miogeosynclinal facies. Some faults in the Northport quadrangle that bring Cambrian limestones in contact with rocks of the transitional facies may be folded thrusts, and it is very possible that thrust faults have shortened the transitional suite, however, no evidence has yet been found for a first order thrust with a displacement measured in scores of miles.

But even these, inferred or conceived, thrust faults cannot be assigned to the mid-Paleozoic with assurance. The best case that can be made for a major thrust is the eastern "embayment" of eugeosynclinal rocks that extends into the miogeosynclinal rocks east of the mouth of the Kettle River (see geologic map of the State of Washington, Huntting, 1961). However, among the eugeosynclinal rocks are fossiliferous Permian rocks, which disqualifies the embayment as a thrust sheet of the Antler orogeny.

Before completely rejecting the concept of a major thrust that separates facies, the Great Basin model should be tested against the geology of eastern Washington. According to the model, an extension of the Antler high or "orogenic belt" would have to lie west of the most western autochthonous Cambrian carbonate rocks—that is, west of the Columbia River. A gravity thrust comparable in displacement to the Roberts Mountain thrust requires on the Antler highland an eastward facing slope at least 160 kilometres broad. This slope, before thrusting would have to be underlain by Devonian or older eugeosynclinal rocks. The location of the slope had to fall between two extremes, called here case 1 and case 2.

Case 1 tests the maximum possible eastern limit of the thrust by assuming that the western depositional limit of the miogeosynclinal facies coincided with the westernmost outcrops of Cambrian carbonate rocks that are along the Columbia River--which is an assumption that is obviously not true but necessary for this analysis. Accordingly, the eugeosynclinal rocks would have to be thrust eastward 160 kilometres from the Columbia River, thus placing the toe of the thrust in Montana. It follows that the crest or axis of the uplift, from which the thrust slid, would have to be at least 160 kilometres west of the Cambrian carbonate outcrops, which places it 30 kilometres or more west of the Okanogan River. However, by placing the eastern slope of the high in the area between the Columbia and Okanogan Rivers, conflicts between areas of deposition and erosion arise, for this is the area underlain by the southern extension of the Shuswap Terrane of British Columbia. On trend to the north Reeser (1970, p. 73) tentatively identified rocks in the complex with the Milford Group, which extends down to the Early Mississippian, and Campbell and Okulitch (1974, p. 25-29) traced mildly metamorphosed rocks into the high grade metamorphic zone of the Shuswap Complex. Among these rocks are those of Middle Mississippian age, which according to the model, is a time when the area of the complex should be undergoing erosion.

To test the opposite extreme (case 2) we assume that the present eastern limit of the eugeosynclinal rocks--and western limit of the miogeosynclinal rocks--represent the toe of the thrust and that therefore the crest of the Antler high must now be at least 320 kilometres further west, across the Cascade Mountains and at the edge of the Puget trough. With this hypothesis we are likewise in trouble, because the Antler high in Nevada was a positive area from the Late Devonian until well into the Pennsylvanian, whereas the area of the western Cascade Mountains and the San Juan Islands--where the model locates the Antler high--was a depositional area during most or all of this time (Danner, 1966, p. 66).

Obviously a model with 160 kilometres of eastward thrusting is not possible, but if the displacement is reduced to 80 kilometres, the model appears more reasonable. In case 1 with 80 kilometres displacement the thrust would not extend beyond the Washington-Idaho boundary and the Antler high would fall in the Okanogan Highlands. In case 2, where the toe of the thrust is the eastern limit of the eugeosynclinal rocks, the axis of the high would be west of the Okanogan River in the Methow Valley. Conditions of case 1 requires that the post-thrust Permian rocks rest upon pre-Devonian or older rocks, because when deposition was resumed over the Antler high in the Permian, the thrust sheet that slid eastward had removed thousands of feet of Devonian and older rocks and post-thrust erosion must have removed additional thousands of feet of even older rocks. Although it is only conjecture to assume that erosion would reach down to the Precambrian, or even the Cambrian, it is reasonably certain under the restrictions of this model that no Devonian or Mississippian rocks could be part of the Shuswap Complex, a condition that is contradicted by the correlations that have been made with the Shuswap rocks. The situation is not greatly different in case 2 where the axis of the high is moved 80 kilometres further east into a terrane of Mesozoic and Cenozoic rocks that lie above schists and gneisses of unknown age. Although such a location is permissive, on existing knowledge, it is improbable. From this hypothetical high, pre-Devonian rocks are presumed to have glided eastward until the toe of the thrust came to rest near the site of the Columbia River. This requires that the eugeosynclinal rocks now located west of the miogeosynclinal rocks are pre-Devonian in age and therefore are not part of an essentially autochthonous post-Devonian sequence transitional to the miogeosynclinal rocks as I believe and have tried to establish in this report.

To remonstrate these objections, the proponent of a major west-over-east thrust might suggest that the facies and thrust boundary falls between the Cambrian limestone and the Ordovician black slates and that all the post-Cambrian rocks up to the Permian belong to the upper plate. He might account for the Permian rocks in this plate by postulating movement on a later thrust correlative with the Golconda fault of Nevada. He nevertheless, would have difficulties fitting the Pennsylvanian Mount Roberts Formation into his model and would have to limit the black argillites to an age no younger than Devonian.

In summary, the extension of the Late Devonian-Early Mississippian tectonism of the Great Basin into northeastern Washington is not possible to demonstrate. Such a fault as the Roberts Mountain thrust, and such a positive area as the Antler orogenic belt, conceivably and by a severe stretch of the imagination, may have existed, but the hypothesis finds no support in the geology of the area as it is now known. On the contrary, the sum of evidence in opposition, seems overwhelming, but until rock ages in the Shuswap Complex are more firmly established, the door can only be closed, not locked.

The tectonism that produced the Antler orogeny in the Great Basin, however, may have been expressed in the Kootenay arc in a different way. The postulated thrust fault that brought together the Metaline and Leadpoint sections is younger than Devonian and older than the Mesozoic folding, which began in the Late Triassic. Conceivably it could be Early Mississippian and thus be time equivalent to the Anler orogeny. The Antler high, however, would lie to the east in Idaho and western Montana, perhaps as a rejuvenated Montana. This high might have extended far to the north in British Columbia, possibly as far as the Cassiar region where the angular unconformity of Lower Mississippian rocks upon Upper Devonian rocks developed on the west slope of the uplift.

The Late Paleozoic

The Pennsylvanian and Permian, is recorded in eugeosynclinal rocks, part of an assemblage from Devonian through Jurassic that extends as far west as the San Juan Islands. The lithologic character of this assemblage corresponds to the commonly accepted definition of the eugeosynclinal facies that is based upon the presence of mafic volcanic rocks within a marine clastic sequence, the implication being that the rocks were deposited or ensimatic or oceanic crust. Although volcanic flows appear as early as the Cambrian in the Paleozoic section, no overtone of volcanism dominated the accumulation until the Pennsylvanian, when the volcanoclastic rocks of the Mount Roberts Formation were deposited, which was the first introduction of clastic detritus from the west.

The presence or absence of volcanic rocks in a depositional assemblage, however, is not itself diagnostic, or even indicative, of the type of underlying crust. As recorded in this report, volcanic rocks appear in different environments throughout the section, for example, in the Belt Supergroup and in the miogeosynclinal quartzites of the uppermost Precambrian. Certainly no one would challenge the cratonic nature of the crust underlying these two examples, any more than one would challenge the concept of oceanic crust as the depositional basement for the Franciscan rocks. Fortunately in most marine sections, where volcanic rocks are an important element, the associated rocks contrast with those that accumulated on or at the edge of the continent. In contrast to the quartzites and thick limestones of miogeosynclines, the sandstones of the eugeosyncline are graywackes, which are immature, unsorted and multimeralic, and contain rock fragments, and the uncommon limestones incline towards thinness and limited horizontal extent.

As long as a physical barrier between facies is absent, the boundary must be gradational or intertongued, therefore one assemblage can be "more eugeosynclinal" than another. If one arbitrarily accepts the Franciscan assemblage (Bailey, Irwin, and Jones, 1964) as the ultimate eugeosynclinal assemblage, other assemblages can be ranked accordingly. Top rank demands the presence of the basic graywacke, shale and volcanic rocks, and in addition pillow basalts, banded radiolarian cherts, and alpine type ultramafics, considered direct evidence of oceanic crust. None of the rocks in the study area qualify for this top rating. The most eugeosynclinal, is the Permo-Triassic sequence, which lacks the banded radiolarian cherts, and pillow basalts. The least eugeosynclinal, is the transitional assemblage of largely dark colored argillite with occasional volcanic flows.

Without the evidence of the ophiolites, the interpretation of an underlying oceanic crust comes from the alpine type ultramafics, the serpentines, considered as tectonic samples of a mantle at shallow depth. These occur not only in the study area, but also in the Orient quadrangle to the west, where they are associated with gabbros (Bowman, 1950). In fact, the province of eugeosynclinal rocks that lies along the west border of the arc contains numerous but small serpentine bodies. If it were not for these serpentines, the case for oceanic crust underlying late Paleozoic and early Mesozoic rocks everywhere west of the Columbia River would lose considerable strength.

The problem of the kind of basement of the eugeosynclinal assemblage is involved with the evolution of the crystalline rocks of the Shuswap Complex. The location of the complex, surrounded by eugeosynclinal rocks that apparently once extended across the complex, appears anomalous to the interpretation that the assemblage overlies oceanic crust--because the complex apparently is not of the eugeosynclinal facies. Because it is moderately free of volcanic rocks and contains fairly clean quartzites and limestone, it could be classified as transitional. Jones (1959, p. 11) says, in speaking of the rocks of the Monashee Group (the crystalline rocks of the complex), "Metamorphic rocks that may have been derived from basic lava flows are scarce." The complex being surrounded by the eugeosynclinal assemblage poses special problems in facies relations and also invites questioning the criteria used to designate crustal types.

