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Geology of the Chewelah-Loon Lake Area, Stevens and Spokane Counties, Washington

GEOLOGICAL SURVEY PROFESSIONAL PAPER 806

Prepared in cooperation with the Washington Division of Mines and Geology

Geology of the Chewelah-Loon Lake Area, Stevens and Spokane Counties, Washington

By FRED K. MILLER and LORIN D. CLARK

With a section on POTASSIUM-ARGON AGES
OF THE PLUTONIC ROCKS

By JOAN C. ENGELS

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CONTENTS

ILLUSTRATIONS

TABLES

GEOLOGY OF THE CHEWELAH–LOON LAKE AREA, STEVENS AND SPOKANE COUNTIES, WASHINGTON

By FRED K. MILLER and LORIN D. CLARK

ABSTRACT

The report area, two 15-minute quadrangles, is in the southern part of Stevens County and the northern part of Spokane County, about 30 miles north of Spokane. Much of the area is underlain by two great Precambrian sections—the Belt Supergroup and the Deer Trail Group. The Deer Trail Group appears to be equivalent to the upper part of the Belt Supergroup, although differences in thickness and stratigraphy suggest that the sites of deposition of the two sections were much farther apart than the sections are now. The Precambrian Huckleberry Formation and Monk Formation unconformably overlie the Deer Trail Group, but not the Belt Supergroup. Both Precambrian groups are overlain by the Cambrian Addy Quartzite. Cambrian, Devonian, and Mississippian carbonate rocks are found above the Addy Quartzite where it overlies the Belt Supergroup.

Nine plutons, representing three periods of plutonic activity, intrude the Precambrian and Paleozoic rocks. They range in composition from granodiorite to alkali-rich quartz monzonite. The oldest is the Flowery Trail Granodiorite, an isolated pluton near the center of the area. It appears to have been intruded about 200 million years ago. Five others, mainly in the northern part of the area, were intruded about 100 million years ago, and intrusion of the remaining three, the Silver Point Quartz Monzonite and two small satellitic plutons in the southern part of the area, apparently climaxed plutonic activity about 50 million years ago.

Andesite of Oligocene(?) age occurs in one small area southwest of Chewelah. Yakima-type(?) flows of the Columbia River Group are preserved at lower elevations in the southern part of the area. Small patches of conglomerate, possibly of Tertiary age, unconformably overlie the Huckleberry Formation southwest of Chewelah and the Starvation Flat Quartz Monzonite near Cliff Ridge. Quaternary glacial and alluvial deposits cover large parts of the west half of the area at lower elevations.

The Belt Supergroup and the Deer Trail Group appear to be confined to different structural blocks that are separated by a major structural discontinuity. The chief structures of the Belt Supergroup block, which underlies most of the eastern part of the area, are a roughly north-south-striking anticline and syncline. Both folds are overturned to the west in the northern part of the area. Six large northwest-striking high-angle faults appear to predate the folding and are probably Precambrian. Although the sense of movement on these faults is not well established, five show apparent left-lateral slip, and one shows apparent right-lateral slip. The five could well be dip-slip faults that have displaced rocks on the north sides downward relative to the south sides.

The north-south folds and northwest faults of the Belt Supergroup block are apparently truncated by the Deer Trail Group block, which underlies the western part of the area and strikes N. 30°–40° E. The consistency of the different trends of, and in, the two blocks and the apparent differences in facies and thickness of probably equivalent rocks in each block suggest that the blocks have been juxtaposed along a thrust fault.

High-angle faults, which are found in both structural blocks, strike about N. 50°–60° E. and have displaced rocks on the south sides downward relative to the north sides. These faults may be contemporaneous with a set of approximately north-south-striking faults which are inferred along the east side of the Colville River valley.

The major thrusting is interpreted as having taken place less than 100 million years ago. The inferred north-south faults on the east side of the Colville River valley appear to cut the 50-million-year-old Silver Point Quartz Monzonite but do not offset the Miocene and Pliocene Columbia River Group.

INTRODUCTION

LOCATION AND ACCESSIBILITY

The report area, which coincides with the east half of the Chewelah 30-minute quadrangle, covers about 400 square miles of Stevens and Spokane Counties in northeastern Washington (fig. 1). Although the 30-minute topographic map is now out of print, 7½-minute topographic coverage is available for the entire quadrangle. The north half of the report area comprises the Calispell Peak, Cliff Ridge, Chewelah, and Goddards Peak 7½-minute quadrangles, and the south half the Nelson Peak, Valley, Springdale, and Deer Lake quadrangles. In addition, the north half is topographically mapped at a scale of 1:62,500 and is named the Chewelah Mountain quadrangle.

Figure 2 shows the location of the report area in relation to Spokane, the nearest major city, and in addition shows many of the geographic features frequently referred to in the text. South of Chewelah, the west edge of the area follows the west edge of the Colville Valley, and the east border roughly follows the divide separating the Colville and Pend Oreille River valleys.

Most of the roads in the area are unsurfaced. Numerous county, logging, and mining dirt roads provide easy access to all parts of the area except the northeast corner. The Burlington and Northern Railroad links Chewelah, Valley, Springdale, and Loon Lake with

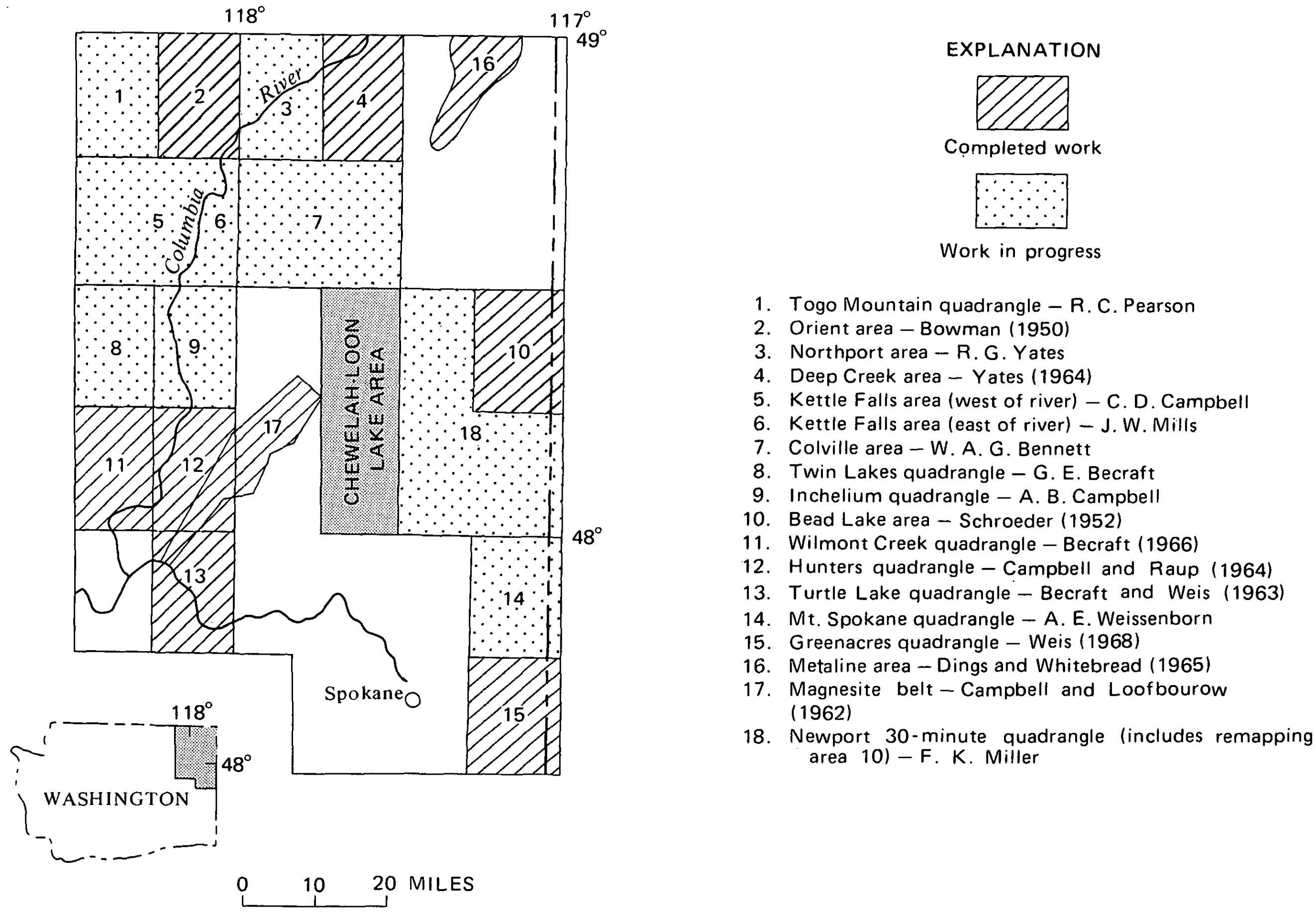


FIGURE 1.—Location of the Chewelah–Loon Lake area and surrounding areas for which geologic maps at a scale of 1:62,500 or larger are available or mapping is in progress. Date indicates published report available; see list of references in text.

Spokane to the south, and Nelson, British Columbia, to the north.

PREVIOUS WORK

Weaver (1920), in a comprehensive report on the mineral resources of Stevens County, published the first geologic map of the area and attempted to establish a stratigraphic section. Jones (1928) mapped the Chewelah 30-minute quadrangle in considerably more detail than Weaver. He retained most of Weaver's units, differentiated several more, and recognized the conglomerate at the base of the Huckleberry Formation. Studies since Jones' have concentrated mainly on the northeastern Washington magnesite belt, immediately west of the report area (fig. 2). The magnesite belt is underlain by many of the Precambrian and lower Paleozic units found in the report area.

Bennett (1941) prepared the first good geologic map of the magnesite belt. He improved on earlier structural and stratigraphic interpretations, and he assigned a Precambrian age to the thick section of fine-grained clastic and carbonate rock beneath the Huckleberry Formation. Besides mapping in the magnesite belt, Bennett prepared a map of the south half of the Colville 30-minute quadrangle, a generalized version which appeared in Mills and Yates' report on high-calcium limestone (1962, pl. 6).

Campbell and Loofbourow (1962) prepared a detailed geologic map of the entire magnesite belt and assigned all rocks beneath the Lower Cambrian Addy Quartzite to the Precambrian. They also compared the rocks of the Deer Trail Group with those of the Belt Supergroup (p. F20).

Campbell and Raup (1964) mapped the Hunters quadrangle, which covers most of the southwestern part of the magnesite belt. Their map, although at a slightly smaller scale than Campbell and Loofbourow's, is more detailed.

Schroeder (1952) mapped an area about the size of a 15-minute quadrangle around Bead Lake, east of the report area (fig. 2). He correlated parts of a thick section of quartzites, argillaceous sandstones, and argil-

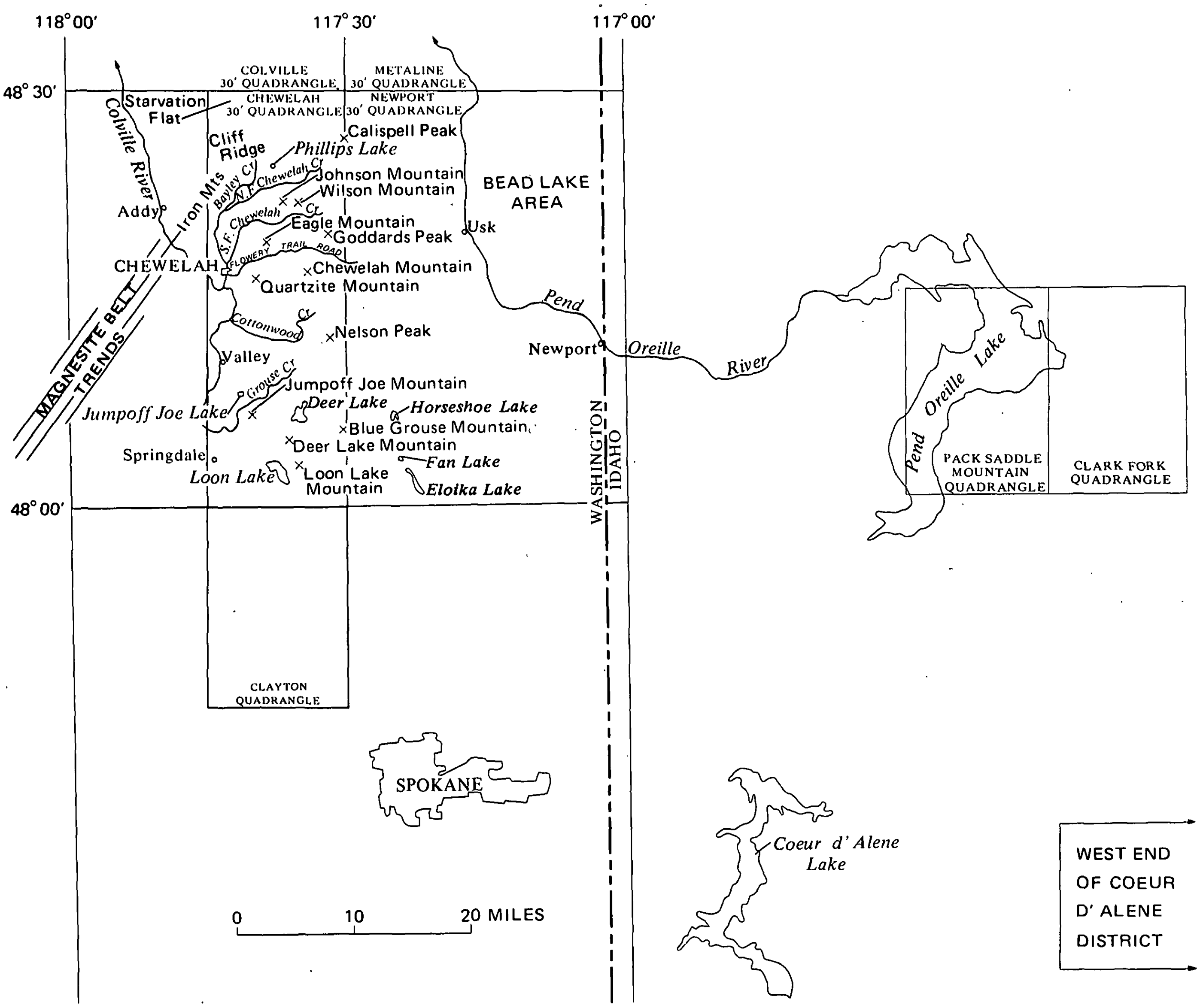


FIGURE 2.—Location of some geographic features referred to in the text.

lites there with the Prichard Formation and the Ravalli Group of the Belt Supergroup. This section of Belt rocks is the nearest one to the report area that has been relatively well studied. Pre-Tertiary sedimentary rocks known to exist between the report area and the Bead Lake area probably consist entirely or in part of Belt rocks also (A. B. Griggs, oral commun., 1967).

PRESENT WORK

Work in the report area began in June 1962 as a cooperative project of the Division of Mines and Geology of the Washington Department of Natural Resources and the U.S. Geological Survey. The authors mapped most of the Chewelah Mountain quadrangle in 1963 and 1964. Miller finished the mapping in 1965 with the help of J. C. Moore. Mapping of the southern 15-minute quadrangle was begun by Miller in 1966, continued with the help of R. C. Reynolds in 1967, and completed in 1968. Two preliminary geologic maps have been published by the Washington Division of Mines and Geology (Clark and Miller, 1968; Miller, 1969).

ACKNOWLEDGMENTS

The authors would like to thank Mr. Phillip P. Skok for his cooperation and help during the study of the mines in the Eagle Mountain area. Mr. Skok, representing several of the mine owners, very generously allowed many of the company mine maps to be duplicated in the Chewelah Mountain quadrangle preliminary report.

We are particularly indebted to Allan Griggs, of the U.S. Geological Survey, who introduced us to the Belt Supergroup in the Coeur d'Alene district and then helped with many of the stratigraphic problems we encountered in the Belt Supergroup in the Chewelah–Loon Lake area. We are grateful for the professional courtesies extended by Marshall Huntting, W. A. Bennett, and Jerry Thorson of the Washington Division of Mines and Geology, who helped in all aspects of the project from the fieldwork to the final editing. Bennett's intimate knowledge of the region saved both authors' time and effort on more than one occasion and several times directed us to pertinent features we might otherwise not have found. Finally, we would like to express our appreciation to Robert G. Yates and the late Arthur B. Campbell, of the U.S. Geological Survey, who helped us to become acquainted with the regional geology of northeastern Washington.

REGIONAL GEOLOGIC SETTING

In an excellent summary of the regional geology of northeastern Washington, northern Idaho, and northwestern Montana, Yates, Becraft, Campbell, and Pearson (1966, p. 47) recognized an eastern and a western geologic province, divided by the Columbia River valley. They defined the eastern province as characterized by Precambrian Belt and Windermere rocks and miogeosynclinal lower and middle Paleozoic rocks, and the western province as composed of eugeosynclinal upper Paleozoic and Mesozoic rocks. The report area lies about 15–20 miles east of the boundary between the two provinces.

Both provinces are underlain by numerous bodies of Mesozoic plutonic rock, which Yates, Becraft, Campbell, and Pearson (1966, p. 56) divided into two major batholiths. As they have defined the boundaries of these batholiths, the report area includes parts of both batholiths. Most of the extensive basalts of the Columbia River Group lie south of the report area, although a few prongs extend into the southern part.

Broad open folds and high-angle normal faults which trend north-northwest are found in the Montana part of the eastern province. The complexity of the structure increases towards the west, and in Idaho west-northwest-trending strike-slip faults and locally overturned folds appear (Yates and others, 1966, p. 52).