The structural and temporal relations between the two facies is best presented by a resume of ideas that have developed about the complex. The gneiss terrane of south central British Columbia has long been a subject of controversy, especially the age of the metamorphism. Although the southern part of the complex was mapped by G. M. Dawson (1898) about the turn of the century and described by R. A. Daly (1912) shortly thereafter, it was not until the 1930's that more detailed work was done by Cairnes (1939). Dawson saw the complex as a product of pre-Permian metamorphism and the rocks as definite stratigraphic position, an opinion later shared by Jones (1959), who favored an Archaen age. Cairnes, on the other hand, prescribed a Mesozoic metamorphism, which he unfortunately related to the emplacement of granitic plutons. Most present workers agree that the metamorphism is post-Mississippian, possibly post-Triassic (Hyndman, 1968), and that it is unrelated to plutonic intrusions. Reeser (1970, p. 85) correlated rocks of the complex with the rocks of the Windermere, in particular those of the Horsethief Creek Group and those of the Lower Cambrian. The deeper parts of the complex and the gneiss domes that blister its surface, form an infrastructure (the Monashee Group) that is locally overlain by medium and low grade metamorphic rocks, the suprastructure, which is mainly preserved along the margin of the complex. The separation between high and low grade metamorphic rocks is in places abrupt, along faults; in other places, gradational, the upper, apparently younger, low grade metamorphic rocks, are eugeosynclinal and include rocks such as, "phyllite, greenschist, meta-andesite, and fragmental volcanic rocks as well as marble and carbonate and(or) diopside-bearing quartzite" (Reeser, 1970, p. 77).

The stratigraphic section appears to be, as it is in the arc, a vertical transition from miogeosyncline to eugeosyncline. The eugeosynclinal facies accordingly overlaps the miogeosynclinal facies for more than 150 kilometres. An alternate explanation is to interpret the facies boundary as a thrust fault. Such a solution, however, is in conflict with the problems of the serpentines, because the serpentine not only intrude the supracrustal rocks, but at least in one place (Jones, 1959, p. 157) intrude the Monashee Group as well.

Assuming that Reeser's (1970, p. 85) correlations of rocks of the complex with rocks of the arc are correct and that an upward gradation from miogeosynclinal to eugeosynclinal rocks exists, one can postulate that the characteristics of a depositional facies at or near the continental margin is less influenced by the underlying crust than by the tectonic state of the basement. In an Atlantic type continental margin, the join between continental and oceanic crusts is a stable relation; the two crusts are welded together and act as a unit. A potential difference between crusts, could be that the load strength of the oceanic crust is less, therefore it subsides more readily when loaded and might tend to crack up and permit magma or serpentine to leak up from the upper mantle. Such is the situation envisioned for the continental margin in northern Washington during the Paleozoic: a stable join between continent and ocean forming a single plate--when one part moved the other part moved with it. The sedimentary prism of the miogeosyncline extended across the join and over the oceanic plate. As more and more sediments accumulated, the prism extended further seaward. In shape it may have been that of a double prism, a miogeosyncline or that of a single prism, the miogeocline of Dietz and Holden (1967).

One can easily visualize the extension of miogeosynclinal rocks over oceanic crust under the above described conditions, but it is less easy to visualize what causes a vertical change in section from miogeosynclinal rocks to eugeosynclinal rocks. The hope of solution is perhaps less bleak than it appears; because the assemblage of stratified rocks that form on the sea floor are not determined by the type of sea floor, but by the tectonic processes that go on above and below that floor. Current direction and velocity, water depth, temperature, salinity, etc. have an important effect, particularly upon the products of the miogeosynclinal environment but a much lesser effect upon the eugeosynclinal environment, which is largely determined by the kind and quality of igneous activity and tectonism originating below the sea floor. Whether or not an oceanic crust is covered by miogeosynclinal deposits, it should react as oceanic crust, whenever and wherever its weld to the continent is broken and the two plates begin to move towards each other and overlap. But long before the actual break occurred and subduction took palce, the oceanic crust was bowed down and weakened by the ever increasing load placed upon it, tensional fractures opened, magma was tapped and rose to become flows on the sea floor and volcanic cones that grew to become islands. Silica concentration increased, radiolarian and other siliceous organisms thrived, cherts formed. The volcanic islands were eroded and the debris mixed with that derived from continental sources, probably transported by turbidity currents. A eugeosynclinal assemblage that is vertically gradational to an underlying miogeosynclinal assemblage has been created.

What is described above resembles in many respect the "Atlantic Phase of ocean floor spreading and expansion" (Dewey, 1969, p. 189-197) whereby continent and ocean are not interacting and accordingly contrast with the "Pacific Phase" where an orogen develops when a plate is cleaved and divided. Both phases can be recognized in the geologic history of eastern Washington and one can be dated as occurring before and the other after the beginning of folding in the Late Triassic. By more than coincidence, this approximately coincides with the Early Jurassic opening of the Atlantic Ocean and movement of the North American Plate Pacificward. Being a rigid body, the North American plate immediately reacted along the Pacific margin--which had previously been weakened by overloading--the junction of continental and oceanic crusts cleaved and lithosphere was lost at the junction by subduction of the Oceanic plate. And so the Mesozoic orogeny was initiated.

The Mesozoic, an Orogenic Era

To describe the Mesozoic as an orogenic era is not to imply that the Mesozoic was one continuous orogeny from the beginning of the Triassic to the close of the Cretaceous. The title is deserved because the one and only tectonic episode in northeastern Washington that involved folding, faulting, volcanism, and plutonism was in the Mesozoic. It could also be called the <u>era of compression</u> in contrast with the extension of the Cenozoic.

In the Kootenay arc the Mesozoic record is of such variable completness, both geographic and chronologic, that it is doubtful if one can generalize for the whole of the arc. At no place in Washington east of the Methow-Pasayten graben have Upper Jurassic and Lower Cretaceous rocks been identified. Lower Triassic fossils have been reported from only one place (Kuenzi, p. 88-89). Mesozoic sedimentary and volcanic rocks occur only on the western side of the arc and--except the Sophie Mountain Conglomerate, of Late(?) Cretaceous age--belong, or are closely related to, the eugeosynclinal facies. North of the International Boundary, Upper Triassic and Lower Jurassic rocks are moderately common. The northern part of the arc may have had a different depositional history than the southern part, or a different erosional history. For example, the upper Triassic rocks, that occur in abundance not only to the north in British Columbia (little, 1960, Map 1090A) but to the east in the Curlew quadrangle (Parker and Calkins, 1964, p. 27-35) have not been identified in northeastern Washington, but they may have been deposited and later eroded rather than never deposited. The time needed for such an erosion could be bracketed by the Pennsylvanian-Jurassic unconformity recorded by Little (1960, p. 48) at Paterson, just north of the Boundary. Further north an unconformity between two Triassic units, the Kaslo Group and the Slocan Group, although of uncertain age, indicates another, but older erosional period. There may be more.

The possibility that some rock units may be misdated is ever present. As mentioned in the section of the black shale facies, the Flagstaff Mountain sequence conceivably could be Triassic in age. No direct evidence supports this view, but if true, my concept of the magnitude of displacement on the fault that separates the Flagstaff Mountain rocks from the Cambrian carbonate rocks needs serious revision, a revision that eliminates any possibility that the fault could be a structure of the Antler orogeny.

Although no specific chronology of Mesozoic events can be made for the arc, a few broad generalizations outline the tectonic evolution. The eugeosynclinal facies of the Permian continued on through the Triassic and through the first half of the Jurassic. Folding appears to have begun in the Late Triassic and was repeated in the Late Jurassic, or it may have been restricted to the Late Jurassic, depending upon the evidence one accepts. The Jurassic folding began after the Lower Jurassic Rossland Group was deposited. By Late Cretaceous the eugeosyncline had retreated west of Vancouver Island and continental conglomerates were being deposited just west of the arc. The compressional phase of the Mesozoic probably ended by the Early Cretaceous. Block or extensional faulting probably began during the Late Cretaceous and continued through the Eocene.

Time and intensity of folding

In northeastern Washington in the southern part of the arc, because of the paucity of layered Mesozoic rocks, we are less concerned with depositional history and more concerned with the tectonic and plutonic history. To interpret and date the folding, the geology both north and south of the International Boundary must be examined. The arc rocks could not have been folded before the Late Triassic because Triassic rocks in apparent conformity with older rocks are folded in arc trends in the west Nelson Map area of British Columbia (little, 1960, Map 1090A). An upper age limit on the earlier folding is fixed in British Columbia by the cross cutting relations of the 162 million year old Toby stock in the East Lardeau map area of British Columbia (Reeser, 1973, p. 93) and in Washington by the 200 million year old Flowery Trail pluton in the Chewelah Mountain quadrangle (Miller and Clark, 1975). The intrusion of the Toby stock roughly coincides with the end of Jurassic deposition, whereas the Flowery Trail pluton was intruded slightly before the beginning of the Jurassic. If we reject the 200 million year age of the Flowery Trail stock, we simplify the tectonic history by recognizing only one broad period of folding, however, the evidence favors accepting the pre-200 m.y. folding, judgement being influenced by the well established Triassic age of folds in the Okanogan region (Rinehart and Fox, 1972, p. 72-73) and by the knowledge that dual folding had been recognized much earlier north of the boundary by Fyles and Hewlett (1959, p. 47). From here on, I shall refer to the Triassic folding as event 1 and the Jurassic folding of lower Jurassic rocks as event 2. The selection of <u>event</u> instead of <u>phase</u> avoids confusion with the terminology of Fyles (1964, p. 41) and others and the inference of unintentional correlation.

Folding in the southern half of the arc ended before mid-Cretaceous, because the latest compressional movement, the cross fold event, is older than the 100 m.y. old Spirit pluton. All folding, therefore can be bracketed between the Late Triassic and the Mid-Cretaceous and it is reasonably certain that the event 1 folding was not much before the Jurassic and the event 2 folding was no older than Late Jurassic. A possible third and final compressive event, the kink folding, may be independent and early Cretaceous, or may be closely related to, and contemporaneous with, event 2. Dating these tectonic events depends mainly upon the radiometric ages of cross-cutting intrusive igneous rocks, which, if in error, could destroy the concept of two separate foldings.

Multiphase folding is much better developed in the northern part of the arc than in the southern part. In British Columbia, Fyles (1964) and later Crosby (1968), and Read (1973) have mapped, both macro and meso structures, showing two phases of uniaxial folding and one or more phases of cross-folding that agree in orientation and relative age but not in time scale age. Although Mills (1973) has described refolded folds in the core of the Columbia anticline in the Northport quadrangle, Washington, the concept of two phases of folding along the same axis is not as firmly established in the southern part of the arc as it is in Canada.

In the Kootenay arc south of the boundary, I have observed and measured refolded folds (meso), but have found them to be uncommon and almost entirely restricted to incompetent beds where flowage was obvious and, in a few places, where flowage clearly occurred before consolidation. The preconsolidation folds agree in trend with the major folds, which is in agreement with the concept that trends of major folds are commonly influenced by and inherited from trends of depositional features--mainly stratigraphic thickness--and thus parallel strand lines and facies boundaries, which in turn contour the depositional slope. This is not to infer that the folds measured in British Columbia are depositional features--which they obviously are not--but to point up the contrast between the abundances of refolded beds north and south of the boundary.

The relative abundances or even presence, of refolded folds may be an indication of depth of structural development, differences in tectonic environment, or differences in structural style. R. B. Campbell (1970, p. 67-72) has demonstrated quite convincingly that the complexity of structure and the degree of metamorphism in the Cariboo Mountains of British Columbia, 300 miles north of the International Boundary is a function of the depth of burial. The deepest exposed structures are isoclinal folds in rocks of staurolite-kyanite grade that pass upward through simpler fanfolds into concentrically folded unmetamorphosed rocks broken by normal faults. Such extreme contrast is not present in Washington--probably partly because of less structural relief--but what may be a deeper structure, the Columbia anticline, does have the tightest and most complex folds. The tectonic complexity that Mills (1973) found in the Columbia anticline probably results from the influence of depth of deformation and the reorinetation of fault-bounded blocks during the cross fold event.

The influence of depth on structural style perhaps explains an anomaly. Folding in the Kootenay arc of British Columbia began in the late Triassic and ended by or before mid-Cretaceous, but the Front Range thrust faults of Alberta transported unfolded rocks eastward from the longitude of the Arc--or at least from that of the anticlinorium--during the Late Cretaceous or early Cenozoic (Bally, Gordy, and Stewart, 1966, p. 370). If the displacement and timing of the thrusts is correct, the folds of the arc and anticlinorium passed upward into unfolded and gently warped beds, a conclusion that supports Campbells' (1970) contention. To add to the puzzle, even before thrusting, the Nelson batholith had been deroofed and granitic pebbles dated between 113 and 174 million years (Norris and others, 1965; see Bally, p. 369) were deposited in a Lower Cretaceous (Albion) conglomerate at the east front of the Rocky Mountains.

An example of deep structures that may underlie the arc occur 40 km west northwest of Northport in the Grand Forks area of British Columbia, where Preto (1970, p. 69-74) describes two closely related phases of deformation in the Shuswap complex; the first and strongest produced folds with trends slightly north of west and gentle plunges to east or west, and the second phase, of minor importance, produced folds of northerly to northeasterly trends and east over west transport. The second phase he relates to the emplacement of the Nelson batholith; the first he regards as a trend "not peculiar to the Grand Forks complex" but "found throughout the Shuswap Terrane" and a trend nearly normal to the regional trend of the Shuswap complex. This east-west trend is also strongly discordant to the trend of the Kootenay arc.

The trends of compressional structures of the cross fold event are so close to the near east-west trend characteristic of the Shuswap Terrane (see fig. 28) that a correlation seems logical. If the correlation is correct, the earliest identified folding in the Grand Forks area particularly in the crystalline rocks of the Shuswap complex (Reeser, 1970, and Wheeler, 1966) is the youngest compressive event in the southern Kootenay arc.

Preto (1970, p. 70), however, provides for the possible existence of a deformation in the Shuswap Terrane that is older than that which produced the east-west folds, by pointing out that many areas of infrastructural deformation similar to that of the Grand Forks area have "a long and complex structural evolution, the beginning of which was marked by a period of isoclinal folding accompanied by metamorphism." Although Preto failed to find any evidence of isoclinal folding, he realized that such structure is possible and notes that metamorphic layering developed and granitic sills were emplaced before the initial folding, which are observations that suggest slippage in a horizontal plane and softening of the rocks before folding. This could coincide with the earliest folding in the arc and with the Triassic folding in the Okanogan area. An important point is that regional metamorphism of amphibolite grade probably went on in the Shuswap Terrane at the same time or before the arc was being folded.

Differences between tectonic environments in the northern and southern parts of the arc are more difficult to evaluate. The buffering by rigid masses--if it existed--could affect one part of the arc differently than the other. Crosby (1968, p. 78) and Ross, J. V. (1970, p. 64) interpret the arc as folded against "the more rigid Purcell basement," but fail to explain why the Purcell (Belt) rocks should be more rigid than overlying rocks, especially since Precambrian folding of Belt-Purcell rocks was very gentle. The shape of the Belt prism[1/] probably was a strong influence on the arcuate

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[1/] Prism is used in this report in a loose sense for an elongate assemblage of sedimentary rocks inferred to be wedgeshaped or double wedgeshaped in cross section.

trend of the arc, but because arc and ànticlinorium were folded together, it could not have acted as a rigid basement and, if it had, the buffering effect should be uniform along the entire length of the arc.

The third possible explanation of the scarcity of refolded folds in the southern half of the arc is a variation in structural style. The major folds of the north are absent in the southernmost part of the arc in the Chewelah 30-minute quadrangle. Possibly compressive stress was relieved by movement on deep rooted thrust faults and not by folding.

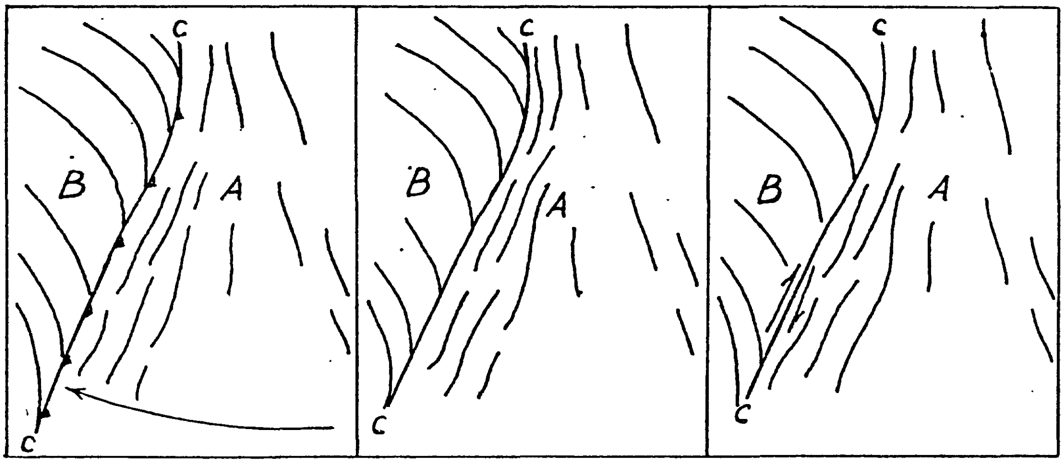
The Kootenay arc as a foldbelt

The ancestral Kootenay arc, which was born in Proterozoic time as a sedimentary prism and continued to grow during the Paleozoic, became a foldbelt in the Mesozoic (see fig. 27A). As now seen, the arc is a bend in the northwesterly trending Eastern Fold Belt of British Columbia, but until the Mesozoic it was nothing more than an arcuate depositional basin, reflecting the contour of the continental margin, which was located an unknown distance west of Idaho. During the Paleozoic, sediments accumulated to great thicknesses in the basin--or trough as it was becoming--and the accumulation weakened and accelerated downwarping of the underlying crust. In the Mesozoic this weakened crust and overlying prism of sediments was compressed into a foldbelt that preserved the trend of the Proterozoic basin.

By definition the eastern boundary of the arc is the Purcell Trench and the Precambrian rocks to the east belong to the Belt-Purcell anticlinorium. This boundary although convenient, disregards stratigraphic distinction, but is tenable because it separates tight folds from open folds. Near the trench, fold trends tend to parallel arc trends. To the east in Montana, away from the structural confusion around the Osburn and Hope faults, the folds of the anticlinorium trend about N.20°W. The northeast and northwest fold trends converge in Canada, giving the anticlinorium a northwestward plunge. The axis of the anticlinorium, as taken from the Geologic Map of British Columbia (Little, 1961), also has an arcuate form similar to that of the Kootenay arc. The eastern limit of the anticlinorium disregards the trend of the axis and continues southeastward into Montana without a bend. In a sense, the anticlinorium governs the arc, or stated differently, the folds of the arc are molded against the anticlinorium. The folds of the anticlinorium are inferred to be the same age as those in the arc--but conceivably could be earlier. Porbably the last stage of Mesozoic folding did not refold the folds of the anticlinorium.

The western border of the Kootenay arc is a compromise between a border defined by stratigraphy and one defined by structure. In a gross manner, the rocks of the arc are miogeosynclinal and those to the west in the Pacific Border province are eugeosynclinal, but because of the transitional facies that separates the miogeosynclinal from the eugeosynclinal, no precise boundary can be drawn on stratigraphic boundaries. Structurally the arc, having predominant northeast fold axes, contrasts with the northwest trends of the Pacific Border province, but the junction of these trends is not well defined. If one through-going fault separated the two, division would be simple, but such a fault has not been found, although theoretically it should exist in broken parts. Not even a hypothetical fault can be drawn that would make a clean division between facies and between fold trends. This western boundary contrasts with the eastern boundary where fold trends gradually change from northeasterly to northwesterly across the breadth of the anticlinorium. At the western boundary, trends of the Pacific Border province appear to swing into the northeast trends of the arc (fig. 28). This feature is particularly well developed in the Salmo Map area of British Columbia and north of the junction of the Columbia and Spokane Rivers.