In northeastern Washington, the folding and faulting are noticeably more complex and irregular. In the report area, strike-slip faults, thrust faults, and locally, large overturned folds are cut by northeast- and north-striking high-angle dip-slip faults. The major movements probably occurred in the Precambrian, middle and late Mesozoic, and early Tertiary.

PRECAMBRIAN ROCKS

BELT SUPERGROUP

The Belt Supergroup is a well-known section of Precambrian sedimentary rocks extensively exposed in western Montana, northern Idaho, and eastern Washington. In Canada it is known as the Purcell Series. It is generally divisable into a carbonate-rich eastern facies and a more clastic western facies. Many formations, some of which are correlative, have been named and delineated by various workers. The large number of Belt units, confusing to many who have not studied these rocks, is an indication of the complexity of the internal stratigraphy of the group. Detailed stratigraphic descriptions and measured sections of the Belt rocks are difficult or impossible to obtain in the report area because of poor exposures and extensive forest cover.

The report area contains the westernmost exposures of the Belt Supergroup. If the Deer Trail Group is correlative with part of the Belt, however, the Belt terrane extends about 15 miles further west.

PRICHARD FORMATION

DISTRIBUTION AND TOPOGRAPHIC EXPRESSION

The Prichard Formation is a thick sequence of interbedded argillite, siltite, and fine-grained quartzite. It underlies 38 square miles in the report area. An additional 12 square miles is covered by surficial deposits or occupied by younger plutons. About 13,000 feet of the formation is exposed in the core of a large, highly faulted anticline, which occupies the entire east-central part of the report area. Only the west half of the anticline is found here, but reconnaissance of the rocks to the east suggests that Belt strata continue in that direction and that much of the other half may have been replaced by plutons. The southeast limb of the fold may be exposed along the west shore of Horseshoe Lake about 7 miles east of Deer Lake. There, a thick section of dark laminated argillite, strongly resembling the upper part of the Prichard Formation, strikes northeast and is truncated by plutonic rocks just north of the lake.

The Prichard Formation was originally named the Prichard Slate by Ransome (1905, p. 280–281) for the exposures along Prichard Creek in the Coeur d'Alene district, Idaho, where the unit, consisting chiefly of argillite and well-indurated sandstone, was estimated to be more than 8,000 feet thick. The base has not been found in the report area or anywhere else.

The formation forms much of the higher mountainous country along the east border of the area. Although they do not crop out as well as the argillite, the quartzitic parts of the formation are probably responsible for the rugged topography formed by the unit. The road along the high ridge crest between Chewelah Mountain and Nelson Peak traverses quartzite for much of its length. Joint-bounded blocks of all sizes litter the areas underlain primarily by quartzite, but almost all have been heaved, rotated, or otherwise moved by surface or near-surface processes. In-place outcrops are abundant in areas underlain by predominantly argillitic parts of the formation, but topographic expression is more subdued.

STRATIGRAPHY AND THICKNESS

The Prichard Formation can be divided into at least three members in the report area. Although they have not been distinguished on the geological map because of lack of time, for descriptive purposes they are informally referred to as the upper, middle, and lower members in the following paragraphs. The upper member (fig. 3) is predominantly dark-gray to black laminated argillite, with a lesser amount of siltite, and contains only an occasional bed of quartzite. The middle member consists of interbedded siltite, quartzite, and minor argillite. The lower member probably consists of about 6,000 feet of interbedded siltite and argillite, although information is available only from sparse outcrops and float.

The upper member is locally divisible into three subunits, in descending order: 600 feet of argillite, 800 feet of siltite, and 3,000 feet of argillite. All the argillite in the upper member is well bedded to laminated and light to dark gray; locally it contains thin bleached zones, whose shape and orientation seem to be controlled by bedding. Very little quartzite or siltite is interbedded within any of the argillite of the upper member. The siltite subunit is fairly distinctive owing to its color, gray to dull yellowish gray, and its thick indistinct bedding, vivid contrasts with the relatively dark laminated argillite zones above and below. The subunit also includes a number of quartzite and argillite beds. The lower argillite subunit is generally darker than the upper one. It is the thickest relatively pure argillite zone in the report area.

The transition into the quartzitic middle member of the Prichard Formation begins about 4,400 feet from the top of the formation, where quartzite and siltite beds start becoming thicker and more numerous (fig. 3). The base is also a transition zone. The average thickness of the member is about 3,200 feet thick. Although it probably consists of about 50 percent quartzite, numerous zones within the member are similar to the transition zone at the top. Relatively pure argillite zones as much as several hundred feet thick also are found in this member, but most are distinctly

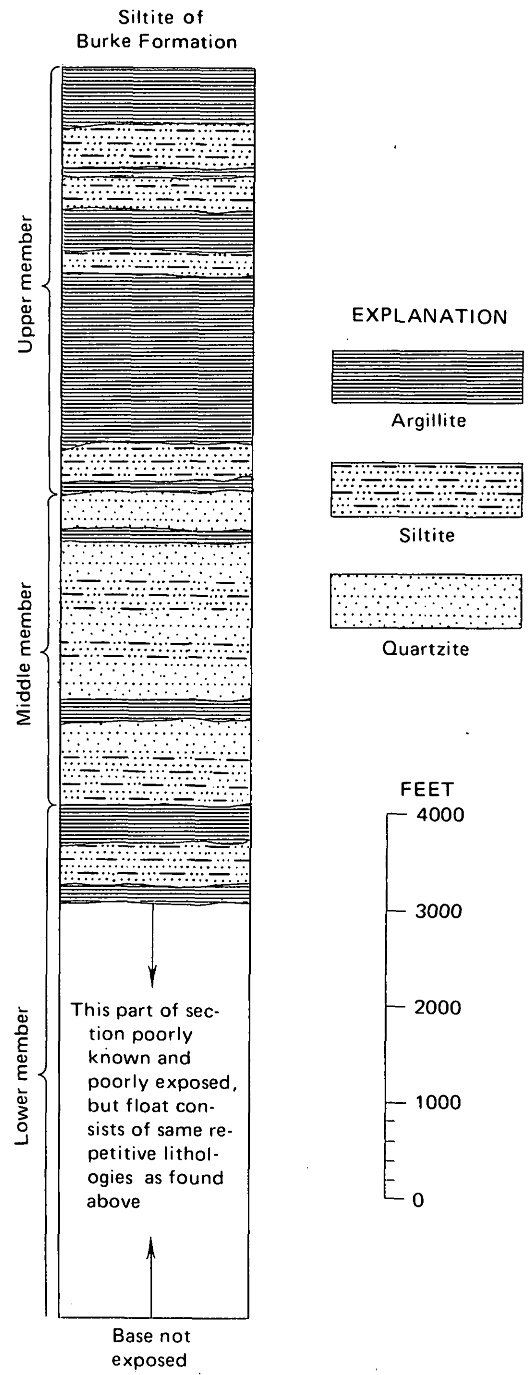


FIGURE 3.—Generalized columnar section of the Prichard Formation in the Chewelah–Loon Lake area. Lower, middle, and upper members are informal subdivisions and are not shown on the geological maps.

lighter colored than the argillite of the upper member.

The quartzite is light gray to white, fine grained, and vitreous. All gradations from quartzite to white siltite are found. Beds of both quartzite and siltite average between 1 and 5 feet in thickness and weather out in large joint-bounded blocks. In general the more siltitic to argillitic the rock, the thinner the beds.

The lower member only occurs east of the divide that connects Chewelah Mountain and Nelson Peak. Scattered outcrops, a few exposures in logging roadcuts, and sparse float suggest that it is composed of the same recurring lithologies as the other members but that siltite and argillite are predominant.

The top of the Prichard Formation in the report area is defined as the top of the highest 600-foot-thick argillite zone beneath the quartzite, siltite, and argillite transition beds of the Burke Formation. This is contrary to the placement of the Prichard-Burke contact in most other Belt sections and should be taken into consideration when the thicknesses of the Prichard and Burke Formations are compared with those in other sections, especially the relatively near well-studied sections around Pend Oreille Lake and in the Coeur d'Alene district. The upper part of the Prichard Formation in the report area appears to be more akin to that at Pend Oreille than that around Coeur d'Alene.

Most workers agree that the bulk of the Belt Supergroup was deposited in shallow water (Ransome and Calkins, 1908, p. 17; Yates and others, 1966, p. 47; Hobbs and others, 1965, p. 33). Hobbs, Griggs, Wallace, and Campbell (1965, p. 33) interpreted the relatively clean, well-sorted quartzites at various levels in the section to be suggestive of a nearshore environment. Others have pointed out that most of the sedimentary structures in the Prichard Formation are confined to the uppermost part (Campbell, 1960, p. 552; Campbell and Good, 1963, p. A8; Hobbs and others, 1965, p. 34).

In the report area, both ripple marks and mud cracks, although not abundant, have been observed in the upper part of the highest member. Cross laminations and graded bedding on a fine scale are found throughout the formation but are quite subtle. The pyrite and pyrrhotite and accompanying stained surfaces described by almost all investigators are also found in the report area.

Several sills of greenish-black amphibolite are found in the middle and lower members of the formation. They have been thoroughly recrystallized, probably from diabase, and the primary texture has been completely destroyed. Although they appear to have caused only slight recrystallization of the host rocks, they seem to have "set up" the rocks for metamorphism by later intrusive bodies. In the vicinity of younger plutonic rocks, the siltite and argillite adjacent to the sills show pronounced recrystallization two to three times as far from the plutonic contact as other rocks of comparable composition not located near the sills. Most of the sills appear to be strikingly concordant, with the possible exception of the large body northeast of Chewelah Mountain. This sill may be discordant, as it appears to be higher in the section in the northern part of the area. It is difficult to be certain, however, because of the strong overturning in this area and because of the possibility of additional undetected structures.

RAVALLI GROUP

BURKE FORMATION

DISTRIBUTION AND THICKNESS

The Burke Formation consists primarily of light-gray siltite with lesser amounts of argillite and quartzite. It crops out in patches along a belt extending from east of Deer Lake northward to the South Fork of Chewelah Creek. The belt is offset by many northwest-trending faults.

The Burke Formation was named and first described by Ransome (1905, p. 280–281) and Ransome and Calkins (1908, p. 32) for the exposures around the town of Burke in the Coeur d'Alene district. They reported a section about 2,000 feet thick "composed of rocks that show all gradations from nearly pure quartzite to siliceous shale." Hobbs, Griggs, Wallace, and Campbell (1965, p. 34) reported a similar thickness but also found that it varied from place to place in the district, generally thickening to the west.

The Burke Formation averages about 3,700 feet in thickness in the report area. Miller (1969, p. 1) erroneously reported an average thickness of 4,500 feet, which was calculated for the thickest exposed section of the formation, located along the margin of the major valley northeast of Deer Lake. Other thicknesses have been calculated; the minimum, 3,100 feet, was calculated for a section north of Goddards Peak. Although some thickness variation may actually exist, most of it is due to differences in where the contacts are placed, especially the upper one.

STRATIGRAPHY

The Burke consists mostly of light- to medium-gray siltite. The basal zone, 150 feet of alternating argillite and siltite, forms a transition between the argillite of the underlying Prichard and the siltite of the Burke. About 500 feet above the contact, the siltite grades into a quartzite zone approximately 300 feet thick. This quartzite zone, in turn, grades upward into a predominantly siltite section. Zones of quartzite or argillite as much as a few tens of feet thick are interlayered with the siltite, mostly in the lower half of the formation. The upper 400 feet is the only other part of the unit that contains a significant proportion of quartzite.

About 3,000 feet above the base is a distinctive 100–300-foot-thick zone of maroon to lavender argillite, siltite, and quartzite, an almost perfect duplicate of typical St. Regis lithology. The argillitic parts of the zone contain mud cracks, mudchip breccias, and ripple marks. The quartzite is very fine grained and in places coarsely laminated. Laminations average about an eighth of an inch thick and exhibit fine planar crossbedding internally.

Much of the Burke Formation, particularly the siltite, is medium to light gray with a subtle, barely noticeable purple cast. Some of the argillite beds in the uppermost and lowermost parts of the Burke Formation have a gray-green cast, as does some of the siltite. This coloration is restricted to specific parts of the Burke here but appears to be typical of much of the formation at Coeur d'Alene (Hobbs and others, 1965, p. 35) and Pend Oreille (Harrison and Jobin, 1963, p. K10). The light-weathering rind on Burke siltite described at both of the just-mentioned localities is present in the report area but is not nearly as well developed nor as extensive.

Beds range in thickness from less than 1 inch to more than 10 feet but commonly are about 6–12 inches thick. In general, the finer grained rocks form the thin beds, and the relatively coarser grained rocks the thick beds. Argillite, particularly in the uppermost and lowermost parts of the formation, appears to constitute a larger proportion of the section than it actually does. This is because thin argillite beds and partings form skins on large exposures of bedding surfaces.

Oscillation and current ripple marks are locally abundant in the upper part of the formation. Some quartzite strata at various horizons are crossbedded, and much of the siltite is finely cross laminated. In some layers the fine but somewhat indistinct laminations in the siltite are broken and disrupted, presumably by slumping or some other form of contemporaneous deformation.

The mineralogy of the Burke Formation, like that of the other Belt units, is relatively simple. Quartz, sericite, plagioclase, potassium feldspar are the most common minerals, generally in that order of abundance. Zircon, the most common heavy mineral, is found throughout the formation. Small magnetite crystals are also common.

REVETT FORMATION

DISTRIBUTION AND THICKNESS

The Revett Formation is predominantly white to light-gray fine-grained quartzite. Within the report area, the distribution of the Revett is similar to that of the Burke Formation. It parallels the latter and is disrupted about equally by the same cross-faults. East of Deer Lake, the formation continues to the southeast, but a few miles outside the area it bends to the east and terminates against a crosscutting intrusive body.

The Revett Formation was initially named the Revett Quartzite by Ransome (1905, p. 280–281) and Ransom and Calkins (1908, p. 35) in the Coeur d'Alene district, Idaho. Harrison and Campbell (1963, p. 1418), in a discussion of correlation problems within the Belt Supergroup, noted that the Revett is no more homogeneous than any other Belt formation in northern Idaho or western Montana and suggested that the name be changed to Revett Formation. Their observation regarding the inhomogeneity of the quartzite is quite apparent in the report area, and the new name they suggest will be used in this report.

Poor exposures prevent detailed measurement of any Revett sections in the report area, although from outcrop width the thickness of the formation appears to vary little from an average of about 3,100 feet. As Hobbs, Griggs, Wallace, and Campbell (1965, p. 37) pointed out, however, at least part of the discrepancies in thickness of the Revett Formation from place to place may be due to differences in the way that its contacts have been placed by different workers.

STRATIGRAPHY AND LITHOLOGY

Most of the Revett Formation in the report area is composed of fine-grained conspicuously clean looking vitreous quartzite. Although, as Harrison and Campbell pointed out, the Revett Formation is probably as heterogeneous as any other formation in the Belt Supergroup, the clean looking quartzite beds that characterize the Revett contrast rather strikingly with the predominantly impure clastics of the other formations.

In the report area, the depositional boundaries between all formations are transition zones, rather than sharp contacts. The transition zone between the Burke Formation and the overlying Revett Formation is the widest in the Belt Supergroup in the report area. The same criteria used by Hobbs and others (1965, p. 35) to draw the contact in the Coeur d'Alene district are used in this area; the base of the Revett Formation is the horizon above which beds of vitreous quartzite predominate over siltite beds. In the south half of the area, the transition zone has been reported to be about 200 feet thick (Miller, 1969, p. 1). In the northern half, where the contact zone is better exposed, as on Jay Gould Ridge east of Chewelah, this zone is closer to 400 feet thick.

The lower 200–300 feet of the Revett Formation contains a fair proportion of siltite beds. These beds decrease upward in number and, less consistently, in thickness. Above the siltitic base is about 1,000 feet of relatively pure vitreous quartzite. Even in this interval, however, there are both single and multiple beds of siltite.

Near the middle of the formation is about 500 feet of interbedded siltite, siltitic quartzite, and quartzite. The rocks of this zone are almost identical with the bulk of the Burke Formation and, lacking continuity with the more distinctive rocks above and below, could easily be erroneously assigned to the Burke Formation. Just below this zone is 50–100 feet of dark, banded quartzite, some of which shows signs of crossbedding. This quartzite is fine grained but coarser than most in the formation.

The middle siltitic part of the Revett Formation is overlain by a thick quartzite zone that contains relatively few siltite beds. About 200 feet from the top of the formation siltite beds are again noticeable and become progressively more abundant toward the contact with the overlying St. Regis Formation. In this upper 200-foot interval, the quartzite and siltite have a slight lavender tint, making the gradation into the overlying St. Regis not only one of grain size but of color also.

Beds range in thickness from less than 1 inch to more than 20 feet; average thickness is about 1–2 feet. Extensive talus composed of quartzite blocks commonly accumulates at the foot of slopes underlain by the Revett Formation.

Most of the quartzite is light gray to white. Siltite beds commonly are the same color, but many are darker gray, similar to the color of those in the Burke Formation. Thin bands of opaque minerals impart a faint banding to the quartzite but are not abundant. Disseminated dark minerals are common in most of the quartzite but do not materially affect the color of the rock. In places, quartzite beds contain pale-yellow roughly spherical pea-sized areas in which part of the cementing agent of the rock has been removed. Similar features, mentioned by Hobbs, Griggs, Wallace, and Campbell (1965, p. 38), appear to be more characteristic of the formation in the Coeur d'Alene district than in the report area. The weathered quartzite surfaces have a pock-marked appearance where these small vuglike features have been leached out by surface waters.

The siltite of the Revett Formation has virtually the same mineralogy as the Burke Formation. The quartzite is relatively pure but contains an unusually large amount of heavy minerals, as much as a half of a percent zircon and magnetite. Sericite and plagioclase are the most common impurities, in that order of abundance. Grains of potassium feldspar are common but not abundant.

Crossbedding and ripple marks are the only common sedimentary structures in the Revett Formation within the report area. Neither structure appears to be as well developed or as abundant as in the Coeur d'Alene district. A few beds containing mud-chip breccias are found in argillaceous beds on the mountain north of Deer Lake, but these appear to be local.

ST. REGIS FORMATION

DISTRIBUTION AND THICKNESS

The St. Regis Formation consists of maroon and green interbedded argillite and siltite. Numerous quartzite beds are interlayered with the finer grained rocks near the base. Good outcrops of the St. Regis Formation occur in only a few places in the large area underlain by the formation. Its distribution is roughly the same as that of the Revett and Burke Formations, except in the area around Wilson Mountain and the area south of the Flowery Trail Granodiorite. Around Wilson Mountain the formation is folded into a strongly overturned syncline. The dips of both normal and overturned limbs are fairly low. The St. Regis Formation is relatively thin for a Belt unit, but its outcrop width is about 2 miles in that area because of the low dip of the beds and the relatively subdued topography. (See cross section A–A′, pl. 1.) From the south margin of the Flowery Trail Granodiorite to Grouse Creek, most of the St. Regis Formation is covered by glacial debris.

In the report area, the average thickness of the St. Regis Formation is about 1,600 feet. No sections were measured, and at only a few localities are both contacts sufficiently well located so that a reasonably accurate thickness can be calculated.

Ransome (1905, p. 280–282) named and first described the St. Regis Formation in the Coeur d'Alene district. Because of the excellent cliff exposures there, Calkins (in Ransome and Calkins, 1908, p. 37) was able to accurately measure the thickness of the formation. In the east-central part of the district he measured 996 feet of section but mentioned that the formation appears to be thicker to the east and to the south. Hobbs, Griggs, Wallace, and Campbell (1965, p. 39) measured 1,400 feet for the same section but included a green quartzose argillite zone in the St. Regis Formation which Calkins put in the overlying Wallace Formation. In the report area, the green beds are also included in the St. Regis Formation.

According to Harrison and Jobin (1963, p. K12), the St. Regis Formation is from 600 to 1,100 feet thick in the Clark Fork quadrangle, 50 miles east of the report area. However, they assigned the distinctive green argillite beds, which are about 200 feet thick there, to the lower Wallace Formation; thus the range in thickness in the Clark Fork quadrangle can be increased to 800–1,300 feet.

STRATIGRAPHY AND LITHOLOGY

The St. Regis Formation is one of the more distinctive units in the Belt Supergroup because of its striking color. It is typically red-purple or lavender, except for several hundred feet of siliceous carbonate-bearing argillite at the top of the section, which is a distinctive yellow green.

The lower 50–200 feet of the St. Regis Formation is transitional into the dominantly quartzitic Revett Formation below. The lower contact has been placed where the red-purple argillaceous and siltitic rocks predominate over the relatively clean white and lavender quartzite of the upper part of the Revett Formation. Defining the contact in this manner makes its placement relatively simple in the field because most of the quartzite beds in the lower part of the St. Regis Formation are much more highly colored. White quartzite and red-purple argillite and siltite beds are actually interbedded in only about 50–100 feet of the transition zone.

The lower three-fourths of the formation consists of interbedded argillite, siltite, and quartzite. In general, the finer grained rocks are more highly pigmented. The quartzites are pale lavender, and the argillites deep reddish purple; the siltite ranges from one color to the other. In this part of the formation, the rocks grade from predominantly quartzite and siltite near the base, to predominantly argillite and siltite in the upper part. The unit, as a whole, appears to be more quartzitic in the report area than in the Pend Oreille or Coeur d'Alene areas.

The upper quarter of the formation is composed of alternating zones of red-purple clastic rocks and yellow-green carbonate-bearing siliceous argillite, which are 10–50 feet thick. The red-purple zones are siltite and argillite and are restricted to the lower two-thirds of this part of the formation. The yellow-green argillite in turn grades upward into the Wallace Formation, mainly through an increase in carbonate-bearing quartzite and black laminated argillite beds.

Individual beds range in thickness from 1 inch to about 4 feet in the lower part of the formation. The thicker quartzite beds are faintly banded by darker purple layers. The maroon argillite and siltite in the upper part of the formation are very thinly bedded and in places laminated. Almost all siltite and quartzite beds are separated by at least a film of red-purple argillite. Bedding characteristics of the middle part of the formation are not well known owing to poor exposure.

Mud cracks and mud-chip breccias are abundant throughout the St. Regis Formation, especially near the top. Even where the unit is mapped largely on the basis of float, many chips and pieces of float have mud cracks on them. Ripple marks and cross-laminations are also common but not nearly as abundant as in the Coeur d'Alene district, perhaps because exposures in one area are so much better than in the other.

In the open, south-plunging syncline on Blue Grouse Mountain, the maroon color of the St. Regis is completely bleached out in places. Along strike, the rocks grade from maroon, through mottled white and maroon, to white or pale green. The white and pale-green rocks are poorly indurated and altered looking. All lithologies are affected. The bleaching appears to be confined to a restricted area of hydrothermal alteration, which affects rocks underlying about half a square mile. The same type of bleaching is present around some of the large quartz veins on Blue Grouse Mountain, Deer Lake Mountain, and Bald Mountain.

WALLACE FORMATION

DISTRIBUTION AND THICKNESS

The Wallace Formation is a thick unit made up of argillite, quartzite, siltite, and carbonate rock. Much of the quartzite and siltite are carbonate bearing. The distribution of the Wallace in the report area is considerably more irregular than that of most other Belt formations. It forms a roughly linear but highly faulted belt around the margin of the anticline in the east half of the area, but also underlies part of Deer Lake Mountain and Loon Lake Mountain in the southern part. In the north half of the area, the formation forms part of an up-faulted block on the west flank of Eagle Mountain; the block extends northward to McDonald Mountain.

The formation is divided into an upper and a lower part in the report area. Hobbs, Griggs, Wallace, and Campbell (1965, p. 41) divided it in the same manner in the Coeur d'Alene district. Because the lower part of the formation forms fewer outcrops than any other Belt unit in the report area, it is difficult to obtain an accurate estimate of its thickness. The average calculated thickness of the lower part of the formation at the homoclinal section of the west flank of Bald Mountain is about 3,500 feet. This is probably as accurate a figure as can be obtained in this area.

A complete section of the upper part of the Wallace Formation is not exposed in the report area, as a fault marks the upper contact everywhere except on Deer Lake Mountain, and the base of the upper part is not exposed there. The thickest preserved sections are on McDonald Mountain, Eagle Mountain, and just east of Quartzite Mountain. The section on McDonald Mountain is not as thick as the map seems to show, because about one-third to one-half of the outcrop width is made up of sills and dikes of leucocratic quartz monzonite. A calculated thickness of 2,500–3,000 feet for the section east of Quartzite Mountain probably represents a maximum for this part of the Wallace Formation. That section also is fault bounded, although it appears to be relatively undisturbed internally. On Eagle Mountain, the upper part of the Wallace Formation is highly sheared and may contain unrecognized faults.

The formation was first named in the Coeur d'Alene district by Ransome (1905, p. 280, 282). Calkins (in Ransome and Calkins, 1908, p. 40) estimated the thickness there to be at least 4,000 feet, but he did not divide the formation. Hobbs, Griggs, Wallace, and Campbell (1965, p. 42), who studied that area in greater detail, found no well-exposed uninterrupted sections of the formation. From measurements of several partial sections they concluded that the total thickness is between 4,500 and 6,500 feet.

STRATIGRAPHY AND LITHOLOGY

LOWER PART

The lower part of the Wallace Formation is composed of argillite, siltite, quartzite, highly impure carbonate rock, carbonate-bearing quartzite, and carbonate-bearing siltite. It is varied in lithology, but is generally easy to recognize not only because of the presence of carbonate-bearing rocks but because of bedding characteristics which are unique in the Belt Supergroup.

The lower contact of the Wallace Formation is assigned on the basis of criteria that Hobbs, Griggs, Wallace, and Campbell (1965, p. 43) used because this part of the section is similar to that in the Coeur d'Alene district. Hobbs drew the contact where the alternating beds of quartzite and argillite of the Wallace Formation became more numerous than the laminated green argillitic rock of the St. Regis Formation. In the report area, this transition zone is about 100 feet thick and is well exposed below the upper Cottonwood Road east of Parker Mountain and in the streambed of Grouse Creek north of Bald Mountain.

Many of the transition beds are carbonate bearing, as are some of the beds for several hundred feet below and several thousand feet above the transition. The lower part of the formation is one of the main carbonate-bearing units in the western Belt Supergroup, but in the report area almost none of this rock resembles normal limestone or dolomite because the carbonate minerals usually make up less than 50 percent (average is about 20 percent).

As well as can be determined from the small isolated outcrops, the lower part of the formation appears to grade upward, with numerous local reversals, from predominantly carbonate-rich quartzite and siltite to siltite or argillite and lesser amounts of quartzite. The rocks in the upper part of the lower Wallace, particularly near the transition into the predominantly argillitic upper Wallace, contain less carbonate minerals than the lower part of the formation.

The unit is characterized by alternating beds of light-colored quartzite and siltite and dark argillite. Carbonate minerals are confined mostly to the quartzite and siltite. The thickness of these relatively coarse grained beds is highly variable not only from bed to bed but within a single bed as well. The most striking lateral changes in bed thickness appear to occur in carbonate-bearing quartzite, which, in the extreme, thins from 5 feet to a few inches within a distance of less than 10 feet. Most of the quartzite beds are thinner, however, and repeatedly pinch and swell. Hobbs, Griggs, Wallace, and Campbell (1965, p. 44) suggested that this boudinagelike appearance may be attributed to compaction adjustment and later deformation. The thicker quartzite and siltite beds show almost no internal stratification. Rarely do colored impurities define bedding within these layers, although the upper part of a coarser grained bed commonly grades into an overlying argillite layer not only by a decrease in grain size but more obviously by darkening in color.

The argillite layers, which contain most of the abundant sedimentary structures in the lower parts of the Wallace Formation, are dark gray to black and commonly finely laminated. Exposed bedding-plane surfaces generally have a blue-black phyllitic sheen. Unlike the very even continuous laminations in other Belt formations, the laminations in this part of the section are commonly discontinuous, disrupted by other sedimentary structures, or disrupted by differential compaction. Small discordant bodies of siltite or quartzite, which look like miniature clastic dikes or filled mud cracks, cut the laminated argillite, commonly at right angles. These bodies are generally less than 1 inch long and have been crenulated, probably by compaction. They are found throughout the Wallace Formation and in parts of the Striped Peak Formation, but are most abundant in the lower part of the former. Desiccation mud cracks and, to a lesser degree, ripple marks are common throughout the unit, and in places mud cracks are superimposed on ripple marks. Crossbedding was not seen, although fine cross-laminations were found in some of the darker fine-grained siltite beds.

UPPER PART

Partial sections of the upper part of the Wallace Formation crop out best on Deer Lake Mountain, the south flank of Parker Mountain, and the hill east of Quartzite Mountain. At these localities the rocks are predominantly composed of dark laminated argillite but also contain zones of relatively pure looking limy dolomite and beds of carbonate-bearing quartzite identical with those in the lower part of the formation. In this respect the lithology of the upper part in this area is intermediate between that at Pend Oreille Lake and in the Coeur d'Alene district. Harrison and Jobin (1963, p. K13) divided the Wallace Formation into five members in the Clark Fork and Packsaddle Mountain quadrangles near Pend Oreille Lake. There, the transition from the lower calcareous member to the argillite member, found about 2,500 feet above the base of the formation, appears to represent roughly the same interval as the transition from the lower part to the upper part of the Wallace Formation in the report area and in the Coeur d'Alene district. However, Harrison and Jobin (1963, p. K15) described additional beds resembling those of the lower part in the member above the argillite member, and a calcareous member above that, which is composed principally of dolomitic limestone. Although Hobbs, Griggs, Wallace, and Campbell (1965, p. 44) mentioned a few hundred feet of carbonate-rich beds in the upper part of the Wallace Formation, the section they described at Coeur d'Alene is made up predominantly of dark laminated argillite.

On Eagle Mountain, 3 miles northeast of Chewelah, the upper part of the Wallace Formation contains numerous beds, as much as 10 feet thick, of light-tan limy dolomite. In the two fault-bounded blocks on the northeast flank of the mountain, the unit is made up of about equal parts of black laminated argillite, laminated carbonate-bearing argillite, and light-tan dolomite beds. Here the various lithologies are repeated in groups from a few feet to a few tens of feet in thickness. The aggregate preserved thickness of this part of the unit is probably 800–1,000 feet. Although it is fault bounded, this carbonate-rich zone probably occurs at least 3,000 feet above the transition from the lower part of the formation because the zone is not found in the intervening interval at the relatively well exposed section east of Quartzite Mountain only 4 miles to the south.

East of Quartzite Mountain the unit is exposed better than at any other place in the area and is fairly similar to the upper part in the Coeur d'Alene district. It is predominantly dark-gray to black laminated argillite. There are a few quartzitic zones as much as 50 feet thick which are made up of individual quartzite or siltite beds as much as 1 foot thick. Lensoidal seams of brown siltite averaging about 0.1 inch in thickness are common throughout the unit. Where these lenses are thicker it is seen that they are channels cut in the argillite and are commonly graded.

On the north flank of Deer Lake Mountain, the upper part of the Wallace Formation is well exposed in large roadcuts above the south shore of Deer Lake. The rock there is highly quartzitic and carbonate bearing. It is identical with that in the lower part of the formation and was originally mapped by Miller (1969) as lower Wallace. Natural exposures are slightly better on the west flank than on the north. The carbonate-bearing quartzite beds on the west flank grade upward into typical black laminated argillite of the upper part of the formation. However, 200–300 feet above that the lowermost member of the Striped Peak Formation rests on the black argillite. If that contact is depositional, as it appears to be, then the quartzitic rock exposed in roadcuts along the south shore of the lake must be an anomalous zone in the upper part of the Wallace Formation.

Mud cracks, ripple marks, graded bedding, and, to a lesser degree, cross-laminations are found in this part of the section, although none are extensively developed. The crenulated clastic dikelets or crack fillings so abundant in the lower part of the formation are common but much less numerous. Slump structures and preconsolidation deformation, which appear in some of the similar looking argillites in the Striped Peak Formation, are not found in the upper Wallace Formation.

MISSOULA GROUP

STRIPED PEAK FORMATION

DISTRIBUTION AND COMPARISON WITH OTHER NEARBY SECTIONS

The Striped Peak Formation in the report area is confined for the most part to a narrow band on the east side of the Addy Quartzite, which trends roughly north to south from Eagle Mountain to Jumpoff Joe Mountain. Other outcrops are found on Deer Lake Mountain and possibly beneath the Addy Quartzite on the south flank of Blue Grouse Mountain.

The formation was originally named and described in the Coeur d'Alene district by Ransome (1905, p. 280, 282). Calkins (in Ransome and Calkins, 1908, p. 44) estimated the thickness of the preserved section to be about 1,000 feet. Shenon and McConnel (1939, p. 5) measured 1,500 feet of section about 1 mile southeast of Striped Peak, the type locality.

Although not particularly thick, one of the best and most complete sections of the Striped Peak Formation is well exposed in a large part of the Clark Fork quadrangle, Idaho. Harrison and Jobin (1963, p. K16) divided the formation there into four rather distinct units with a total thickness of about 2,000 feet. The lowest member is about 600 feet thick and composed predominantly of light-colored carbonate-bearing siltite, quartzite, and argillite. The second member is predominantly dolomite and is about 400 feet thick. Member three is composed of about 300 feet of dark well-laminated argillite and siltite, and the fourth, uppermost member consists mainly of distinctive red arkosic quartzite, about 700 feet thick. All but the lowest member correlate well with similar divisions in the report area.

STRATIGRAPHY AND LITHOLOGY

Although that part of the Belt Supergroup above the Wallace Formation in the report area is similar to the Striped Peak Formation at other places, the differences that exist almost justify using a new name here. For several reasons, however, not the least of which is the confusion already existing in Belt nomenclature, the rocks above the Wallace Formation in the report area are assigned to the Striped Peak Formation. Facies changes that are large and rapid relative to facies changes in the lower Belt formations are recognized in the Striped Peak Formation in western Montana and northern Idaho. Most of the differences between the Striped Peak Formation in the report area and the formation in the Clark Fork quadrangle and the Coeur d'Alene district can be explained by facies changes no greater than those described between localities of comparable separation at other places in Montana and Idaho.

The formation is divided into four members, which are designated by letters to avoid confusion with the numbered members in the Clark Fork quadrangle (Harrison and Jobin, 1963, pl. 1). A composite section is illustrated here (middle column, fig. 7A) because no complete and unfaulted section has been found in the report area. Many of the members are fault bounded in much of the area, and the contact between members *a* and *b* is a fault wherever found.

MEMBER *a*

On Jumpoff Joe Mountain and Deer Lake Mountain, the lowest member of the Striped Peak Formation is made up of thin- to medium-bedded siltite with interlaminated argillite partings. Another but smaller fault-bounded section of siltite on the mountain south of Beitey Lake is probably part of this member also. The siltite appears to conformably overlie the black well-laminated argillite of the upper part of the Wallace Formation on Deer Lake Mountain. Although the complete transition from one unit to the other is not exposed anywhere, a composite of the transition zone appears to be about 50–100 feet thick.

About 70 or 80 percent of the lower member is medium-gray to olive-gray siltite. Except for about 100 feet of argillite near the top of the member, the other 20–30 percent of the unit is dark-gray to black argillite, which occurs chiefly as thin bedding-plane partings and in beds less than 1 inch thick. Although they sometimes tend toward pink or dull green, the siltite beds do not vary much in color, perhaps because they have been thermally metamorphosed in varying degree by nearby younger plutonic rocks. On a weathered surface, some siltite beds show faint internal laminations, which are rarely seen on a fresh surface. Most siltite beds are about ¼–⅓ inch in average thickness, but a few quartzitic beds are as much as 3 feet thick. Because of its thin-bedded character, the unit forms platy talus slopes below prominent outcrops.