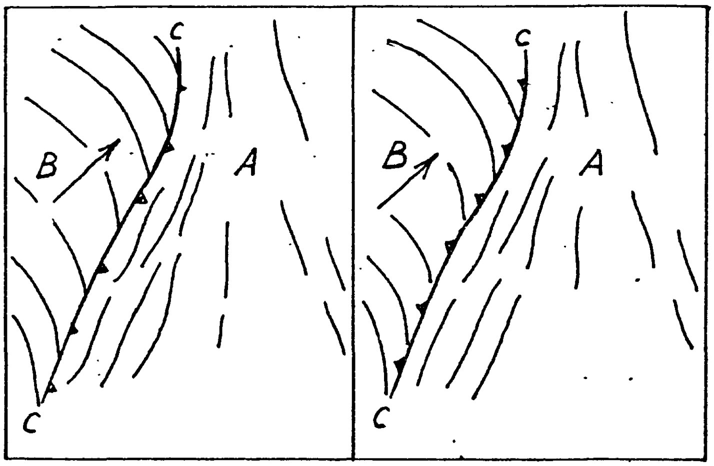
An understanding of these converging trends is of major importance, but is made difficult by the lack of detailed mapping in critical areas and by large areas engulfed by plutonic rocks or buried by glacial deposits. In lieu of direct proof, I approach the problem with models representing several interpretations (fig. 37). In Model 1 contemporaneous folding in the arc (A) and Pacific Border province (B) is postulated, requiring westward thrusting, which is difficult to equate with a single stress direction. Another objection is that this model requires oroclinal bending, which is an unreasonable hypothesis. Model 2 where the B folds are earlier than the A folds, requires the refolding of the B folds into A folds. There is no evidence to support this. Model 3 requires the rocks of B to be younger than the folds of A and also requires right lateral movement on fault c. Arc folding would have to be pre-Late Paleozoic and if the correlation of Shuswap rocks is correct, Precambrian, which is obviously impossible. Model 4, the "subduction model" has the arc folds early and the B rocks thrust under the A rocks. Model 5, the "obduction model," is a variant of model 4, with B thrust over A. In models 4 and 5 the lack of windows and outliers, indicates that any thrust plane between A and B is steep and of no great horizontal displacement. In summary, models 1, 2, and 3 are not applicable; 4 and 5 are theoretically possible, but are not demonstrable from the data at hand.



1. Contemporaneous folds. Requires oroclinal bending and thrusting along C of arc (A) over Pacific Border province (B).

2. B folded before A. Requires that the B folds be refolded into A folds.

3. A folded before B rocks deposited. Strike-slip movement on C confines second folding to B.



4. The "subduction" model. A is early, B is thrust beneath A.

5. The "obduction" model. A is early, B is thrust over A.

Figure 37. -- Idealized diagrammatic sketches of several hypothetical origins of the arc folds.

The construction and analysis of these models has not solved the origin of the two contrasting tectonic trends, but is has pointed up several facts that appear inescapbale. These are: (1) the join between the arc and the Pacific Border province must be a fault or a system of faults, (2) any thrust fault that might separate the two fold systems is not a flat fault, (3) strike-slip right lateral movement at the join cannot be entirely eliminated, and (4) the arc buffered the northwest trending folds and bent them towards parallelims with are trends. An additional fact to keep in mind is that fold trends in the northwestward continuation of the arc, the Eastern Fold belt of British Columbia, are parallel to fold trends in the adjacent Pacific Border province. Whether or not the same two periods of folding occur here as in the arc is an open question (Read, 1973, p. 104-105).

The above conclusions lead one to believe that depth and type of basement had an important influence on the form of the arc and the junction between arc and Pacific Border province. Before the first folding (Late Triassic) the arcuate form of the continental margin controlled the shape of the depositional basin, and this in turn controlled the trend of the first folds. By the time of the second folding, the early folded arc rocks were a semi-rigid part of the continental margin--at least rigid enough to buffer the Jurassic and older rocks that were pushed against them. To produce these two different trends, 70° apart from the same direction of transport, is no more difficult than to produce the northeasterly and northwesterly trends of the arc from a single stress direction.

One might ask, "why did not the second folding reproduce folds of arc trends in the same way it reproduced the axial trends of the early folds in the Eastern Fold Belt northwest of the arc?" Suggested answers are: (1) a change in stress direction, or (2) a difference in basement. A change in stress direction is, at best, only a partial answer because the stress that produced the arc, produced folds that range through 75° of trend from the north to the south end of the arc, thus suggesting there is a wide latitude between stress direction and fold axes. The effect of the basement could be considerable. The first folds are interpreted to occur along the edge of the continent; the second are mainly in rocks believed to have been deposited on oceanic crust. The second folding, however, also effected the first folds, particularly in the northern part of the arc. The southern part of the arc may have escaped much of the compressional thrust of the second folding by release of stress along the fault that separates the arc from the Pacific Border province.

Kink folds and basement structure

Although most workers in the Kootenay arc and the Shuswap complex recognize at least two phases of folding, there is no agreement on the correlation of phases from area to area within the arc nor between arc and complex. In a section above, uncertainty was expressed as to whether the kink fold and related structures of the cross-fold event were formed in a final third period of folding, or whether they were contemporaneous with or a late phase of the second period of folding. Such problems as this can only be resolved by accurate dating of both the second event and the kink folds--which has not been done. Even if one compares the same structural levels, correlations based upon superimposed minor structures if fraught with danger.

The two areas, arc and complex, although both Mesozoic in structural age, are greatly different in structural style. The arc is a linear feature, the complex is--in a relative sense--a planar feature. Macrofolds of the

Shuswap complex, although varying in style with depth and age, are typically isoclinal with attenuated limbs and gently dipping axial planes; folds of the arc range from tight overturned refolded folds with axial planes at medium to high angles in the northern part of the arc, to broad open folds in the central part, and to what appears to be a faulted homoclinal structure in the southern part.

As exposed, the complex is basically a high grade metamorphic terrane, the arc is not, although it has within it areas of high grade metamorphic rocks, which in their deeper parts have the recumbent folds of the complex. This leads one to believe that Shuswap rocks may be the basement for the area that extends eastward from the Okanogan River to the Purcell Trench and that the gneiss domes are blisters of mobilized infrastructure extending upwards from this "basement" that was born of Mesozoic metamorphism.

If the arc is underlain by the Shuswap complex, and the structures of arc and complex are Mesozoic--conclusions I accept--it is logical to assume that the domal structures of the complex should be reflected in the overlying arc rocks. Paradoxically, they do not appear in the suprastructure, a fact which leads to the conclusion that either the areas of high grade regional metamorphism, located east of the complex are not related to the complex--which is unreasonable--or else the gneiss domes as structures, do not extend into the suprastructure, the uplift being compensated by horizontal thinning. As products of viscous flow, the gneiss domes demonstrate that the complex is also the product of partial melting and flow and, during the Mesozoic and perhaps early Cenozoic, was the upper part of a mobilized basement in a region of high heat flow.

The fold structures of both infrastructural and suprastructural levels of the Shuswap complex of British Columbia can be tentatively correlated with the cross-fold event of the arc in Washington. The correlation is based upon fold trends: most trends of the complex fall between N.75°W. and east-west, the cross-folds of the arc range between east-west and N.80°E. The discrepancy between these two trends can be explained by the differences in prefold structures of the two provinces. The folds of the complex that are being correlated are the early folds, which are the isoclinal recumbent folds of the infrastructure and the open folds of the suprastructure. These folds developed from horizontal strata, whereas the structures of the cross fold event of the Kootenay arc are redeformation of beds previously folded in arc trends. The tear faults associated with the thrust faults of the Columbia anticline help cement the tie between the folds of the complex and those of the cross fold event. Their north to N.10°E. trend is approximately normal to the fold axes of the complex and probably represents the direction of regional compression at that time.

One can correlate the cross fold event with structures in the British Columbia part of the arc. In the Salmo lead-zinc area as described by Fyles (1959, p. 55) the east-west to east northeast striking thrust faults are clearly correlative with the cross fold event. In the Kootenay Lake area Crosby (1968, p. 59) describes Phase III structures as cross folds, which in trend and timing correspond with those in the Columbia anticline. Farther north in the Duncan Lake area (Fyles, 1959, p. 46-47) describes no Phase III structures but points out minor folds having northwesterly plunging axes, which are clearly later than the northerly trending arc folds.

Plate boundaries

To locate the boundary of the continental plate during the Mesozoic in northeastern Washington, is little more than speculation. One can make tentative locations, if one assumes--as I have--that both continental and oceanic crusts underlay the area, that the join between crustal types was frozen during the Paleozoic and decoupled during the Mesozoic, and that the trace of a fault zone between subducting oceanic plate and a continental plate would not outcrop, but would be masked by eugeosynclinal rocks, which although formed on the oceanic crust were crumpled and pushed up on the continental plate. The details of such a model are given in a later section of this report.

If one assumes that granitic rocks along continental margins are related to plate boundary interaction and emplaced in the continental plate, the trace of the subduction zone must, therefore, lie seaward of the outermost granitic rocks. By such an assumption all granitic plutons from the San Juan Islands to the panhandle of Idaho are assigned to the continental plate, but were intruded under a depositional terrane that extends from the eugeosyncline, through the miogeosyncline, to the shelf. On the other hand, if one assumes, as was done in a section above, that ultramafic rocks emplaced in a eugeosynclinal assemblage are indicators of an underlying oceanic crust, then the boundary between oceanic and continental crust lies near the course of the Columbia River. The model of the pre-Mesozoic tectonic environment described above depicts Late Precambrian and Cambrian sedimentary sequences extending westward from the miogeosyncline over the continental margin and on to a quiscent oceanic crust. Tectonics during the Paleozoic produced essentially unfolded structures, a situation that changed drastically during the Late Triassic, when the first folds of the arc were formed.

Such a drastic change in structural habit suggests a drastic change in tectonic relations between the two crusts. One might speculate that the junction between continental and oceanic crust was severed and the freed Pacific plate was moved beneath the North American plate. One can theorize that it was differential movement between the two crusts--perhaps initiated by the opening of the Atlantic Ocean--that broke the crustal bond, thus creating two independent plates that moved towards each other with crumping of their margins.

The trace of such a subduction zone would probably never intersect the surface (fig. 38), but would be buried beneath the supracrustal rocks that were sliced from the subducting plate, crumpled and thrust--or obducted--up and onto the continental plate. The serpentine that identified the rocks deposited on oceanic crust, although penetrating Jurassic rocks, could have traveled upwards, cold and diaperically, from an earlier home in the oceanic basement. Because of the compressional jumbling of rocks and later extensional faulting, no single fault plane--if it ever existed--can be identified.

Whether or not a thrust fault obducting the sedimentary prism over the continent existed, its hypothesis adds strength and reason to the causal association between folding and plate subduction. In fact, it adds vigor to the postulation of a subduction zone, by not requiring surficial expression of the join and by providing a mechanism whereby waste from the continent can be returned to the continent. In effect, this recycling of continent derived sediments is a process promoting the conservation of continental mass.

Figure 38.-- Sketch illustrating why the plate junction of a subduction zone is unlikely to be exposed.

Uplift and erosion, the end of an era

By mid-Cretaceous the compressional regime of Mesozoic tectonism had ended; uplift and gravitational tectonics took over. Knowledge of the Early Cretaceous is no greater than that of the late Jurassic, other than that plutonism was active. The lost depositional record, from mid-Jurassic to Late Cretaceous, spans the change from marine to continental conditions. By the Late Cretaceous marine deposition was over and erosion had cut deep into the crust exposing the high grade metamorphic rocks of the Shuswap complex. The only direct clues to the closing history of the Mesozoic are in the Sophie Mountain Conglomerate and the Cretaceous granitic rocks.

The Sophie Mountain Formation

The description of the Sophie Mountain Formation given in this report, particularly that on size, variety, and uncertain provenance of the clasts, indicates a topography of high relief and rapid transportation from a nearby provenances. The eugeosynclinal character of some clasts, such as the radiolarian cherts, and the miogeosynclinal character of other clasts, such as the quartzite, indicate either two separate provenances or tectonic positioning of facies into a single provenance. Tectonic positioning is the preferred explanation, a conclusion based mainly upon the large quartzite boulders which could not have been moved more than a few miles by normal stream transport. The presence of these boulders in a conglomerate that rests on an eroded eugeosynclinal terrane, suggests derivation from the front of a thrust plate now completely eroded. The large size of the boulders also suggests that high angle normal faults produced relief locally great enough to insure rapid transportation. It is highly probable that these extension faults were initiated in the early Late Cretaceous.

The nearest Cretaceous sedimentary rocks are about 250 kilometres to the west in the Methow-Pasayten graben in north-central Washington and adjacent British Columbia. These rocks probably indicate the general conditions that existed in eastern and central Washington during the Cretaceous. Barksdale (1960) and Pierson, T. C. (1973) describe a sequence more than 5,500 metres thick of clastic sedimentary rocks, predominantly well sorted sandstones and minor amounts of shale, conglomerate, and volcanic rocks. The Early Cretaceous rocks predominantly arkose, include a few cobble and boulder beds and contain marine fossils. The Late Cretaceous rocks, resting unconformably on the Early Cretaceous, represent a vertical transition to nonmarine and finally to a volcanic environment. The clasts indicate provenance was both to the east and to the west, a conclusion that suggests an ancestral graben or a narrow embayment, perhaps opening southward into the gulf of the Cretaceous sea that extended into eastern Oregon. The Sophie Mountain Conglomerate, although separated during its deposition from the Methow-Pasayten embayment by an upland, was similarly fault controlled. Whether other embayments of the Early Cretaceous sea extended into northeastern Washington and were removed by mid-Cretaceous erosion before the deposition of nonmarine conglomerate cannot be determined, but the presence of an unconformity within the Methow-Pasayten section supports this suggestion.

The rocks in the Methow-Pasayten graben were folded before 86 million years ago (Staaz and others, 1971, p. 32), the age of a cross-cutting igneous intrusion, which is younger than any known folding in the Kootenay arc. Because the rocks of the graben were folded before 86 m.y., very little of the Late Cretaceous can be represented in the section and if this folding is correlative with the second period of folding to the east, which was sometime between 100 m.y. and 165 m.y. only lower Cretaceous rocks can be present.

I am inclined to assume that folding occurred sometime after 100 m.y. ago and is therefore not correlative with but younger than the second and youngest folding in the arc. This conclusion is consistent with the general decrease in age of folding from east to west. For example, all folding in the arc was over before the Eocene, whereas in the Methow-Pasayten graben Cenozoic volcanic rocks, possibly Eocene, are folded (Staatz and others, 1971, pl. 1). Further south near Yakima (Huntting and others, 1961) in what would be structurally west of the Methow-Pasayten graben, the Miocene Columbia lavas are folded along northwesterly trends. There is no indication, however, that the extensional faulting followed this regional westward progression in time.

Plutonism and the subduction model

The granitic rocks of northeastern Washington range from the syenite rocks of the Coryell batholith to quartz dioritic phases of granodiorite plutons. They fall into few age groups, approximately 50 m.y., 100 m.y., 170 m.y., and 200 m.y., an arrangement that suggests cyclic plutonism. This suggestion is not confirmed if the sampling of plutons extends into British Columbia, where gaps between age groups are filled and more continuous spectrum of plutonism results, nor is it representative for all of the State because the Cascade Range of western Washington has plutons of Miocene age. The filling of gaps in the age sequence by expanding the sample area horizontally might also be accomplished by expanding vertically; that is, a sample plane 5 kilometres below the present surface might include 75 and 150 m.y. old rocks.

Plate tectonic theorists relate foldbelts, batholiths, and other phenomena as expressions of plate interference at continental margins. Accordingly, the structural model for the tectonic evolution of a terrane should be different, if plutonism was continuous rather than episodic. The continuous generation of magma for 200 million years would suggest that the conditions that prompted magma generation--plate movement--continuous. One cannot, however, as readily assume the reverse, that cyclic plate motion is indicated by cyclic magma generation or cyclic folding, because continuous application of stress can, and does, produce cyclic events; for example, continuous stress applied to the San Andreas fault is episodically relieved, suddenly and violently. Such reasoning leads to the conclusion that it is unnecessary to postulate an interruption in plate movement pattern until contact with another plate makes it mandatory. In theory, both plutonism and folding can be cyclic without change or interruption in spreading rate.

In a preceding section, I suggested that folding was cyclic and was followed by plutonism, but I am unable to demonstrate that the seeming cycles of folding and plutonism are not near surface expressions of deepseated continuous flow that is capable of storing energy that can be released episodically. Whether cyclic or continuous, by the end of the Eocene, tectonism and plutonism had welded a granitic basement within the lower part of the prism of sedimentary and volcanic rocks that were originally deposited on the oceanic crust of the Pacific Border province. This concept of continental expansion by sedimentary extension, although not supported by the age distribution of plutons, is supported by the westward decrease in the age of folding. Except for the youngest granitic rocks, which are restricted to the western Cascade Mountains, and the 200 m.y. old rocks that are concentrated along the eastern side of the Cascade Mountains and rare examples in the Kootenay arc, the age distribution of plutons is random. The 100 m.y. old Cretaceous plutons are widespread from the western border of the anticlinorium, across the arc and through out the Pacific Border province. They, as well as the Eocene and Triassic plutons are compositionally unrelated to any age pattern.

Two exceptions to the above generalization that emphasizes the lack of pattern in pluton distribution, are in themselves generalizations that contradict each other. The first is that the quartz diorite line of Moore (1959), which separates potash-rich from potash-poor rocks, can be drawn through the Cascade Mountains of northern Washington, nevertheless, the line has no respect for age of plutons, except--and this is the second exception--that all Miocene plutons fall west of the line, and these are plutons that are theoretically and paradoxically too high in potash for their location.

Further pursuance of this subject enters the highly controversial and speculative field of origin of granites, which although too involved a problem to discuss in depth, cannot be entirely side stepped. Several significant facts concerning granite rocks in Washington worthy of emphasis are: (1) lack of systematic distribution of plutons by age, (2) lack of systematic change in composition by age, and (3) lack of systematic relation to tectonic environments. From these negative relations one can infer that the granitic rocks are not the product of plate subduction such as the model proposed by Dickinson (1970, p. 841-844), which relates generation of granite magmas to a steeply dipping plate boundary where subsidence of the ocean plate induces partial melting and fractionation of more silicic and alkalic magmas in a space-time-composition order. Such a model demands systematic change in age and composition of plutons in the direction of plate movement. The unsystematic distribution of plutons nevertheless, is a valuable clue to the tectonic evolution of the region in the Mesozoic and Cenozoic.

The pre-Mesozoic rocks from the Cascade Range to the Purcell anticlinorium were not folded until the Triassic at the earliest, when horizontal strata were deformed into the low angle isoclinal folds and metamorphosed into the high grade metamorphic rocks of the Shuswap Terrane (Reeser, 1970, p. 83), and of the panhandle of Idaho (Crosby, 1968, p. 72). It requires no stretch of the imagination to visualize these high grade metamorphic rocks grading downward into a zone of partial melting from which granitic magma was derived.

Reeser (1970, p. 86) points out that stratigraphically correlative rocks of comparable metamorphic grade in the Purcell-Selkirk Mountains and Shuswap Terrane lie at the same topographic level. In a regional sense, such a surface of melting--which parallels metamorphic isograds--would represent a planar surface, but in detail the surface would be undulating with sharp ridges and troughs. For example, in the Kootenay Lake area, not only the Lower Cambrian rocks, but the Mississippian-Triassic Milford Group rocks are metamorphosed to sillimanite grade (Crosby, 1968, pl. 1). The isograds define a metamorphic "anticline" that parallels the structural trend. From this metamorphic high, metamorphic grade decreases eastwards into the background muscovite-chlorite assemblage of the greenschist facies of the Purcell rocks of the anticlinorium and westward into the contact rocks of the Nelson batholith. It is postulated that the Belt-Purcell rocks of the anticlinorium grade downward into more intensely metamorphosed rocks and ultimately into a migmatite complex and granitic rocks. According to this hypothesis a level of _in situ_ melting was topographically well below a comparable level in the Shuswap Terrane. Although the high grade regionally metamorphosed Precambrian and Paleozoic rocks of the Kootenay Lake area have not been mapped southward into Idaho, rocks of comparable metamorphic grade occur on strike to the south on the west side of the Purcell trench. The rocks here are the oldest of the Belt-Purcell Supergroup, stratigraphically well below therocks of the Shuswap complex. These high grade metamorphic Belt rocks are in upfaulted blocks that were metamorphosed well below a comparable depth in the Shuswap Terrane. From the distribution of regionally metamorphosed rocks I infer a zone of partial melting [1/] that extended from the Cascade Mountains to the Belt-Purcell anticlinorium and during early stages either sloped eastward across essentially horizontally bedded rocks, or extended horizontally across rocks dipping gently seaward.