Approximately the upper 100 feet on Deer Lake Mountain is thinly laminated deep-reddish-purple or maroon argillite and siltite with thinly interbedded lensoidal layers of fine-grained quartzite. Interbedded near the top of the maroon zone are several platy-weathering light-gray to pale-lavender siltite layers 10 feet thick. The Addy Quartzite rests unconformably upon the maroon zone, indicating that some of the formation has been removed by erosion here. On Jumpoff Joe Mountain, pre-Addy erosion has removed the entire maroon zone, and on the mountain south of Beitey Lake the upper contact is a fault. The thickest section of this member appears to be on Deer Lake Mountain, where 900 feet of strata may be present.

Detrital mica is sparsely scattered through much of the rock in both siltite and argillite layers. Large polygonal mud cracks and thin zones of mud-chip breccia are common in all parts of the member except for the upper maroon zone. Small ripple marks are common but not abundant. Salt casts are common in the lower member of the Striped Peak Formation at other places but were not found in the report area.

Although many characteristics of this member are similar to those of the lower member in the Clark Fork quadrangle and in the Coeur d'Alene district, three characteristics are notably different and make it difficult to correlate member *a* with the equivalent part of the section at the other two localities. About 55 miles east in the Clark Fork quadrangle and 75 miles southeast in the Coeur d'Alene district, the lower member is generally red, pink, or green, or it is light gray tinted with one of these three colors (Harrison and Jobin, 1963, p. K17; and Hobbs and others, 1965, p. 45). The red color appears to be an important characteristic of the unit over a very large area. Except for the maroon zone in the upper part of the member on Deer Lake Mountain, the rocks of member *a* are darker and less colorful than those making up the lower member to the east. In addition to a color difference, the lower member in the Clark Fork quadrangle and in the Coeur d'Alene district is characterized by small quartz and calcite-filled vugs, and quartzitic layers which contain carbonate minerals. Although these features were carefully searched for, only local traces of carbonate minerals could be found in member *a*.

MEMBER *b*

Member *b* crops out in a narrow discontinuous belt from Eagle Mountain south to the Loon Lake copper mine, but is well exposed only on Parker Mountain and the mountain south of Cottonwood Creek. The unit is chiefly medium- to thick-bedded dolomite containing many argillaceous bedding-plane partings. Nowhere in the area is a complete section of the member exposed, as the lower contact is everywhere faulted. Because of internal faulting, a reliable composite section can be constructed for only part of the member. From the available exposures, however, considerable lithologic variety is obvious.

On Quartzite Mountain and Parker Mountain the contact with overlying member *c* is fairly well exposed in places. Noticeable differences in the upper part of member *b* are found along strike in the 1½ miles separating these two localities. On Parker Mountain the gradation down section from member *c* to *b* is from gray argillite into purple argillite and subsequently into purple or maroon dolomite. Locally, about 10–30 feet of light-gray dolomite separates the purple argillite from the purple or maroon dolomite. Lower in member *b* at this locality, the maroon pigmentation of the carbonate is increasingly pale and tends toward gray.

On Quartzite Mountain, numerous thin beds of pale-green carbonate-bearing siltite in zones as much as 10 feet thick are interlayered with the dolomite. The same lithologies are found south of Parker Mountain, where member *b* crops out about 1 mile north of Beitey Lake. Here, however, beds of maroon dolomite and carbonate-bearing siltite are interbedded with gray dolomite and green carbonate-bearing siltite, and the contact with member *c* is a fault.

From the south half of Parker Mountain to the southernmost exposures of member *b*, the upper and lower contacts of the unit are both faults. On the mountain south of Parker Mountain, most of the dolomite is medium gray and pale tan; maroon or green beds occur only on the south flank. The gray and tan beds probably represent the lowest part of the member exposed and are also found in the easternmost exposures of the member, near the top of Parker Mountain. In general, the most intense maroon coloring appears to be in the upper part of the member. The coloration decreases in intensity down section into the more abundant tan and gray beds. About half of the unit is pale tan or gray on a fresh surface, weathers tan or red brown, and forms a brick-red soil.

The thick carbonate beds contain closely spaced laminae of silica. Some of these films of silica are nearly perpendicular to bedding and on weathered surfaces commonly exhibit a boxwork structure like that described by Harrison and Jobin (1963, p. K18). In the few places where stromatolites have been found, the form of the fossil is commonly outlined and accentuated by thin films of silica.

The dolomite beds in this member are considerably purer as a whole than in any other Belt carbonate unit in the report area, but contain, even so, several percent sand, silt, and argillaceous material, in addition to the silica already mentioned. On the west flank of Parker Mountain, the dolomite contains an anomalously large amount of sand about 700–800 feet below the top of the member. The sandy zone is only about 100 feet thick and grades downward again to relatively pure carbonate beds with bedding-plane partings of argillite.

Calculated from outcrop width, the maximum thickness of that part of the unit still preserved is about 1,750 feet. Although the least disturbed section was chosen for the calculation, the figure may possibly be too large owing to undetected faults. The preserved part of the member is probably not less than 1,000 feet thick, however, which is considerably thicker than the 400 feet reported by Harrison and Jobin (1963, p. K18) for member 2 in the Clark Fork quadrangle.

Because of stratigraphic uncertainties stemming from the absence of a complete section of the Striped Peak Formation throughout the area, the possibility cannot be ruled out that the predominantly maroon carbonate of member *b* is separated by member *c* from the gray and tan carbonate. If member *c* were overlain by a maroon carbonate unit and underlain by a gray and tan carbonate unit, the stratigraphic section would resemble that shown in figure *7B*. Although correlations with other sections would not be impaired, and might even be improved, this interpretation would necessitate changing several structural interpretations. As best as can be determined, however, all the relatively clean carbonate rock is restricted to a single unit and is overlain by member *c*.

MEMBER *c*

Member *c* is better exposed than any other member of the Striped Peak Formation, and indeed better than any unit in the area with the exception of the Addy Quartzite. It is particularly well exposed on the south and west slopes of Quartzite Mountain east of Chewelah (figs. 4, 9). The member is composed chiefly of medium- to dark-gray laminated argillite interlayered with lighter gray siltite in beds generally less than 1 inch thick. The argillite beds are well laminated, but in some parts of the unit the laminations are obscure except on a weathered surface.

Most of the member is grayish argillite. About 500 feet above the base is 250–300 feet of light-gray, pale-yellow, and pale-green siltite and siltitic quartzite. Small flakes of detrital mica are found all through the rock of this zone but especially along bedding planes. Bedding averages 2–6 inches in thickness. The only other lithology that is notably different from the argil-

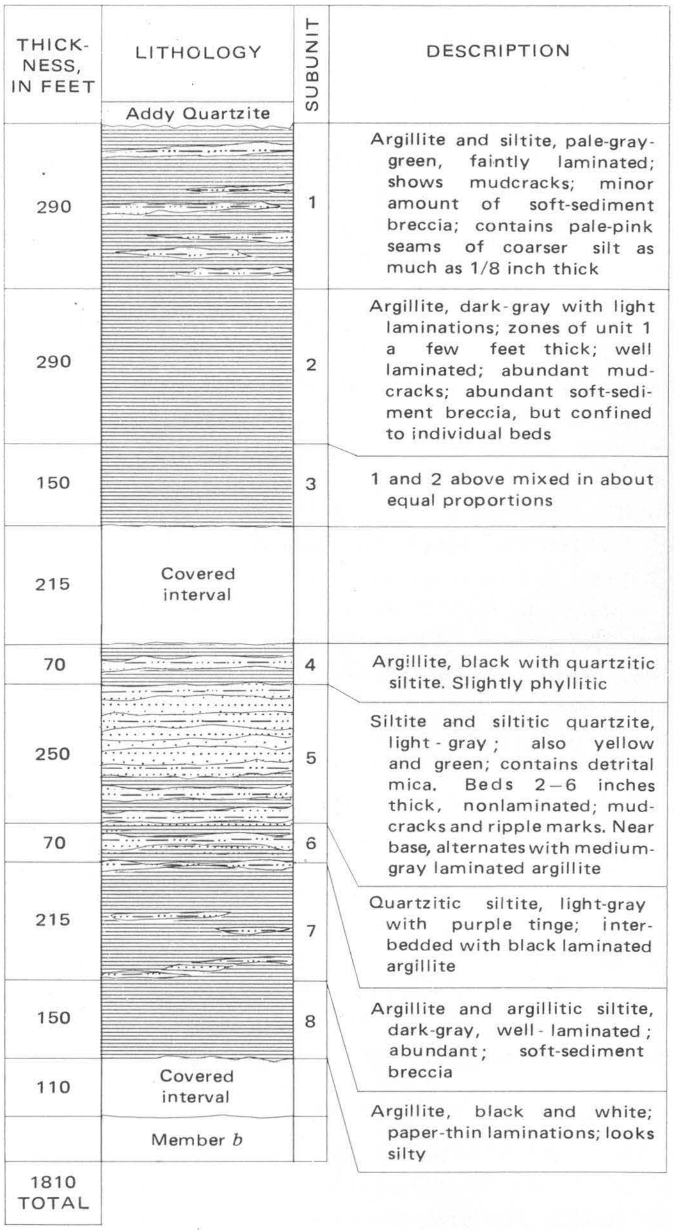


FIGURE 4.—Generalized columnar section of member *c* of the Striped Peak Formation on Quartzite Mountain.

lite is about 300 feet of pale-gray-green argillite and siltite in the uppermost part of the unit. The rock is characterized by thin pink lensoidal seams of coarser silt about one-eighth of an inch thick.

Sedimentary structures are abundant in this member. Mud cracks and small crenulated siltstone dikelets, or filled desiccation cracks, occur throughout the member. Graded bedding is well developed in the siltstone beds, especially in the upper 500–1,000 feet of section on Eagle Mountain, but most hand specimens show only a gradation from light to dark gray because the grain size is so small. Light-gray silt beds grade upward into black argillite beds. Cross-laminations are common but not obvious. Ripple marks are rare and confined almost entirely to the siltite and siltitic quartzite zone. Soft-sediment breccias are found throughout most of the member (figs. 5, 6). They are apparently the result of slumping or some other penecontemporaneous deformation that occurred before lithification. Most of the soft-sediment breccias on Quartzite Mountain are confined to individual beds and truncated at top and bot-

FIGURE 5.—Argillite and breccia from member *c* of the Striped Peak Formation. *A*, Laminated black argillite. Shows channeling, clastic dikes, and graded bedding. Lighter layers are graded from coarse to fine silt. Collected near top of Eagle Mountain. *B*, Soft-sediment breccia showing disrupted bedding and clastic dikes which presumably formed prior to lithification of the argillite. Collected near the top of Eagle Mountain about 100 feet away from specimens shown in *A* and about 20 feet stratigraphically above it.

tom by perfectly undisturbed beds, as if the disturbed zones acted as a décollement before lithification. On Eagle Mountain many of the beds bounding the soft-sediment breccias are disturbed also and may have formed under slightly different conditions. The disturbed beds there range in thickness from about 1 inch to more than 5 feet.

The two small hills south of the Loon Lake copper mine provide the best exposures of the uppermost part of member *c* and the transition into member *d*. The contact is placed at the base of the lowest thick series of maroon siltite beds. Some maroon siltite beds occur below this, but they are confined to zones less than 3 feet thick and are separated by thicker zones of dark-gray siltite and laminated argillite. Most of the beds in the upper 200 feet of member *c* contain detrital mica. Below the maroon rocks is a zone of finely laminated olive-gray siltitic argillite about 50–100 feet thick. These rocks appear to be highly susceptible to alteration and weather out in large thin plates. Below the laminated rocks is several hundred feet of interbedded dark argillite and siltitic quartzite. Some zones of pure poorly laminated black argillite are as much as 15 feet thick, and some quartzite beds as much as 5 feet thick. Below this, the rocks are similar to those that make up member *c* on the mountain north of Beitey Lake.

The 1,810-foot-thick section on Quartzite Mountain appears to be the thickest in the map area. On the north flank of the mountain, the thickness may reach 2,000 feet where pre-Addy erosion did not cut so deep, but the apparently thicker section could not be accurately measured, as the rocks are very poorly exposed. On Eagle Mountain incomplete exposure and possible structural complications combine to prevent an accurate estimate of the thickness of member *c*. The yellow- and green-tinted siltite and siltitic quartzite zone found on Quartzite Mountain could not be identified on Eagle Mountain, but may have been cut out along the fault bounding the unit on the east. Even so, the width of

B

FIGURE 5.—Continued.

FIGURE 6.—Soft-sediment breccia in the McHale Slate, probable correlative of member *c* of the Striped Peak Formation. Location is a few hundred feet west of the map area, about 4 miles north of Chewelah. Photograph illustrates how soft-sediment breccias are confined to distinct layers and are bounded by undeformed laminated argillite. Compare with figure 5.

outcrop on the south flank of Eagle Mountain suggests that the member is much more than 1,000 feet thick. At least part of this section is stratigraphically higher than the highest part of the member on Quartzite Mountain. A composite thickness of the member on both Quartzite and Eagle Mountains is probably 2,000–2,500 feet. If the uppermost parts of the member, such as found on the mountains north and south of Beitey Lake, are added to this figure, the minimum thickness of the composite section would be between 2,400 and 2,900 feet.

MEMBER *d*

The uppermost member of the Striped Peak Formation is exposed only on the mountains north and south of Beitey Lake and on the two hills south of the Loon Lake copper mine. (See fig. 10.) It is perhaps the most easily identified of all the Striped Peak members because of its deep-maroon to hematite-red color. The only divergence from this color is on the north flank of the hill south of Beitey Lake, where layers of black faintly laminated argillite and siltite in zones as much as 40 feet thick are interbedded with the maroon rock. The chief rock type is siltite, but argillite is common in thin layers and as bedding-plane partings. Thin beds of quartzite also are found throughout the unit, but they are not as abundant as in member 4 in the Clark Fork quadrangle. Bedding thickness ranges from about ¼–3 inches and averages about three-fourths of an inch. Internally some of the beds are faintly laminated. Because it is thin bedded, this unit almost always breaks into plates or chips.

A few feet of pale-green siltitic quartzite occurs in the transition between this and the underlying member *c*. These beds are similar to those between members 3 and 4 that Harrison and Jobin (1963, p. K19) described in the Clark Fork quadrangle.

The Cambrian Addy Quartzite unconformably overlies member *d* in the report area. The member is thickest, 500–800 feet, just southwest of Beitey Lake.

Ripple marks, mud cracks, and salt casts are well developed and abundant throughout the member. Cross-lamination, scour and fill, and graded beddings are found locally and are relatively small scale features.

DEER TRAIL GROUP

The rocks assigned to the Deer Trail Group are largely those that were originally called the Deer Trail Argillite by Weaver (1920, p. 59). Bennett (1941, p. 7) elevated the Deer Trail Argillite to group status, and one of the carbonate zones in the section to formational status. Campbell and Loofbourow (1962, p. F8), in a more detailed report on approximately the same region that Bennett studied, divided the Deer Trail Group into the Togo Formation, Edna Dolomite, McHale Slate, Stensgar Dolomite, and Buffalo Hump Formation.

All the formations are extensively exposed in the Huckleberry Mountain southwest of Chewelah. In that area, the Deer Trail Group and younger rocks form what has become known as the magnesite belt for its large deposits of magnesite. This region is referred to frequently in the following formation descriptions because the various units are much more extensively exposed there than in the report area. For a more complete description of the Deer Trail Group, the reader is referred to Bennett (1941), Campbell and Loofbourow (1962), Becraft and Weis (1963), and Campbell and Raup (1964).

The assignment of rocks in the report area to the Deer Trail Group is questionable because the rocks closely resemble the upper part of the Belt Supergroup. Without exception, however, all the rocks so assigned are overlain or intruded by the greenstone of the Huckleberry Formation. The greenstone has not been found associated with the known Belt rocks east of the main north-south belt of Cambrian Addy Quartzite. Therefore, where considerable doubt exists, the presence of associated greenstone is used as a criterion in assigning rocks to the Deer Trail Group.

Pronounced facies changes in the Deer Trail Group over relatively short distances are well documented in the magnesite belt by Campbell and Loofbourow (1962, pl. 2) and Campbell and Raup (1964), and equally large changes, but on a regional scale, are found in the upper Belt Supergroup (A. B. Griggs, oral commun., 1969; Harrison and Campbell, 1963, p. 1424). Therefore, slight differences between the Deer Trail units in the report area and the type formations in the magnesite belt are less remarkable than would be similar differences between two sections in the lower Belt Supergroup.

All workers who have studied these rocks since Weaver have considered both the Deer Trail Group and Huckleberry Formation to be Precambrian in age.

The Deer Trail Group is truncated by intrusive rocks north of Chewelah but continues on the other side of the batholith in the Metaline quadrangle with the same unvarying northeast trend that it has in the magnesite belt. Possible correlatives of the Deer Trail Group in the Metaline quadrangle have been called the Priest River Group (Park and Cannon, 1943, p. 6), and the same rocks in Canada have been called the Priest River Terrane (Daly, 1912, p. 258). Becraft and Weis (1963, p. 16) were the first to suggest a correlation between the Priest River and Deer Trail Groups but were unable to match individual units.

TOGO(?) FORMATION

Rock that may belong to the Togo Formation underlies only one small patch in the report area, about 2 miles east of Valley. The rock is sheared dark-gray-green argillite which looks identical with that found in at least five other Precambrian formations in the area. It is assigned to the Togo Formation only because it appears to underlie the Edna Dolomite.