If one can accept this hypothesized zone of melting, it is easy to believe that from it the Mesozoic granites were generated. It is only one step further into the boundless field of speculation, to postulate that the thermal high of the Mesozoic persisted into, or more likely, was rejuvinated in, the Eocene and that the Eocene granites were generated by a similar mechanism. The late Tertiary plutons of the western Cascade Mountains do not fit this pattern and because of their beltlike alignment and association with andesite volcanic rocks may have a different origin.

[1/] This is not to be confused with the partial melting zone postulated to explain the seismically quiet (low velocity) zone in the upper mantle at a depth between 100 and 200 kilometres.

According to the above hypothesis, the quartz diorites of the Cascade Range are differentiates from a remelted segment of upper oceanic (basaltic) crust (see Misch, 1966, p. 106) and the more alkalic intrusions. such as the Golden Horn batholith (Misch, 1966, p. 139), from a particularly silica and potash-rich infold of eugeosynclinal sediments. Rocks east of the quartz diorite line, as for example the Colville "batholith" and within the Pacific Border Province would have been derived by anatexis of the Eugeosynclinal prism that had been deposited on oceanic crust. Those intrusions still further east, such as the Kaniksu batholith and Spirit pluton that intrude the miogeosynclinal assemblage and the Belt rocks, would be derived from the lowermost Belt rocks and the Hudsonian crystalline basement.

At the present state of knowledge, explanations of the rise in heat that softened the basement more properly belongs in the field of metaphysics. To relate the heat rise directly to a conventional plate subduction model is difficult, because the usual model controlled by the Benioff zone is constructed with a steep slope and requires a systematic arrangement by composition and age of the igneous rock products.

A Benioff type model of a subducting plate that generates andesitic and granitic magma (Dickinson, 1970) is commonly constructed with the plate junction a plane inclined at 45°, although the zones range from 30° to 70°. If a postulated subduction zone projects to the surface immediately west of the Cascade Range--and the most western granitic rocks--it would lie more than 400 kilometres deep in Idaho. It is highly unlikely that similar granitic magmas could be contemporaneously generated and emplaced through a vertical range of 400 kilometres and equally unlikely that magma could be generated from the same basaltic crustal layer of the descending slab and move upward through cooler mantle rocks--even diapirically--for 400 kilometres. The Benioff model could better explain magma generation if the dip of the subduction zone was reduced to 6° or less, but this negates any relation between deep focus earthquakes and an active plate junction.

In figure 39 I have reconstructed a plate model that appears to be capable of reproducing the geologic history of northeastern Washington and adjacent areas. By using average depths to the Conrad and Moho discontinuities on oceanic and continental plates and gross values for stratigraphic thicknesses and by assuming there was no tectonic junction between continental and oceanic crusts and that continental waste was deposited west of the continental margin, I arrive at a cross-section (fig. 39A) representing the continent as it was in the Early Triassic. Figure 39B shows the margin in the Late Jurassic after the oceanic crust was sheared and subducted beneath the continent. The two plates are moving towards each other, but as the oceanic plate advances its front descends, permitting the continental plate to over-ride and push back the descending part of the oceanic plate, causing the axis of descent, or "turn-down," to retreat towards a spreading center to the west. In this model oceanic crust is generated and consumed, but as the oceanic plate grows, its width, paradoxically shrinks.

To facilitate folding of the sedimentary prism above the continental plate, a décollement was introduced at and above the Conrad discontinuity. The décollement is "nailed" down at both ends so that as the two plates advance towards each other, the prism compensates for the shortening by wrinkling and thrusting, which results in a pileup of rocks at the front of the North American plate. In the early and intermediate stages of plate interaction (fig. 39B) the prism rocks are piled upon the continental

Figure 39.-- Geologic cross-sections speculating, in the context of plate tectonics, on the Mesozoic history between Puget Sound and the Purcell Trench.

A. --Before the Mesozoic orogeny. Illustrates the nondisruptive join between continental and oceanic crusts. Western limits of stratigraphic units are interpretations without direct support. Red arrows shown here and in cross-sections B and C indicate direction of heat flow.

B. --Disruption of the bond between continental and oceanic crust and subduction of the oceanic plate. Postulated heat cell is divided and flow reoriented, heat under edge of continent is increased. As oceanic plate descends, North American plate moves westward negating the advance of the oceanic plate. Sedimentary overload on continent side of subduction is detached by a fault from the underlying plate and it crumples as the underlying part of the North American plate moves westward.

C. --Final stage of the compressional history of the orogeny. Increased heat flow causes partial melting of the lower siliceous crust with production of anatectic granties and a zone of flow upon which the overlying crust "floats." There can be no complete separation of the events and processes shown in B and C: those beginning in B continue into C and those shown in C began in B. Sedimentation is continuous. For simplicity the diagrams omit unconformities and facies changes and neglect the significance of a trench and volcanic arcs.

plate, but as the pile increases in mass it overlaps the oceanic plate, the décollement becomes unnailed at its western end permitting it to advance westward carrying the continent derived prism of sediments over the ophiolite suite. Loading the continental margin by thrusting the sedimentary prism eastward while folding it to several times its depositional thickness pushes the continental plate downward. The result is a tightening of the join between the two plates and consequent increase in frictional heat. In this model none of the continent derived waste is carried beneath the continent on a subducting plate. The continents grows by extensional loading of continental waste on oceanic crust, a process that can be termed, supra-accretion. Other models of continent and ocean plate junctions tuck the sedimentary prisms down the maw of the junction, supposedly to subject the sedimentary and volcanic rocks to the high pressure requirements of blue schist metamorphism, the products of which by some mysterious process are regurgitated to the surface. The lower part of the prism, milled by the movement on the plate junction, eventually reaches a level where partial fusion rejects much of the silica and large cations, which move upward to rejoin the continental crust. Such is infra-accretion. This, however, only balances the loss by erosion; overall growth must come from the mantle.

To avoid confusion, figure 39B omits the generation of granitic magma, although by the Late Jurassic plutonism was well advanced and an important process in the Mesozoic orogeny. On the other hand, figure C diagrammatically represents the conversion of the sedimentary prism into crystalline continental crust [1/], although it was not drawn to represent any particular time in the Mesozoic. Because plutonism, whether continuous or episodic, was of such long duration, it is probable that the lower continental crust under the western part of the Cordillera was in various states of melting during Mesozoic. Such a hypothesis, supported by the close association in time and space between metamorphic terrane and granitic rocks, provides a mechanism that interrelates folding, granite emplacement, and the thrust faults of the Rocky Mountains.

[1/] The above described model generates granitic magma by anatexis of crustal rocks, which at the time of writing is in cyclic disfavor, mainly because of the faith placed in isotopic correlations. Admittedly proof of anatexis is not available: sampling of possible source rocks is lacking, nor is there precise control on the compositional variations of the plutonic rocks across the cross-section area. Nevertheless, to deny the contribution of continent derived sediments to magmas generated along and near continental margins, is to deny a fundamental law of physics and the thesis of global tectonics--that mass or energy can neither be created nor destroyed.

In figure 39B the folds and thrust faults terminate at the décollement, consequently shortening of the continental plate is restricted to the sedimentary prism. The shortening not only increases the thickness of the prism, but concentrates greater mass over the subduction zone. For some reason not fully understood, this shortening is accompanied by a heat rise. Several of the many factors that might contribute to this are: (1) adiabatic increase caused by concentration of load, (2) movement of the North American plate towards the spreading center, and (3) change in the pattern of convecting cells caused by the subduction of the oceanic plate. It is doubtful if the heat produced by the compression of the prism could be retained for any useful period of time, although this might be a contributing factor. The second factor accepts the model requirement that the continental plate move towards the spreading center, but is faced with the problem of subducting warm instead of a cold oceanic plate. A warm plate could not be forced down as readily as a denser cold plate. The third factor, change in the pattern of convecting cells appears more promising.

If a convecting cell existed beneath the join between oceanic and continental crust and then was beheaded by a descending oceanic plate, the size of the cell would be sharply reduced and the heat flow substantially increased (fig. 40). Until more is known about convection of heat in the upper mantle, it is difficult to evaluate such a heat distribution system. However, by postulating a heat source closer to the surface, variables are better controlled and the hypothesis becomes somewhat more realistic. Radioactive heat generated in the sedimentary prism would be doubled per unit cross-section area, if the prism was tectonically reorganized to double its thickness. The increase in thickness increases the insulating potential of the prism and conceivably might bring temperatures at the base of the prism

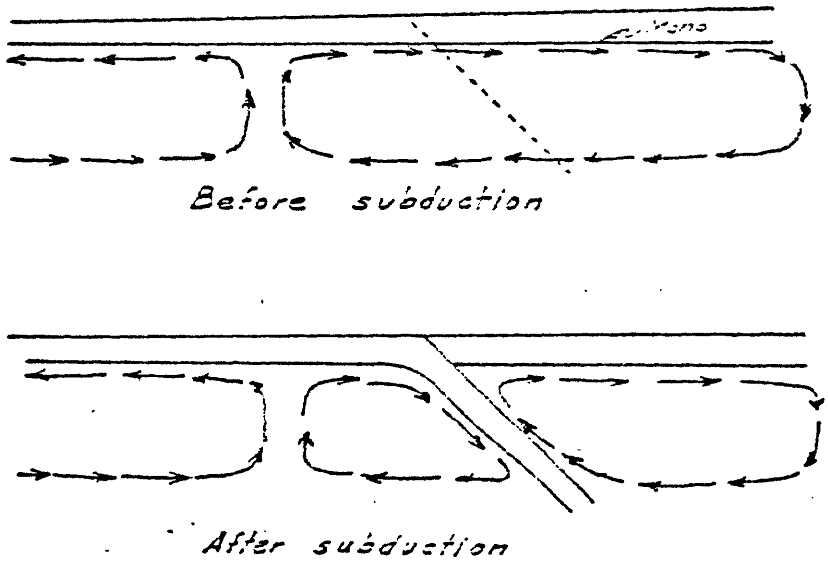


Figure 40.-- Sketch suggesting change in heat potential of a convecting cell that might result if the cell is bisected by a subducting plate.

up to melting. This conflicts, however with the logical contention of Resser (1970, p. 83) that metamorphism and deformation are contemporaneous.

The melting of the lower part of the prism also can be accomplished by the addition of water, which lowers the melting point of the assemblage. This requires and adequate original temperature and sufficient water. In the model shown in figure 39B the most feasible way to add water is to distill it from mantle serpentine in the descending plate. Figure 41 shows that serpentine loses its water by several mineralogic changes: the first loss at the lowest temperature and a geothermal gradient of 25° km and a depth of 20 km would be about 150° lower than the melting point of "wet granite" and the last stage, the breakdown of talc to enstatite, would be at 35 km and 900°, which is well above the melting point of granite. In any case, water would not be released through the entire distance from the surface to a depth of 40 kilometres.

Although the mechanism is uncertain, the presence of a zone of partial melting extending from the Cascade Range to Idaho, seems inescapable. It is assumed that early in the Mesozoic the zone of melting was either horizontal or dipped gently eastward. The décollement shown in figure 39B would fall within the zone of partial melting, because this would be the weakness along which separation would occur. As the North American and Pacific plates continue to move towards each other, folding and thrusting along the decollement compensated for the shortening, but when folding ended and the prism is under extension, thrust faults have to be activated by some other force than inter-plate reaction. As long as the zone of melting is essentially flat, the overlying rocks exert no effective pressure against the crust to the east, but once the zone is tilted or domed, the overlying mass moves downslope and over-rides the rocks beyond the borders of the partial melting zone.

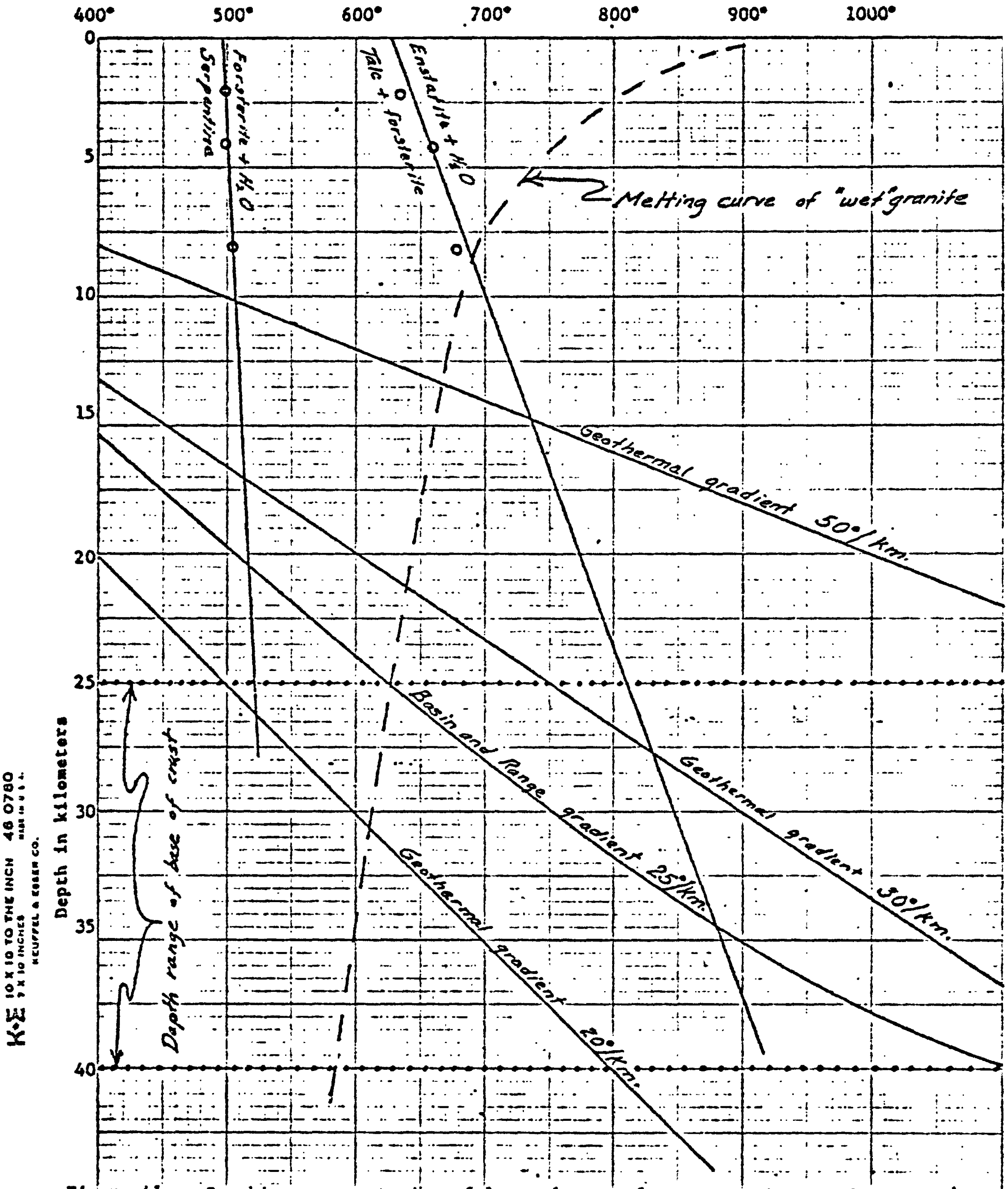


Figure 41.--Graphic representation of loss of water from serpentine resulting from subduction of an oceanic plate. Pressure-temperature data on mineralogic changes from Bowen and Tuttle (1949, p. 439).

The Cenozoic and extension tectonics

In northeastern Washington the Cenozoic regime began in the Cretaceous. If the geologic time scale had been based upon the geologic history of this area, the Mesozoic would have been over by the Late Cretaceous, because in this area, the Eocene is an augmentation of the Late Cretaceous, sharing a common history of continental sedimentation, plutonism, volcanism, and extensional faulting.

The plutonism of the Eocene was not too different from that of the Mesozoic. Aside from the Coryell batholith (north of the Northport quadrangle), which is more alkaline than most Eocene or Mesozoic plutons in the region, average compositions are very similar. Eocene plutonism and volcanism is particularly impressive because of its great regional extent, reaching from the Pacific to the plains of Montana. The generation of granitic rocks over this vast area indicates an area of high heat rise of similar extent. The magmatic hearth of this Eocene plutonic event is not directly evidenced, like that of the Mesozoic, by a terrane of gneiss and migmatite of correlative age, but we can assume that something similar lies below the present surface and is comparable with the exposed Mesozoic Shuswap Terrane that represents an earlier high heat rise. In the following pages I pursue this parallel in order to demonstrate a correlation between plutonism and the Rocky Mountain thrust belt.

The time of thrusting in the Canadian Rockies overlaps both the compressional and extensional events in the Kootenay arc. Active thrusting in the foothills of the Rocky Mountains of Alberta was contemporaneous, or at least penecontemporaneous, with the Eocene plutonic event and the extensional faulting in northeastern Washington. According to Bally, Gordy, and Stewart (1966, p. 370) horizontal shortening on the Foothill thrust faults is post-Eocene and accounts for but 25 percent of the total displacement in the whole of the Rocky Mountain thrust fault belt. These writers believe the remaining 120 kilometres of shortening west of the Front Range of Alberta probably occurred during the Mesozoic. Because the Mesozoic thrusting in the Front Range took place when the arc rocks to the west was being folded by being pushed eastward towards the craton, whereas the Eocene thrusting occurred when the area west of the Rocky Mountain trench was under tension, a change in the thrust activating force is indicated.

A reasonable hypothesis to explain the continuation of thrusting under an extensional tectonic environment instead of a compressional environment must supply a change in the inherited physical setting of the thrusts. The change proposed can be illustrated with an analogy of a boy pushing a wagon up one hill, which represents thrusts produced under compression, and the boy riding the wagon down the slope beyond, which represents the tensional environment of gravity thrusts. The main difference between the analogy and the hypothesis is that the analogy changes hills, whereas; the hypothesis changes the slope of the same hill.

The post-Paleocene thrusting in the Alberta foothills coincides with the climax of uplift and erosion of the terrane west of the Rocky Mountain trench that had been invaded by Eocene granitic rocks, whose presence suggest that the partial melting zone was extant or had been rejuvenated. The uplift of the area west of the Rocky Mountain trench--attested to by the Late Cretaceous conglomerates resting on Middle Jurassic rocks and Eocene volcanic rocks resting on Precambrian Belt rocks--tilted the partial melting zone eastward, thus converting the western part of the décollement from a compressional thrust to a gravity thrust as far east as the Rocky Mountain trench. East of the trench the décollement follows the top of the Hudsonian basement, which slopes westward. The uplift of the western part of the décollement raised the overlying rocks that were of sufficient mass to more than counterbalance the mass of rocks above the décollement east of the trench. To equalize the unbalanced masses resting on opposing and facing slopes, the larger, western mass pushed the eastern mass upslope as thrusts broke from the décollement in the foothills. This mechanism resembles the model of Hose and Danes (1973, p. 437, fig. 6) except that their décollement is lubricated by a zone of high fluid pressure instead of a zone of partial melting, which in effect are the same.

As the rocks above the décollement moved eastward, variations in frictional drag caused the plate to pull apart along faults of northerly trends that perhaps originated during the cross fold event. These are listric, downward flattening, faults and include the fault of the Purcell trench, the Flume Creek fault, the Leadpoint fault, north trending faults of the Chewelah and Newport quadrangles, the Kettle River fault, and faults of the Republic graben.

The development of graben and normal faults accompanied the regional uplift that began in the Late Cretaceous. These faults represent in part the tectonic gap, or compensating extention, that Price (1971, p. 1134) gives as an essential balance in any thin skin thrust complex. Although no measurements or estimates of horizontal displacments on these faults have been made, some idea of the potential can be obtained through their vertical displacement. In contrast with a vertical fault with a horizontal displacement of zero, a normal fault that has a vertical displacement of 6 kilometres, such as the Flume Creek fault, and flattens in depth to near horizontal, has a horizontal displacement approaching 6 kilometres figure 42. The sum of extension on the normal faults from the Cascade Range to the Rocky Mountains front certainly would not compensate for the 160 to 200 kilometres Price and Mountjoy (1970) and Bally, Gordy, and Stewart (1966) proposed for horizontal movement in the southern Canadian Rockies but it could account for the 40 kilometres Bally and others indicate for the post-Paleocene movement, which is the only period of time when the thrusts need extensional compensation.