The lithology is similar to that making up part of the type Togo Formation in the magnesite belt to the west (Campbell and Loofbourow, 1962, p. F9). Campbell and Loofbourow (1962, p. F8) named the unit for the exposures around the Togo mine about 15 miles due west of Springdale. There, the formation consists of several thousand feet of dark-gray to black laminated argillite. The upper part of the formation consists of about 1,000 feet of relatively pure quartzite at the latitude of Valley but thins to the north. Both Campbell and Loofbourow (1962, p. F9) and Campbell and Raup (1964) noted that the argillitic part of the formation is carbonate bearing near the top. No accurate estimate of the thickness of the Togo Formation has been made because of inability to detect structure in the monotonous section of argillite. Campbell and Loofbourow (1962, p. F9) estimated that the unit is at least 4,000 feet thick in the southern part of the area, but it may be much thicker. Becraft and Weis (1963, p. 7) estimated the thickness from outcrop width in the same general area to be about 20,000 feet.

EDNA DOLOMITE

Light-gray to tan dolomite crops out on three small hills immediately west of the Togo (?) Formation in sec. 18, T. 31 N., R. 41 E., northeast of Valley. It is identical in all respects with the type Edna Dolomite which was named and described by Campbell and Loofbourow (1962, p. F9). The rock is medium- to thin-bedded impure dolomite, but it has a massive appearance. It weathers light tan and forms a bright-hematite-red soil. A completely fault-bounded discontinuous belt that extends about 4 miles north from Chewelah also is considered to be a part of the Edna. There, the lithologies are the same as those northeast of Valley, except for a zone of vitreous quartzite about 300 feet thick in the lower part of the exposed section and several zones of argillite as much as 50 feet thick interbedded with the dolomite. Only small partial sections are exposed in the report area, and so the stratigraphy cannot be described in detail.

Campbell and Loofbourow (1962, p. F10) reported a zone of slate near the middle of the formation in the magnesite belt and probably more than one zone of vitreous quartzite. Campbell and Raup (1964) noted several discontinuous quartzite zones in the Hunters quadrangle, which covers most of the southern part of the magnesite belt. In the report area quartzite and argillite zones were found only northeast of Chewelah and could not be matched unit for unit with those in the magnesite belt.

The west boundary of the formation is a fault in this area, and so only a minimum thickness can be estimated. From outcrop width, the preserved section is calculated to be 850 feet thick. Campbell and Loofbourow estimated that the thickness is between 1,500 and 2,500 feet in most of the magnesite belt but could not accurately measure it because of poor exposures.

McHALE SLATE

Black, gray, and green laminated argillite underlies about half a square mile just east of Valley and is considered to be part of the McHale Slate. As with all units of the Deer Trail Group in this area, the assignment is not above question, but it is the best correlation based upon available evidence. The argillite resembles the type McHale Slate more closely than it does any other formation in the Deer Trail Group and displays many of the distinguishing characteristics of the formation.

The McHale Slate was named by Campbell and Loofbourow (1962, p. F11) for the exposures in McHale Canyon, 4 miles west of Chewelah. There the unit is between 1,000 and 1,500 feet thick, as it is in most of the magnesite belt. The dominant lithology, which varies little, is gray, green, and brown laminated argillite. The McHale Slate contains numerous beds of soft-sediment breccia, which were not described by Campbell and Loofbourow. The breccia resembles that found in member *c* of the Striped Peak Formation. The unit also contains numerous thin graded beds composed of silt- and clay-sized particles.

In the report area the rocks of this unit are generally phyllitic, although bedding is usually recognizable. The formation is highly faulted, and so no section is complete. One mile east of Valley about 2,000 feet of argillite is preserved between two faults but is not well exposed; it also may be faulted internally. This apparently is the thickest possible section of McHale Slate in the map area.

STENSGAR DOLOMITE

The Stensgar Dolomite crops out north of Valley and southwest and north of Chewelah. The lowest contact of the unit is nowhere exposed. It is either concealed by younger rocks or cut off by faulting. The upper contact on the hill north of Valley is gradational into the argillite and siltite of the overlying Buffalo Hump(?) Formation. In that area, the dolomite is predominantly light gray or light tan with interbeds of maroon dolomite, argillite, or siltite. Bedding thickness ranges from 1 inch to about 2 feet, and internal laminations or algal structures are generally present in the thicker beds. Maroon argillaceous partings occur between some beds but are not numerous.

Weaver (1920, p. 58) named the Stensgar Dolomite and included it in his Deer Trail Argillite. Bennett (1941, p. 7) changed the name Deer Trail Argillite to Deer Trail Group and elevated the Stensgar Dolomite to a formation. The magnesite in the area is confined to this formation, which is found almost the entire length of the magnesite belt. Thickness ranges from 300 to 1,200 feet in the magnesite belt. There it is chiefly a fine-grained light-blue or pinkish-gray dolomite, which is thin bedded except where extensively recrystallized (Campbell and Loofbourow, 1962, p. F13). Campbell and Loofbourow (1962, F15) noted that a zone of reddish dolomitic slate is commonly found near the top of the unit and that a similar zone about 100 feet thick was also found at the base in a measured section at the north end of the belt.

The thickness of the formation in the report area cannot be accurately estimated, owing to poor exposure. All the exposed sections are incomplete; 1 mile southwest of Chewelah about 500 feet of the unit is preserved. North of Valley a partial section of the unit is about 1,000 feet thick, as calculated from outcrop width, but is probably faulted internally.

BUFFALO HUMP(?) FORMATION

About 1,000 feet of argillite and siltite on the hill north of Valley is questionably assigned to the Buffalo Hump Formation. The name Buffalo Hump was given by Campbell and Loofbourow (1962, p. F17) to a thick section of argillite and quartzite that overlies the Stensgar Dolomite in the magnesite belt. They estimated that as much as 3,000 feet of the unit is preserved near the middle of the magnesite belt but found that it thins to the northeast and southwest because of erosion prior to deposition of the Huckleberry Formation.

The formation consists chiefly of quartzite and argillite and exhibits striking facies changes in the magnesite belt. Where it intersects the Hunters-Springdale road, the unit is about 25 percent quartzite and 75 percent argillite. Only 5 miles along strike to the southwest, the proportion of quartzite is 85 percent (Campbell and Raup, 1964, calculated from their map). Much of the quartzite is confined to zones and is vitreous, medium to fine grained, and light colored. The argillite is slaty and dark gray; much of it is not laminated, unlike most of the argillite in the Deer Trail Group (Campbell and Loofbourow, 1962, p. F18, F20).

In the report area, most of the rocks assigned to this unit are well laminated or thin bedded; about 100 feet of siltitic argillite is medium to thick bedded. The rock is primarily medium to dark gray, but has a faintly reddish cast locally. Vitreous quartzite beds are not present, and in this respect the rocks differ from those in the magnesite belt. The rocks in the report area are assigned tentatively to the Buffalo Hump Formation chiefly because they conformably overlie the carbonate unit containing maroon dolomite, which is considered part of the Stensgar Dolomite.

On the hill north of Valley, the upper contact of the unit appears to be faulted, but at least 1,000 feet of section is preserved and fairly well exposed. This figure is probably an accurate estimate of the thickness of the unit preserved in the area.

DEER TRAIL GROUP, UNDIVIDED

Rocks presumably belonging to the Deer Trail Group crop out on the hill north of Springdale and north of that on the mountain west of Jumpoff Joe Lake. The rock at both localities is not distinctive enough to assign to any specific formation.

Black laminated argillite and some thin-bedded siltite make up the section on the mountain west of Jumpoff Joe Lake. These rocks most closely resemble the McHale Slate or parts of the Togo Formation. The rocks north of Springdale are about the same, except for a much higher proportion of siltite to argillite.

RELATION BETWEEN THE BELT SUPERGROUP AND THE DEER TRAIL GROUP

From the preceding descriptions it is apparent that both the Belt Supergroup and Deer Trail Group are thick Precambrian sections composed chiefly of fine-grained clastic rocks which usually exhibit a very low grade of metamorphism. Part of the Belt Supergroup is strikingly similar to the Deer Trail Group and may be an approximate time equivalent. There are, however, obvious differences between these two sections, and their proximity to one another necessitates either facies changes or a large structural discontinuity.

The Belt Supergroup underlies many thousand square miles and has been studied by many geologists, a few of whom, such as Schroeder (1952, p. 19) and Ross (1963, p. 88), briefly considered the relation between the Belt and Deer Trail. In all previous studies, however, investigators were more handicapped than us in attempting correlations, because the sections compared were relatively far removed and, as a result, they were unable to make unit-by-unit correlations.

Schroeder (1952, p. 18–20) compared the rocks in the Bead Lake area with both the Deer Trail Group in the magnesite belt and the Belt Supergroup in the Clark Fork district. He recognized differences between the two sections and correctly concluded that the rocks in the Bead Lake area, which he named the Newport Group, were more similar to the Belt Supergroup. Campbell and Loofbourow (1962, p. F20) noted that the similarity in lithology and sedimentary structures and the great thickness of the Deer Trail Group combined to make "the resemblance to the Belt Series [Supergroup] striking." Although they discussed the problem more thoroughly than anyone had before, they were unable to make unit-by-unit correlations. The correlation of the two sections was also discussed, briefly, by Becraft and Weis (1963, p. 17).

We have drawn heavily on the valuable work by these men and have combined it with our own data to make tentative unit-by-unit correlation between the Deer Trail Group and part of the Belt Supergroup. The correlations proposed here are tentative, mainly because of the composite nature of the Striped Peak section in the report area, but also because of major stratigraphic and structural differences between these two thick sections and the units overlying them. Figure 7 summarizes the possible correlations between the Deer Trail Group of the magnesite belt and the Belt Supergroup of the report area and also shows the relation of the Belt of the report area to the Striped Peak Formation of the Clark Fork quadrangle. The Striped Peak units of the report area that are shown in figure 7 are pieced together as accurately as possible, but faults that interrupt the section may introduce errors in thickness and even in the order of superposition. Figure 8 shows the present spatial relations between the Deer Trail Group and the Belt Supergroup in the report area.

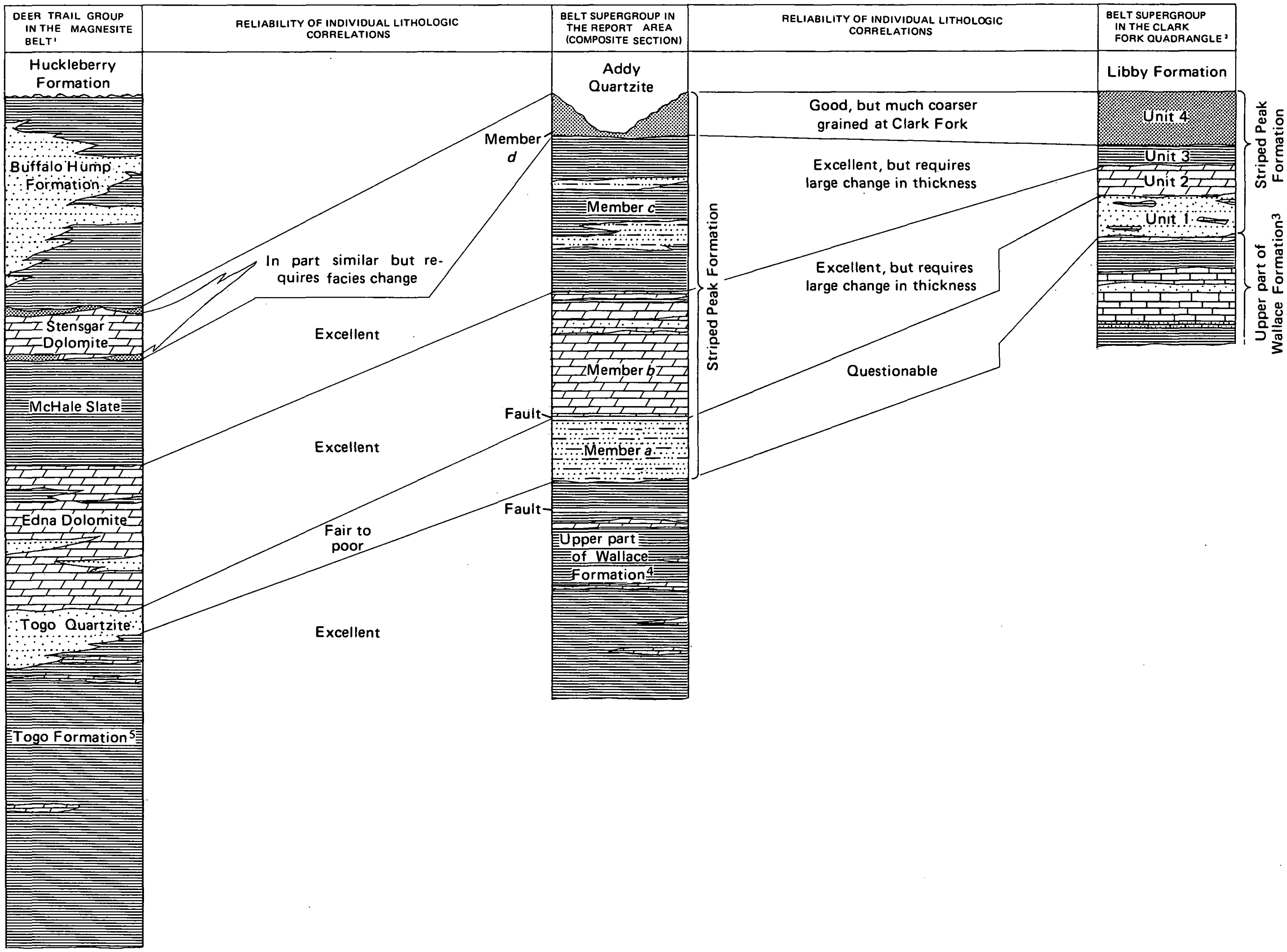
In both the Belt Supergroup and Deer Trail Group, the basal formation is considered at least twice as thick as any other formation in that section. The Togo Formation has been estimated to be as thick as 20,000 feet, and the only Belt unit to which it could possibly be compared on the basis of both thickness and lithology is the Prichard Formation.

Notably contrasting lithologic differences make correlations between the Prichard and Togo Formations suspect, even if the Togo is considered to be as thick as the Prichard. Although both are made up of argillite, siltite, and quartzite, the relative amounts in the two formations are considerably different. The Togo Formation is composed chiefly of dark-gray argillite with local, discontinuous carbonate zones and has a medium-grained quartzite zone as much as 1,000 feet thick at the top. The Prichard Formation has several distinct argillite and quartzite zones; all the quartzite is fine grained, and most of it is 4,000 feet or more below the top of the unit (fig. 3). Both quartzite and siltite are more abundant in the Prichard Formation, and iron sulfides, which are characteristically found in the unit, have not been described in the Togo Formation.

The obvious lithologic dissimilarities between the formations above these basal units may have tended to discourage previous attempts at unit-by-unit correlation. Neither the Edna Dolomite nor the Stensgar Dolomite, both relatively pure compared with Belt carbonates, resembles any of the 10,000–18,000 feet of Belt section between the Prichard and Striped Peak Formations. The only other unit in the Belt Supergroup that is lithologically comparable with the Togo Formation is the upper part of the Wallace Formation, which is probably only slightly more than 2,500 feet thick, thus considerably thinner than either the Prichard Formation or Togo Formation. Nevertheless, units overlying the Togo and upper Wallace Formations correspond well. If this correspondence is meaningful and the Togo Formation and the upper part of the Wallace Formation are correlative, then the Togo Formation may not be as thick as was estimated. Indeed, evidence of another sort suggests that the Togo Formation is considerably less than 20,000 feet thick.

Almost all estimates of the thickness of the Togo Formation are based on the excellent work of Bennett (1941), Campbell and Loofbourow (1962), Becraft and Weis (1963), and Campbell and Raup (1964), who are chiefly responsible for the present understanding of the Deer Trail Group. Although each of these investigators has provided additional refinements on earlier work, any study of the Togo Formation in the magnesite belt is hampered by the incredibly poor exposure and a lack of any traceable marker units other than the quartzite at the top. This combination makes it impossible to systematically delineate structures in this thick argillite unit. The result is well illustrated by the wealth of complex structures which have been mapped in the rest of the Deer Trail Group (Campbell and Loofbourow, 1962, pl. 1; Campbell and Raup, 1964) and the almost total absence of faults and folds shown in the Togo Formation, even where attitudes are locally highly divergent. It would seem reasonable that faults and folds are as numerous in the Togo Formation as in the rest of the Deer Trail Group and that the formation may therefore be thinner than previously estimated. If the fault density in the rest of the Deer Trail Group is superimposed on the Togo Formation, a thickness of 3,000 feet could more than adequately account for the observed outcrop width.

The Togo Formation and the upper part of the Wallace Formation are similar enough lithologically to be considered correlative. Both are thick sections made up chiefly of laminated dark-gray argillite, both contain carbonate-rich zones, and both are overlain by sections containing a significant proportion of relatively coarse


DEER TRAIL GROUP IN THE MAGNESITE BELT[1]
RELIABILITY OF INDIVIDUAL LITHOLOGIC CORRELATIONS
BELT SUPERGROUP IN THE REPORT AREA (COMPOSITE SECTION)
RELIABILITY OF INDIVIDUAL LITHOLOGIC CORRELATIONS
BELT SUPERGROUP IN THE CLARK FORK QUADRANGLE[2]
Huckleberry Formation
Buffalo Hump Formation
Stensgar Dolomite
McHale Slate
Edna Dolomite
Togo Quartzite
Togo Formation[5]
In part similar but requires facies change
Excellent
Excellent
Fair to poor
Excellent
Addy Quartzite
Member d
Member c
Member b
Fault
Member a
Fault
Upper part of Wallace Formation[4]
Striped Peak Formation
Good, but much coarser grained at Clark Fork
Excellent, but requires large change in thickness
Excellent, but requires large change in thickness
Questionable
Libby Formation
Unit 4
Unit 3
Unit 2
Unit 1
Striped Peak Formation
Upper part of Wallace Formation[3]

A

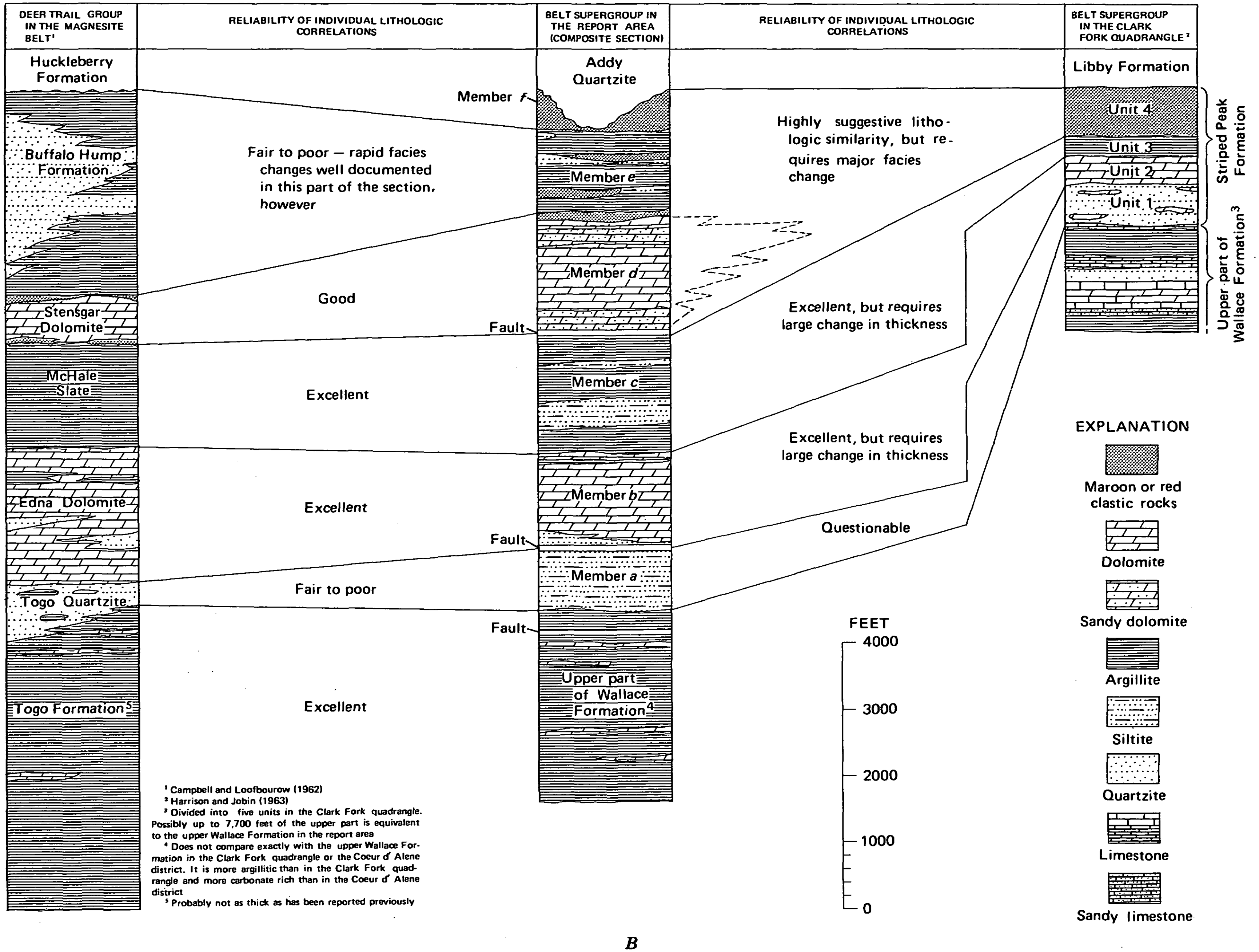
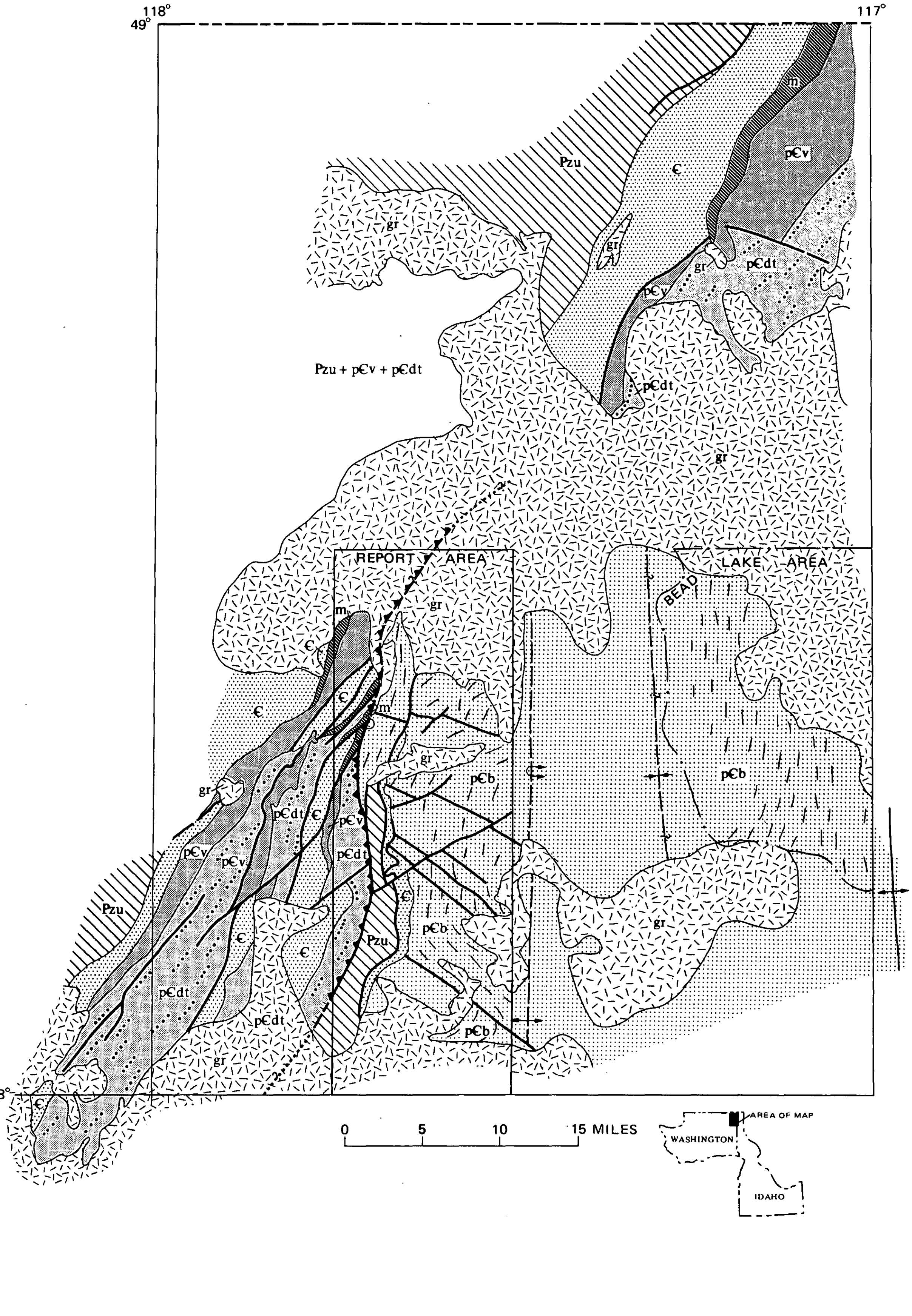


FIGURE 7.—Possible correlations between the Deer Trail Group and the Belt Supergroup. *A*, A single carbonate unit assumed to occur in the Striped Peak Formation of the report area. Compare with *B;* see text for discussion. *B*, Two separate carbonate units assumed to occur in the Striped Peak Formation of the report area. See text for discussion.


118°
117°
49°
48°
Pzu
€
m
p€v
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p€dt
Pzu + p€v + p€dt
REPORT AREA
LOON LAKE AREA
BEAD
p€b
0 5 10 15 MILES
AREA OF MAP
WASHINGTON
IDAHO

EXPLANATION

gr

Tertiary and Mesozoic intrusive rocks

Pzu

Paleozoic rocks, undivided

Є

Cambrian rocks

Mostly Addy Quartzite and Gypsy Quartzite, but includes Maitlen Phyllite and some Metaline Formation in places

m

Monk Formation

pЄv

Huckleberry Formation, Leola Volcanics, and Shedroof Conglomerate

pЄdt

Deer Trail Group and Priest River Group

pЄb

Belt Supergroup

Contact

Fault

Dashed where approximately located

?......?....

Fault separating Deer Trail Group and Belt Supergroup

Dashed where approximately located; dotted and queried where concealed or destroyed by younger rocks. Sawteeth on presumed upper plate; hachured where fault is downdropped by younger faults

Normal Overturned

Axis of anticline, showing direction of dip of limbs

Dashed where approximately located

? ?

Axis of syncline, approximately located

Queried where concealed

Deer Trail Group Belt Supergroup

Generalized strike of strata where reliable data are available

FIGURE 8.—Generalized geologic map showing the distribution of the Belt Supergroup and the Deer Trail Group in and around the Chewelah–Loon Lake area. No postintrusive rocks shown. Geologic data modified from Huntting, Bennett, Livingston, and Moen (1961), Schroeder (1952), Park and Cannon (1943), Bennett (1941), Campbell and Loofbourow (1962), and Campbell and Raup (1964).

clastics and carbonate rock. Unfortunately, distinguishing internal characteristics are generally absent in both units. Although only about 2,500–3,000 feet of the upper Wallace Formation is preserved in the report area, the unfaulted thickness before erosion could have been somewhat greater.

The quartzite at the top of the Togo Formation has a counterpart in member *a* of the Striped Peak Formation, although the lithologic correlation is not as close as that between the other units correlated. The Togo quartzite is chiefly light-colored medium- to fine-grained quartzite with a small amount of interbedded argillite. Member *a*, however, is predominantly darker siltite interbedded with some argillite. The Togo quartzite contains ripple marks and mud cracks similar to those that are so abundant in member *a*, but the two units differ from the others correlated in the two sections in that they are not look-alikes. However, the Togo quartzite exhibits major facies and thickness changes within the magnesite belt. Equal or larger differences could exist between the magnesite belt and the report area, especially since the Precambrian sections of the two areas appear to have been juxtaposed by a thrust fault and thus were probably widely separated when the rocks were deposited.

Member *a* does not closely resemble member 1 of the Clark Fork and Coeur d'Alene sections either, even though both are siltites. Rocks similar to those in member 1 thin from a thickness of over 1,500 feet south of the Coeur d'Alene district (Shenon and McConnel, 1939, p. 5) to about 600 feet in the Clark Fork quadrangle (Harrison and Jobin, 1963, p. K17), but the units are lithologically similar at both localities. If member 1 varies so greatly in thickness from north to south, its lithologic character may possibly change slightly to the west, and member *a* and the Togo quartzite may be reflections of this change. Because only part of member *a* is preserved in the report area, it is also possible that the carbonate-bearing siltite, which characterizes unit 1 to the east, was deposited in the report area but because of faulting is no longer exposed at the surface.

A remarkably good unit-by-unit match is evident between the remainder of the Deer Trail Group and the Striped Peak Formation. The Edna Dolomite is indistinguishable from the light-tan and gray carbonate that characterizes all but the upper part of member *b*. The McHale Slate is almost identical with member *c* and even displays the same type of soft-sediment deformation that characterizes the member. Differences in appearance result largely from the greater dynamic metamorphism of the McHale Slate. Although no carbonate unit of significant thickness which could correlate with the Stensgar Dolomite is known to overlie member *c*, the maroon argillite, siltite, and quartzite of member *d* might be compared with the maroon argillite that bounds the Stensgar Dolomite and seems to thicken toward the northeast end of the magnesite belt.

The Buffalo Hump Formation at the top of the Deer Trail Group has no equivalent in the Belt section of the report area, as the Cambrian Addy Quartzite rests unconformably on the eroded surfaces of member *d* and older rocks. The Buffalo Hump Formation may be roughly the same age as parts of the Missoula Group that A. B. Campbell (1960, p. 560) described in the St. Regis–Superior area to the east or as parts of the Libby Formation. Either of these correlations would imply considerable changes in facies and thickness. Within the magnesite belt, the Buffalo Hump Formation changes from predominantly argillite at the northeast end to predominantly quartzite at the southwest end (Campbell and Loofbourow, 1962, pl. 2; Campbell and Raup, 1964); thus major facies changes do occur in this part of the section.

METAMORPHIC ROCKS, UNDIVIDED

Deformed mica schist injected by leucocratic dikes and sills occurs in several small areas in sec. 29, T. 34 N., R. 41 E., and sec. 34, T. 35 N., R. 41 E. These highly metamorphosed rocks cannot be reliably assigned to any formation. In addition, they lie near the boundary between the Deer Trail Group and Belt Supergroup, and so they cannot even be assigned to one or the other.

Highly recrystallized quartz-mica schist, quartzite, and amphibolite underlie a narrow strip about 9 miles long between Nelson and Goddards Peaks along the east border of the report area. All the rock in this area was derived from the Prichard Formation. It is mapped separately from the Prichard Formation because a short distance east of the report area it is not possible to distinguish from which formation the metamorphic rocks were derived. The metamorphic rocks, which are intruded by many large and small bodies of two-mica quartz monzonite, extend to the west side of Pend Oreille Valley, 6 miles east of the report area. The extreme recrystallization of the rocks was caused by the numerous bodies of two-mica quartz monzonite. The contact with the Prichard Formation cannot be located precisely because a gradational zone as much as 1 mile wide separates the units and because there are very few exposures east of the divide separating Nelson and Goddards Peaks.

WINDERMERE GROUP

HUCKLEBERRY FORMATION

A thick section of conglomerate overlain by an equally thick section of greenstone constitutes the Huckleberry Formation in the magnesite belt. The formation rests unconformably on the Deer Trail Group there. The two lithologies and their stratigraphic relation to the rocks above and below them were first recognized by Jones (1928, p. 115), who referred to them collectively as the greenstone phase of the Chewelah Argillite. Weaver (1920, pl. 1) earlier differentiated small areas of greenstone locally but at most places included both conglomerate and volcanic rock in the Addy Quartzite. Bennett (1941), p. 8) divided the conglomerate and greenstone and described them in detail, but he assigned these rocks to the Lower Cambrian. He recognized the major unconformity between the conglomerate and the underlying Deer Trail Group and coined the names Huckleberry Conglomerate and Huckleberry Greenstone. Campbell and Loofbourow (1962, p. F24) showed that the contact between the greenstone and Addy Quartzite is also an unconformity and assigned the conglomerate and greenstone to the Precambrian. They gave these two units member rank in a formation they named the Huckleberry Formation. That usage is followed in this report.

The Huckleberry Formation in the report area is made up almost entirely of the greenstone member. Conglomerate, which is found only on the hill southwest of Chewelah, was not differentiated from the greenstone on the geologic map. The largest area underlain by the formation is about 8 miles due north of Chewelah. Here the greenstone is faulted against the Addy Quartzite on the east and is unconformably overlain by the Monk Formation on the west. It is not known whether the absence of the conglomerate at this locality is due to faulting or to nondeposition, but the latter is suspected. On their geologic map of the magnesite belt, Campbell and Loofbourow (1962, pl. 1) showed the formation thinning toward the southeast, east, and northeast, and as the formation is thin on the hill 1 mile southwest of Chewelah, it may wedge out a few miles north of the town.

The large area of greenstone north of Chewelah consists of flows, breccias, and a minor amount of light-colored tuff. Petrographically, much of it resembles volcanic breccias and aquagene tuffs in British Columbia that Carlisle described (1963, p. 57). Fine-grained dark-green basalt appears to be the most common rock type in the greenstone. Although these rocks are mildly chloritized and epidotized, the primary igneous textures are perfectly preserved in many places. In several specimens, pyroxene crystals (augite $Ny=1.686$, $2V=54°$) are completely unaltered, although almost all plagioclase reflects a low-grade metamorphism in that it is much more sodic ($An_{8\pm2}$) than an unaltered basalt should be. The Na_2O content of the rock, however, is too low for a spilitic rock, although alkalies may have been removed during metamorphism.

Analyses (table 1) of three samples considered to be representative of the basalt flows and an analysis of a fine-grained tuff show that the rock is chemically a basalt but contains unusual amounts of a few components. The total iron content is high, and the calcium low. The analyses are most like those of a tholeiitic basalt (Nockolds, 1954, p. 1030), except that the silica content averages 1 or 2 percent low. Normative quartz is calculated in three of the specimens because of the extremely low alkali content.

Whereas almost no planar structures, primary or secondary, are developed in the greenstone north of Chewelah, southwest of the town both the greenstone and the conglomerate are cut by a well-developed cleavage which imparts a phyllitic character to the rock and in most places masks any bedding or layering that may have been present.