The Late Cretaceous sedimentary rocks and the Eocene volcanic rocks from the Methow-Pasayten graben to the Purcell trench are almost everywhere on the downside of northerly trending extension faults. The association is not genetic; the late rocks are there because they are protected from erosion. Only in the Republic graben (Muessig, 1967, pl. 1) are igneous rocks intruded along faults. From the Republic graben eastward, the east side of the fault is the down side, however, the displacements are not cumulative, but are compensated by westward tilting of the faulted blocks (fig. 43).

The concept of listric normal faults as applied here can be extended to faults in the Newport 30-minute quadrangle and Panhandle of Idaho that have been interpreted as thrust faults (Miller, 1971 and Miller and Engels, 1975) because they have dips as low as 30°. Descriptions of these faults are in all respects compatible with the interpretation that they are the flatter, deeper parts of listric faults and represent extensional movement. The descriptions indicate that upper plate movement is everywhere downward and lower plates everywhere represent deeper crustal levels than do the upper plates. The broad cataclastic zones described in the lower plate rocks are not characteristic of thrust faults, but are to be expected in association with listric normal faults where downward change in dip causes new planes of movement to develop as faulting progresses. The faults

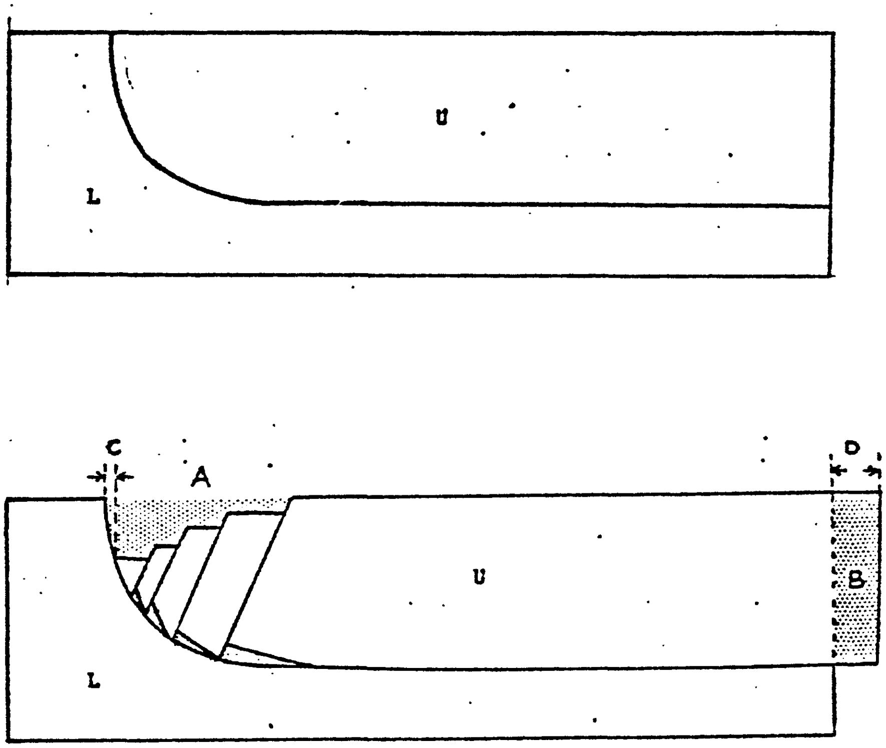


Figure 42.-- Idealized sketch showing relation between horizontal and vertical displacements on a listric fault. The example shown is the extreme case, where the fault changes from a vertical dip near the surface to a horizontal dip at depth. The shaded area at A is equal in cross-section-area to that at B. Horizontal displacement measured at C, is not a measurement of horizontal extension between blocks L and U, which is correctly indicated at D. The true extension is also the sum of displacements on all faults at A, but if the geology is such that displacements on left dipping faults cannot be measured or if the hypothetical void created by the pull-apart is filled -- as it develops -- by shear folding in the area A, the extension measured at C may be the only measurable displacement. This type of fault can have application where a shallow zone of partial melting exists and the listric faults merge with this zone, which is capable of horizontal flowage.

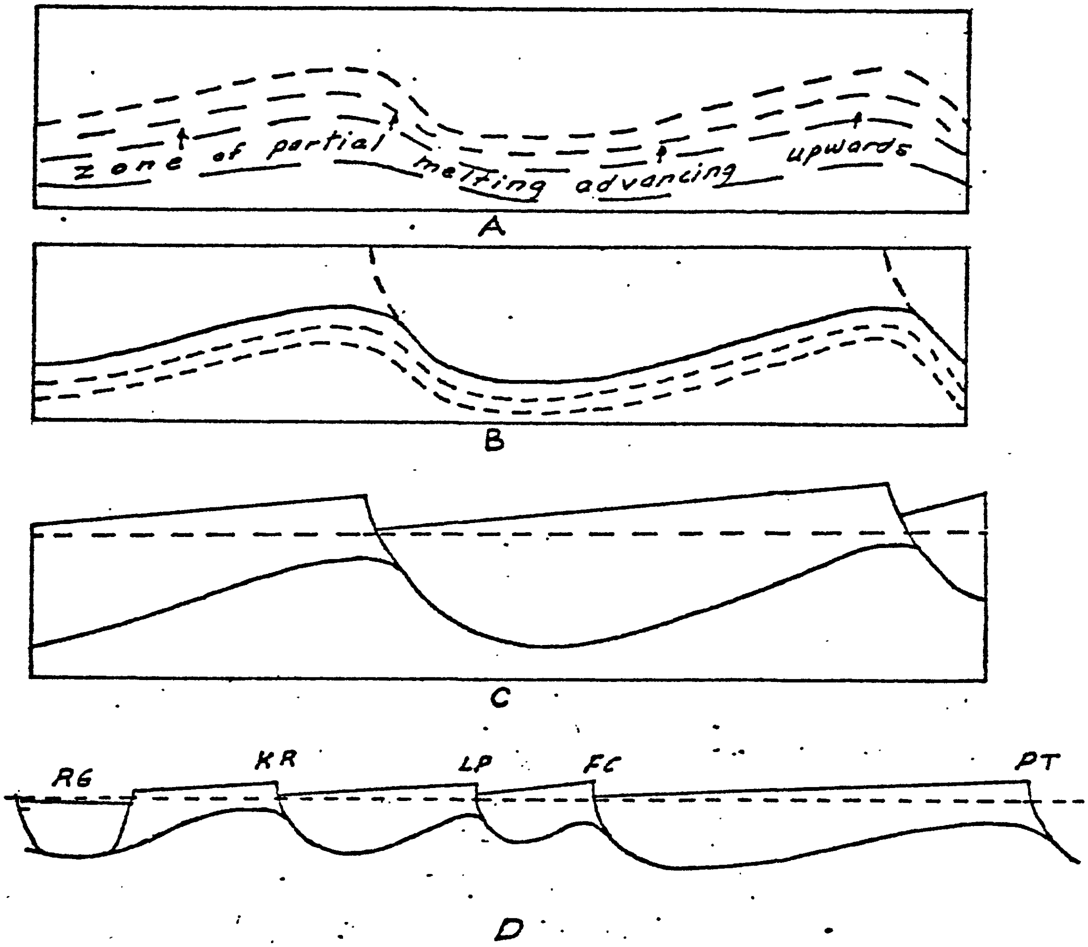


Figure 43.-- The concept of listric faults and tilted blocks.

A - High heat rise produces partial melting zone, which advances upward.

B - Partial melting reaches a level where crust is thinned to a point of failure along listric normal faults that merge ;e downward with the zone of flow (partial melting).

C - Failure is complete, segmented crust begins to sink into partial melting zone. Blocks rotate because of uneven distribution of mass. Thick portion descends, thin portion tilts upward, but aggregate movement is downward. Sinking ceases when partial melting zone freezes. Dashed line is a horizontal reference plane.

D - Application of hypothesis to northeastern Washington and northern Oregon. Interpretive cross-section during the Eocene drawn from the Republic Graben (RG) to Purcell Trench (PT). Doming with eastward tilting (not shown) probably accompanied melting. Horizontal control approximate, vertical scale exaggerated. KR - Kettle River fault, LP - Leadpoint fault, FC - Flume Creek fault. Dashed line is a horizontal reference plane.

shown in figure 43 can be thought of as gravity thrusts with both upper and lower plates in motion, but not as thrusts with autochthonous lower plates.

The normal faults may have developed as illustrated in figure 44. Presumably at the time of normal faulting, the area of continental crust under observation was a moderately rigid body floating on a semifluid medium. If the thickness of the crust above the zone of flow was irregular, stresses would develop in the thinner parts. To illustrate this, figure 43 shows a crust of irregular thickness and the "water line" the plate sinks to in the semifluid media of high viscosity. If the plate is rigid, the water line, determined by the total mass of the plate, averages out all thickness variations; however, if the plate separates at stress points KR, LP, and FC, each resulting block, having a different thickness will settle to its own "water line" which is a function of the mass of the block and the time of freezing of the semifluid zone. The thicker parts of the blocks go down faster than the thinner parts (fig. C). By mentally returning the blocks to their original positions in the plate, the points of strain become obvious.

If the concepts expressed in the above interpretation are valid, the form of the partial melting zone before the breakup can be reconstructed from the fault pattern. Because all faults east of the Republic Graben dip eastward and are down on the east, an asymmetric profile of thickening and thinning is indicated and is pictured diagrammatically in figure 43C and 43D as a series of giant ripples.

If the extensional faults are truly listric faults, they are in theory capable of compensating for the 40 kilometres of horizontal displacement on the Tertiary thrust faults in the foothills of Alberta. However, they can produce 40 kilometres of displacement only if there is a comparable amount of thrust displacement. The only other way to extend the crust above a décollement and not expose the décollement, is to lengthen a unit of crust by arching accompanied by extensional faulting. Arching of a 600 kilometre wide section of the earth requires a fantastic amount of uplift to yield 40 kilometres of extension, as can be seen by solving the following equation, $s = 1\ [1 + 2/3(\frac{2d}{1})^2]$ for the unknown d, the amount of uplift for the chord (1) 600 kilometres long to be arched into an arc (s) 680 kilometres long. The solution yields a value in excess of 130 kilometres, which is an inconceivable blister on the face of the earth.

I end this paper on a strong note of "ifs" and if the interpretations based on these ifs stimulate the indignation of my critics into the creation of a new set of its, I will be fully rewarded.

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