TABLE 1.—*Chemical analyses and CIPW norms, in weight percent, of basalt and one tuffaceous rock from the greenstone of the Huckleberry Formation*

[Analysts: P. L. D. Elmore, S. D. Botts, Lowell Artis]

	1	2	3	4	5
Chemical analyses					
SiO_2	42.6	43.8	44.8	47.5	....
Al_2O_3	12.7	12.0	13.2	13.4	....
Fe_2O_3	.95	1.6	1.0	2.4	....
FeO	13.5	12.2	11.5	10.9	....
MgO	7.3	7.1	6.3	7.5	....
CaO	5.0	7.8	6.6	6.7	....
Na_2O	1.9	.95	2.8	2.1	....
K_2O	.59	.28	.85	.9	....
H_2O^-	.14	.15	.18	.43	....
H_2O^+	5.0	5.7	4.3	4.4	....
TiO_2	3.5	3.2	2.8	2.3	....
P_2O_5	.72	.54	.45	.25	....
MnO	.20	.21	.21	.22	....
CO_2	5.5	4.3	4.8	.24	....
Total	100.60	99.83	99.79	99.24	....
Chemical analyses (calculated on H_2O- and CO_2-free basis)					
SiO_2	47.9	48.8	49.5	50.4	48.7
Al_2O_3	14.3	13.4	14.6	14.2	14.1
Fe_2O_3	1.1	1.8	1.1	2.6	1.3
FeO	15.2	13.6	12.7	11.6	13.8
MgO	8.2	7.9	7.0	8.0	7.7
CaO	5.6	8.7	7.3	7.1	7.2
Na_2O	2.1	1.1	3.1	2.2	2.1
K_2O	.66	.31	.94	.96	.64
TiO_2	3.9	3.6	3.1	2.4	3.5
P_2O_5	.81	.60	.50	.27	.64
MnO	.22	.23	.23	.23	.23
Total	99.99	100.04	100.07	99.96	99.91
CIPW norms (calculated on H_2O- and CO_2-free basis)					
Q	1.4	6.3	....	2.0	....
c	1.8	....	....	....	....
or	3.9	1.8	5.6	5.6	....
ab	18.1	9.0	26.2	18.8	....
an	22.6	30.8	23.1	20.0	....
wo	....	3.5	4.1	3.2	....
en	20.4	19.7	12.4	19.8	....
fs	20.9	18.0	12.6	15.6	....
fo	....	....	3.5	....	....
fa	....	....	3.9	....	....
mt	1.5	2.6	1.6	3.7	....
il	7.5	6.8	5.9	4.6	....
ap	1.9	1.4	1.2	.6	....
Total	100.0	99.9	100.1	99.9	....
Percent normative an in plagioclase	55.7	77.2	47.0	58.8	59.6

1. Fine-grained basalt, 700 ft E., 2,100 ft N. of SW cor. sec. 26, T. 34 N., R. 40 E.
2. Fine-grained basalt, 1,050 ft E., 1,850 ft N. of SW cor. sec. 26, T. 34 N., R. 40 E.
3. Fine-grained basalt, 1,500 ft E., 1,550 ft N. of SW cor. sec. 26, T. 34 N., R. 40 E.
4. Fine-grained tuff, 4,200 ft E., 1,700 ft N. of SW cor. sec. 26, T. 34 N., R. 40 E.
5. Average composition of the three basalts.

Attitudes in the large greenstone area north of Chewelah are difficult to obtain because of the almost complete lichen cover and the gradational character of the contacts between flows and breccia horizons. At the few locations where attitudes could be determined with relative certainty, the sequence appears to strike approximately north-south and to dip westward at low angles. In at least one place in the eastern part of the outcrop area, the greenstone dips eastward; thus the great width of the outcrop may be due to a broad open fold.

Because of the difficulty in determining attitudes in the greenstone the thickness of this formation cannot be accurately estimated. The few attitudes available were used in drawing a cross section across the strike of the large outcrop area north of Chewelah. If the cross section is correct, the greenstone must be at least 1,200 feet thick to account for the outcrop width. As the base is not exposed, the unit could easily be much thicker.

In several small areas 6 miles north-northeast of Chewelah and at several places north and east of Valley, greenstone appears to intrude the rocks of the Deer Trail Group. Similar dikelike bodies of greenstone too small to show on the map are found on the hill west of Jumpoff Joe Lake. All these intrusive bodies are probably related to the extrusive rocks of the Huckleberry Formation. Some caution must be exercised in assigning greenstone, whether intrusive or extrusive, to the Huckleberry Formation, for similar-appearing greenstone of Paleozoic age is present 17 miles southwest of Chewelah in the Hunters quadrangle (A. B. Campbell, written commun., 1965).

In the Metaline quadrangle, about 40 miles northeast of Chewelah, Park and Cannon (1943, p. 7–11) described some 5,000 feet of conglomerate that unconformably overlies argillite of the Priest River Group. The conglomerate is in turn overlain by some 5,000 feet of greenstone. They applied the names Shedroof Conglomerate and Leola Volcanics to these two units and correlated them with the formation Daly (1912) named Irene Conglomerate (later named the Toby Conglomerate by Walker, 1926, p. 13) and Irene Volcanics in Canada. Becraft and Weis (1963, p. 173) correlated the conglomerate of the Huckleberry Formation with the Shedroof and Toby Conglomerates, and the greenstone with the Leola and Irene Volcanics.

The conglomerate northeast and southwest of the report area is fairly thick, and its near absence within the area suggests either nondeposition due to a topographic high or rapid thinning in a southeasterly direction. The basin in which the various conglomerate and volcanic units were deposited appears to have been fairly extensive in a northeast-southwest direction (Becraft and Weis, 1963, p. 16), but little is known about its extent, if any, to the northwest or southeast.

MONK FORMATION

The Monk Formation is poorly exposed in the report area, except on the mountain west of Cliff Ridge and for a distance of about 3 or 4 miles to the southwest (outside the quadrangle). The most distinguishing characteristic of the formation is probably its heterogeneity. Although predominantly slate and argillite, it contains numerous beds of dolomite, conglomerate, and quartzite.

The Monk Formation was named by Daly (1912, p. 148) for exposures of slate, phyllite, schist, and conglomerate along Monk Creek in southeastern British Columbia. He measured a section about 5,500 feet thick. Park and Cannon (1943, p. 11) estimated the thickness of the formation to be about 3,800 feet in the Metaline quadrangle. There, it is mostly fine-grained phyllite with numerous interbeds of carbonate rock, quartzite, and grit.

Rocks probably belonging to the Monk Formation are found in the hills bordering Bayley Creek, on the mountain west of Cliff Ridge, and on the hill 1 mile southwest of Chewelah. The patchy occurrence of this unit is due chiefly to faulting, but the variation in thickness also results from Precambrian erosion. The upper contact is everywhere an unconformity with the Addy Quartzite. Some of the rocks, especially those east and southeast of Bayley Creek, may not be part of the Monk Formation, but were so assigned because they are apparently underlain by the Huckleberry Formation.

On the mountain west of Cliff Ridge, dolomite of the Monk Formation rests on greenstone of the Huckleberry Formation. The dolomite is pale yellow to pale gray and highly recrystallized, and it occurs in beds as much as 5 feet thick. It is restricted to the lower 150 feet of the unit within the area and grades upward into a section that consists predominantly of slate, some of it carbonate bearing. About 1.5 miles to the southwest, along strike, the lower carbonate zone thickens and is overlain by rocks displaying an upward gradation similar to that found to the north. From there, for about 2 miles farther southwest, beds of conglomerate and carbonate-bearing slaty argillite separate the carbonate rock and the greenstone. Although poor exposures do not permit detailed stratigraphic observations, the conglomerate part of the unit appears to become progressively thicker to the southwest.

Park and Cannon (1943, p. 12) found a similar sequence of units near the base of the formation in the Metaline quadrangle, except that the thicknesses of the individual units differ from those in the report area. They reported that conglomerate forms the base of the Monk Formation north of the headwaters of Gypsy Creek and is overlain by 200–300 feet of limestone. South of the Gypsy Creek headwaters, however, the conglomerate is missing, and presumably the same limestone rests on greenstone of the Irene Volcanics, the Huckleberry greenstone equivalent. In and just west of the report area, clasts in the conglomerate are pebbles and cobbles of greenstone from the Huckleberry Formation, an andesitic-looking rock not found in the area, and fine-grained white quartzite. The matrix is metamorphosed but appears to have been carbonate-bearing argillite or siltstone that was sandy in places. The slate overlying the dolomite west of Cliff Ridge is the most common lithology in the formation. It is purple to maroon with thin pale-green beds which range in thickness from about one-sixteenth inch to about 2 feet. The purple beds are chiefly argillite, but the pale-green layers are composed of slightly coarser grained silt-sized material. Almost all the argillite is finely laminated, although in much of it the laminations are quite subtle. Both the purple and green layers are carbonate bearing. Higher in the section, the slate contains progressively less carbonate, and in the upper half of the unit almost no carbonate-bearing rocks are found.

The small fault-bounded outcrops of this unit 4.5 miles due north of Chewelah are made up of both the lower dolomite and the purple slate. At this locality, however, the transition zone contains several tens of feet of pisolitic sandy carbonate beds. The pisolitic rock is light yellow brown and forms beds as much as 2 feet thick. It grades upward bed-by-bed into slaty maroon argillite.

The Monk Formation west of Bayley Creek and on the hill 1 mile southwest of Chewelah is made up, in part, of lithologies not found in the more extensive section west of Cliff Ridge and could be part of the Stensgar Dolomite. West of Bayley Creek, the Addy Quartzite rests on about 100 feet of cream-colored dolomite. The dolomite is faulted over about 150 feet of purple to maroon argillitic dolomite and carbonate-bearing argillite, which includes a 20-foot-thick bed of conglomerate and pebbly mudstone near the top. This pebbly mudstone and the carbonate rocks below it may be part of either the Monk Formation or the Stensgar Dolomite.

The Monk Formation is overlain by the Lower Cambrian Addy Quartzite on the hill southwest of Chewelah. There the Monk consists of about 400 feet of medium-gray slaty siltite and argillite and contains beds of light-gray dolomite as much as 50 feet thick in the lower 150 feet. The slaty rock and dolomite overlie, or are faulted against, about 350 feet of highly sheared conglomerate provisionally assigned to the Huckleberry Formation. This rock strongly resembles the conglomerate member of the Huckleberry Formation in the magnesite belt, but is also similar to the conglomerate found in the Monk Formation southwest of Cliff Ridge. The conglomerate overlies the maroon argillite, carbonate-bearing pebbly mudstone, and dolomite believed to be part of the Stensgar Dolomite.

PALEOZOIC ROCKS

ADDY QUARTZITE

The Addy Quartzite, first described and named by Weaver (1920, p. 61–63), is a thick fine- to medium-grained vitreous quartzite. It crops out at many places in the west half of the report area and is one of the best stratigraphic markers in the region. This unit and its probable equivalent to the north, the Gypsy Quartzite, underlie much of northeastern Washington.

The best exposures of the Addy Quartzite in the report area are due east of Chewelah on the west flank of Quartzite Mountain, where about 1,500 feet of the unit forms spectacular, near-vertical cliffs (fig. 10). Eagle Mountain marks the northern limit of an approximately north-south discontinuous belt of Addy Quartzite which extends southward to just beyond Springdale. This belt is bounded on the west by a fault of variable dip and unknown displacement. The fault trace is covered by glacial and alluvial material for the entire length of the belt.

A surface of low relief was eroded across rocks of the Belt Supergroup after they had been faulted and broadly folded. The Addy sediments deposited on this surface therefore rest unconformably on several of the Precambrian units. At most places in the area, however, the angular discordance is so slight that it cannot be reliably measured. On the west flanks of Eagle Mountain and Quartzite Mountain, the Addy rests on member *c* of the Striped Peak Formation, and at various places south of Quartzite Mountain, it rests on members *a*, *c*, and *d* (figs. 9, 10). On the hill 1 mile southwest of Chewelah, and on the mountain 10 miles north of Chewelah, just west of the report area, the formation rests on the Monk Formation. Southwest of Valley it appears to have been deposited on the Buffalo Hump(?) Formation of the Deer Trail Group.

The unconformity at the base of the Addy Quartzite is well exposed on the west flank of Quartzite Mountain and on Jumpoff Joe Mountain. At both localities, the basal 100–300 feet is a distinctive purple quartzite. These beds are well stratified and generally are marked

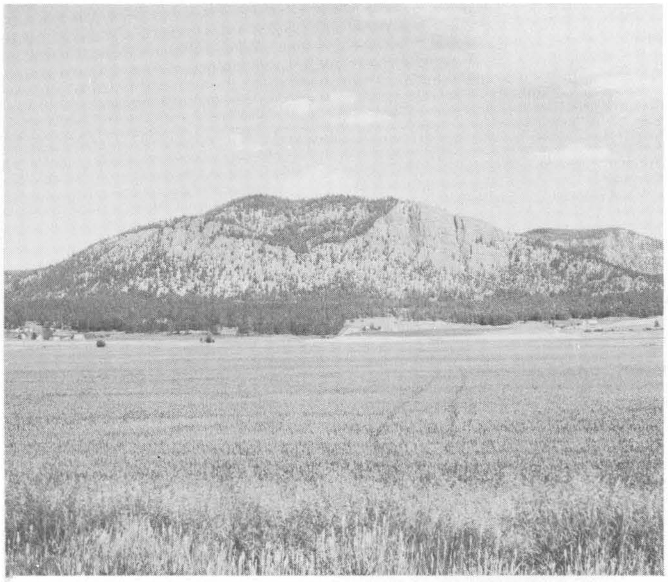


FIGURE 9.—The bold cliffs are dip slopes of Addy Quartzite which form the west side of Quartzite Mountain. Member *c* of the Striped Peak Formation underlies the quartzite just beyond the lip of the cliffs. The exposures formed by member *c* in the low swale near the center of the mountain and those behind the most prominent cliff on the south flank of the mountain are the best of any Belt formation in the area. The steep slopes on the hill in the right background are underlain by the Revett Formation. View east.

FIGURE 10.—Contact between the Addy Quartzite and member *d* of the Striped Peak Formation is at the base of the bold outcrops in the left center of the picture. The contact is well exposed here, but the angular discordance between the two units is so small that it cannot be measured. The base of member *d* is near the break in slope at the right edge of the photograph (arrow). Outcops are on the small hill immediately south of the Loon Lake copper mine. View north.

by a well-developed thin black striping which is easily mistaken for bedding. From the basal beds the Addy grades upward through pink quartzite over a stratigraphic distance of about 100–200 feet into the white and light-gray quartzite which characterizes most of the formation. At several places in the area, the purple quartzite is separated from the underlying Precambrian rocks by as much as 500 feet of white, gray, or pink quartzite and siltite (Reynolds, 1968, p. 11).

The bulk of the Addy Quartzite is white to light-gray vitreous quartzite, although some beds have a pale-yellow or pale-pink cast. Small pink quartz grains are sparsely scattered throughout this rock and are a characteristic of the formation. The beds range in thickness from a few inches to over 20 feet but average about 3 feet. Although a poorly developed bedding-plane parting is usually present and gives the rock a massively bedded appearance, bedding, as defined by differences in grain size or lithology, is not obvious owing to the homogeneity of the formation. Real bedding in the lower purple quartzite, as contrasted to the black striping, is defined by thin partings of argillaceous material or by beds of coarser grained quartz. Beds of purple quartzite average 1–2 feet in thickness, but massive beds as much as 5 feet thick are common.

Reynolds (1968, p. 13, 15) studied the lower part of the formation in detail, including the purple beds, and found that the rocks consist of fine to medium subrounded to well rounded moderately to well sorted quartz grains. These observations also apply to the rest of the formation. Small lenses of pebble conglomerate are common in the lower 200–400 feet and are abundantly exposed on Quartzite Mountain. Pebbles of quartzite are the most common clasts. Pods of conglomerate and sedimentary breccia as much as 5 feet thick occur locally at the contact with the underlying Precambrian rocks, on Deer Lake Mountain, and Quartzite Mountain. The clasts in these pods are composed of the immediately underlying material and are probably locally derived.

Small lensoidal pockets with a friable appearance are another feature characteristic of the Addy Quartzite as a whole but much more common in the lower 500 feet. Much of the cementing material in these pockets has apparently been leached out from between the grains, but the grains are held together strongly by a small amount of silica cement. Where numerous, the pockets impart a vuggy appearance to the rock.

At least 50 feet of shale, some of it sandy, occurs about 1,000 feet above the base of the unit on Quartzite Mountain. This shale is well stratified and light gray, pale green, and pale maroon. It is overlain by more light-colored quartzite identical with that below. The upper part of the formation is preserved only on the hill 1 mile south of Springdale. There, a few hundred feet of medium- to thick-bedded white quartzite is overlain by the Metaline Formation. An interval about 100 feet wide between argillaceous limestone of the lowest Metaline and quartzite of the highest Addy is covered, and so it is not known if there is any slate or argillite between the two units, as there is in sections north of the report area.

The most complete section of Addy Quartzite in the vicinity of the report area is just east of the town of Addy (fig. 2), about 4 miles northwest of Chewelah. There, a north-dipping, apparently homoclinal section of Addy Quartzite about 4,000 feet thick overlies Huckleberry greenstone and is reasonably well exposed. The upper part of the quartzite contact appears to be faulted against the Huckleberry Formation.

Only two fossil localities are known in the formation. One is located about 2,500 feet above the base of the section east of Addy. Fossils from this locality were identified by A. R. Palmer (written commun., 1964) and include *Nevadella addyensis* (Okulitch), *Kutorgina?*, and *Hyolithes* sp. Palmer concluded that the trilobite indicates that the quartzite is very Early Cambrian. The other locality is about 1 mile to the southwest, on the west side of Addy. Among those who have examined fossils from this second locality, Okulitch (1951, p. 405) seems to have made the most detailed study. He reported finding the following fauna: *Micromitra* (*Paterina*) sp., *Kutorgina* cf. *K. cingulata* (Billings), *Kutorgina* sp., *Rustella* cf. *R. edsoni* Walcott, *Hyolithellus* sp., *Nevadia* [=*Nevadella*] *addyensis* Okulitch, olenellid fragments, and fucoids. Okulitch concluded that "The fauna is undoubtedly Lower Cambrian in age; and the presence of the rare genus *Nevadia* [=Nevadella], with its very primitive characteristics, suggests the lower portion of the Lower Cambrian."

Lithologically, the fossil-bearing rock is not typical of the formation as a whole. It is thin- to medium-bedded fine-grained quartzite which is locally argillaceous and contains detrital muscovite. Beds range in thickness from 1 to 12 inches and are interbedded with argillite bands as much as 4 inches thick. The argillite is medium gray, black, and gray green. Some of it is laminated, but most is not. In the upper part of the fossiliferous zone are a few beds of dolomitic siltite. Above the dolomitic zone are more thick to massive quartzite beds typical of the formation. None of these atypical lithologies have been recognized within the report area.

The purple quartzite near the base of the Addy is one of the most important stratigraphic markers in the region, but appears to have been largely ignored by previous workers. Bennett (1941, p. 9) briefly described the formation, but did not mention the purple zone.

Campbell and Loofbourow mentioned a zone in which the bedding planes are emphasized by thin purple bands, but did not specify in what part of the section it occurs. On a brief trip the authors found the purple zone near the base of the formation on Stensgar Mountain, which is about in the center of the magnesite belt. Park and Cannon (1943) did not report a purple zone in the Metaline quadrangle, nor did Yates (1964) in the Deep Creek area.

Becraft and Weis (1963, p. 12) reported that the Addy Quartzite is about 3,900 feet thick in the Turtle Lake quadrangle, 30 miles southwest of Chewelah. They correlated the formation with the Gypsy Quartzite in the Metaline quadrangle. Forty miles northeast of Chewelah, in the Deep Creek area, about 2,900 feet of the Gypsy Quartzite conformably underlies the Maitlen Phyllite (Yates, 1964), but the section is faulted at the base. On Crowell-Sullivan Ridge near the type section of the Gypsy Quartzite in the Metaline quadrangle, the unit is 8,500 feet thick (Park and Cannon (1943, p. 13). About 7 miles west of that, however, the section is only about 5,300 feet thick.

CARBONATE ROCKS

Paleozoic carbonate rocks crop out discontinuously from Eagle Mountain to south of Springdale. Only the Cambrian, Devonian, and Mississippian Systems have been identified, but other systems may have gone unrecognized owing to the sparsity of fossils. Exposures in the area are inadequate to construct a complete composite section, and part of the section may not be exposed owing to faulting or erosion. Thick sections of Cambrian and Ordovician strata not found in the report area occur to the west in the Hunters quadrangle (Campbell and Raup, 1964) and to the north in the Colville area (Bennett, pl. 6, in Mills, 1962), Deep Creek area (Yates, 1964), and Metaline quadrangle (Park and Cannon, 1943, p. 17–22). This distribution may be due to rapid and large-scale facies changes; however, it is equally possible that these units were never deposited in the report area. Most of the Paleozoic rocks here may have been deposited at a distant location and brought in by large lateral movements.

Because of the discontinuous and patchy occurrence of the Paleozoic carbonate rocks that are preserved, the stratigraphic relations of these rocks are not well established. Much of the carbonate rock, especially between Loon Lake and Jumpoff Joe Lake, is exposed only in small, widely separated outcrops and had to be mapped as undivided Paleozoic.

METALINE FORMATION

About 1.5 miles southeast of Springdale, carbonate rocks of the Middle Cambrian Metaline Formation rest with apparent conformity on the Addy Quartzite. According to R. G. Yates, who examined these rocks with the authors in the field, the rocks closely resemble the lower and middle units of the Metaline Formation in the Deep Creek area, about 60 miles north of Springdale. There, the formation varies considerably from place to place owing to both facies changes and large lateral movements along faults; complete sections are more than 5,000 feet thick and consist chiefly of limestone and dolomite (Yates, 1964). In the northeastern part of the Hunters quadrangle, 15 miles northwest of Springdale, Campbell and Raup (1964) reported that as much as 8,000 feet of limestone and dolomite overlies the Addy Quartzite. They assigned these rocks to Weaver's (1920, p. 66) Old Dominion Limestone. Similarities in lithology and faunal assemblages, and relationship to the Addy Quartzite, make at least part of the Old Dominion a most likely correlative with the Metaline Formation in the area.

In the Deep Creek area and in the Metaline quadrangle, the Metaline Formation is separated from the Gypsy Quartzite (correlative with the Addy Quartzite) by the Maitlen Phyllite, which is over 5,000 feet thick in both areas. In the Hunters quadrangle and the report area, the Old Dominion Limestone and the Metaline Formation may rest directly on the Addy Quartzite. A covered area about 100 feet wide separates the highest outcrops of Addy Quartzite from the lowest Metaline carbonate on the hill southeast of Springdale, and conceivably that much phyllite could separate the two units. As no phyllite float was found in the covered area, the carbonate rock is assumed to rest directly on quartzite.

The absence of the Maitlen Phyllite in the southern part of Stevens County is apparently due to nondeposition rather than faulting. A major fault with large lateral transport separates sections containing the phyllite from those without it. The fault separating the Belt Supergroup and Deer Trail Group may also divide the Paleozoic section of the report area from that typified by the Paleozoic rocks in the Deep Creek area.

Southeast of Springdale the base of the Metaline Formation consists of gray to blue-gray limestone. Irregularly shaped yellow-brown-weathering argillaceous seams are distributed throughout the rock in an interwoven network. The surface of the weathered rock is typically grooved or fluted, resembling a rillenstein like surface. The rock is thick to thin bedded and generally fine grained. About 4,000 feet northeast of the Addy-Metaline contact, most of the lower part of the Metaline Formation is repeated by a fault. There, numerous beds of black shale as much as 1 foot thick, which are not exposed to the southwest, are interlayered with the argillaceous limestone. Above the argillaceous limestone thick- to thin-bedded tan to light-gray dolomite is interlayered with a small number of thin dark-gray shale layers.

The partial sections are very poorly exposed and may be repeated more than once by faulting. The upper contact is a fault, and so the thickness of the Metaline Formation in the area is not known. Because of poor exposures and uncertain structure, the thickness of even the preserved part of the formation cannot be accurately estimated. The part immediately above the Addy Quartzite and southwest of the fault that repeats the section may be homoclinal. If this portion of the section is not faulted, it is about 2,000–2,500 feet thick.

Trilobites and brachiopods were collected in the SW¼ sec. 35, T. 30 N., R. 40 E. A. R. Palmer examined two collections and identified the trilobites as *Peronopsis*, *Bathyuriscus*, *Olenoides*, and undetermined ptychopariods, and the brachiopods as *Linnarssonia* and *Acrothele*.

Palmer (written commun., 1968) stated that the collections contain undoubted Middle Cambrian fossils and also that "The * * * [fauna] * * * is most probably of late Middle Cambrian age, and could be the same age as the earlier faunas from the Metaline Limestone [Formation]. *Olenoides* is a long-ranging genus and precise placement * * * within the Middle Cambrian on its own merits is not possible."

DEVONIAN OR MISSISSIPPIAN CARBONATE ROCK

Some of the carbonate rocks in the Loon Lake quadrangle that Miller (1969, p. 3) described as Mississippian(?) may include strata as old as Devonian and are here referred to as Devonian or Mississippian. Limestone containing fossils overlies these rocks with no apparent unconformity. The lower age limit is less precisely based on Upper Devonian fossils found in isolated outcrops of carbonate rocks. The lower dolomite of the earlier report is divided into two units, and the overlying dolomite and calcareous slate are retained as a third.

UNIT 1

The lowest unit in this sequence is a dark-gray dolomite that crops out discontinuously from the south flank of Eagle Mountain to the mouth of Cottonwood Creek. It also crops out in a small area along the Burlington Northern Railroad tracks 1.5 miles northeast of Springdale. Although not very extensive, the best exposures of the unit are on the hill in sec. 19, T. 32 N., R. 41 E.

The dolomite is uniformly dark gray, except for a few zones which are a mottled light gray. No limestone beds were found in the unit, and in fact, very little of the rock is even slightly limy. In places the dolomite appears to be argillaceous, and it contains sparse sand grains throughout. Bedding is well defined, ranging from a few inches to 5 feet in thickness. The texture varies from aphanitic to coarse grained.

Sedimentary structures are found throughout the unit and along with the dark color serve to distinguish the dolomite from all other carbonate units. Oolites are common, usually confined to individual beds less than 1 foot thick but locally occurring as irregular patches. Dolomitic conglomerate is also found throughout the unit. It is composed of angular to slightly rounded clasts as much as 1 inch across, which are almost identical in appearance with the surrounding carbonate matrix. Randomly oriented curved chips of white coarse-grained dolomite scattered through most of the unit may be recrystallized fossils. Some resemble the curved cross section of a brachiopod shell, whereas others are cylindrical and may be pelmatozoan remains. The exposed portion of the unit is estimated to be between 600 and 700 feet thick.

The lower contact is covered by surficial deposits everywhere in the report area. Beneath the younger cover unit 1 must be faulted against either the Addy Quartzite or Metaline Formation. An exploratory oil well drilled by the Empire Exploration Co. in the NE¼ sec. 30, T. 31 N., R. 41 E., penetrated carbonate rock, most of it dolomite, for 2,192 feet. At 2,664 feet quartzite was penetrated immediately below a fault zone, although it is difficult to ascertain from the description of well cuttings if the drill was in unit 1 before hitting quartzite.

UNIT 2

The contact between units 1 and 2 is reasonably well exposed on the hill in sec. 19, T. 32 N., R. 41 E. Where examined, it is marked by a rather abrupt change from dark-gray to almost white dolomite. Like unit 1, unit 2 is internally homogeneous. The beds of white dolomite at the base are representative of the unit as a whole, except that large oolites or pisolites are found only in the lower 10–20 feet. Other than bedding and the oolites near the base, the unit contains no obvious sedimentary structures, perhaps because of recrystallization. Although the units bounding it are predominantly fine-grained or aphanitic carbonate rock, unit 2 is almost uniformly coarse grained.

Bedding is less well defined than in unit 1. Beds range from a few inches to about 5 feet in thickness; the average is about 3 feet. The thickness of the beds gives the rock a massive appearance, but close examination of weathered surfaces shows that most of the thicker beds are thinly laminated internally. A. K. Armstrong of the U.S. Geological Survey examined specimens collected from this unit and concluded that "Oolites and coated carbonate lithoclasts in a matrix of smaller pieces of the same material suggest a shoaling environment for part of the carbonate" (oral commun., 1967). These shoaling rocks alternate bed by bed with others containing algal structures suggestive of an intertidal environment (Miller, 1969, p. 3). No fossils other than the nondefinitive algae have been found in the unit.

The unit is 500–650 feet thick on the hill in sec. 19, T. 32 N., R. 41 E. The faulted section that is well exposed along the Burlington Northern Railroad tracks 2 miles northeast of Springdale is nearly horizontal and at least 550 feet thick. The rocks at this locality are very light gray rather than white. Because most of the units in this area are fault bounded, the rocks assigned to unit 2 here could conceivably be a different part of the section. They have been assigned to unit 2 because they contain oolites, especially near the base, and are underlain by dark-gray dolomite which resembles unit 1.

UNIT 3

Overlying the white dolomite is an extremely distinctive unit which, unlike some of the carbonate units, is recognized with confidence wherever mapped. It is not internally homogeneous like the other two units but is made up chiefly of light-colored dolomite, maroon slate, and all gradations between the two. The base of the unit is predominantly light-gray and cream-colored dolomite and rests with apparent conformity on the white dolomite of unit 2. About 40 or 50 feet above the base, several beds as much as 2 feet thick of pale-gray-green argillaceous dolomite and maroon slate are intercalated with the carbonate rock. The maroon beds become thicker and more numerous upward in the section. Two to three hundred feet above the base is a zone of thin-bedded maroon slate about 100 feet thick. The maroon rock grades into pale-green argillite both above and below the zone. These color changes are especially well displayed in the exposures northeast of Springdale. The slate is generally well laminated and contains thin seams of dolomite. It grades upward into light-gray and cream-colored dolomite similar to that in the lower part of the unit.

In the section northeast of Springdale, predominantly light-gray to white dolomite that strongly resembles the rocks of unit 2 is found 500–600 feet above the base of unit 3. The contact between this rock and the underlying cream-colored dolomite of unit 3 is not exposed but is thought to be a fault. If it is a fault, the preserved part of unit 3 is between 500 and 600 feet thick.

Bedding thickness in the dolomite ranges from a few inches to about 10 feet. Although the dolomite is cream colored (described as tan or gray tan by Miller (1969, p. 3)), the weathered color, which is almost pale orange, is more characteristic of the unit. The texture is partly saccharoidal, but some of the rock is sufficiently aphanitic to resemble chert.

MISSISSIPPIAN CARBONATE ROCK

Limestone and dolomitic limestone containing Mississippian fossils are found on the hill 1 mile north of Springdale and on a smaller hill 1.5 miles east of Valley. Most of the limestone is medium gray to blue gray. The lower part of the unit contains a few zones of tan to tan-gray dolomitic limestone, some of which are laminated. Bedding thickness ranges from less than 1 inch to more than 15 feet but averages about 2 or 3 feet. Chert, in beds and nodules, is common but confined to definite zones. On the small hill near the center of sec. 27, T. 30 N., R. 40 E., about 400 feet of limestone in which cherty rock alternates with noncherty is well exposed. The cherty zones are about 80 feet thick, and the zones without chert about half that. Most of the chert beds are only a few inches thick, but some are as much as 10 inches. Slightly higher in the section to the northeast, some of the thicker beds appear to have been bioclastic limestone that has been locally replaced by chert.

The grain size of the limestone on the hill north of Springdale varies from coarse to fine and to some degree may coarsen in the direction of the coarse-grained quartz monzonite. Where dolomitic, the rock tends to be slightly coarser than the limestone and ranges from medium to coarse grained. Approximately 500 feet above the base of the preserved section there is about 10 feet of dark-gray to black aphanitic limestone which serves locally as a marker within the unit. Both the very fine grain size and the color of this zone distinguish it from the other rocks in the unit.

Neither the upper nor the lower contact is exposed in the area. Northeast of Springdale the upper contact is a fault, and the lower part is concealed beneath surficial material. A thickness of 600–700 feet for the preserved part of the unit is calculated from outcrop width, but this figure may be in error because of undetected faults.

Fossils are found throughout the unit but are abundant in only a few places. The tan to tan-gray beds in the lower part of the unit contain abundant fenestellid bryozoans and a few solitary corals. The medium-gray to blue-gray beds in the same general part of the section contain abundant pelmatozoan debris and a few fenestellid bryozoans, but all are partly recrystallized. Near the middle part of the exposed section northeast of Springdale is a dolomitic limestone bed which contains abundant corals, brachiopods, gastropods, and pelmatozoan debris. A. K. Armstrong of the U.S. Geological Survey identified the following fossils (written commun., 1967):

Coral:
Amplexizaphrentis sp.

Brachiopods:
Unispirifer sp. indet.
Spirifer sp. indet.
Pseudosyrinx? sp. indet.
Gastropod:
Platyceras sp.

Armstrong reported the following:

> The fossils are poorly preserved as fragments which have been replaced by coarse chalcedony within recrystallized dolomitic limestone. Acid etching has yielded a small fauna. The corals are typical representatives of Mississippian *Amplexizaphrentis*. A number of fragments of a brachiopod with a high cardinal area, non-plicated sinus and apparently devoid of a syrinx may possibly belong to the genus *Pseudosyrinx*. The original shell structure was not preserved and the fragmentary nature of the material makes positive generic identification impossible. Fragments of pedicle valves strongly indicate the presence of both the genus *Spirifer* and *Unisperifer*. The fauna suggests a Mississippian age.

The limestone on the hill 1.5 miles east of Valley presumably belongs to the same unit. A. J. Boucot (written commun., 1966) recovered the following brachiopods and suggested that the rocks were probably Mississippian in age:

Rhipidomella sp.
Composita sp.
indet. spirifer (fine ribs)
indet. spirifer (coarse ribs)
Crurithyris sp.
indet. brachiopods

Gilbert Klapper (written commun., 1966) examined rocks from the same locality for conodonts. Samples of a light-tan dolomitic limestone bed were collected about 30 feet stratigraphically below the gray limestone collected by Boucot and yielded the following:

Siphondella isosticha (Copper)
Pseudopolygnathus multistriata Mehl & Thomas
Gnathodus sp.

Klapper concluded that "The * * * [age of the rocks] * * * is definitely Mississippian. It contains about the same fauna as that known from the basal Banff Formation of Crowsnest Pass, Alberta * * *. The fauna is one zone higher than the fauna described from the basal Lodgepole * * *". (See Klapper, 1966, p. 5.)

Embysk (1954, p. 14) identified the following ostracodes from the same locality:

Graphiodactylus tenuis
Jonesina craterigera
Cavellina aff. *C. corelli*

Except for these sparse outcrops in the report area, no Mississippian rocks have been reported in northeastern Washington or northern Idaho. The significance of this fact cannot be competently evaluated by work in and around the report area, however, because the entire Paleozoic is so incompletely exposed and poorly understood.

PALEOZOIC CARBONATE ROCKS, UNDIVIDED

Much of the Paleozoic carbonate rock is mapped as undivided because of the lack of distinguishing characteristics needed to either assign the rock to established units or define a new unit. In some of the areas shown as bedrock on the geologic map, the exposure consists of only a few widely separated small outcrops. Some of the undivided carbonate rock undoubtedly belongs to units already described, but much of it definitely does not. Although lithologic descriptions of many individual outcrop areas would serve little purpose, some localities should be mentioned because the rocks there contain a few fossils.

Jones (1929, p. 43) reported finding brachiopods in a 2-foot-thick bed of purplish-red limestone in the SE cor. sec. 22, T. 33 N., R. 40 E. These were identified by Branson (1931, p. 70) as *Kutorgina cingulata* (Billings) of Early Cambrian age. The limestone is enclosed in light- to dark-gray dolomite, cream colored in places. Most of the carbonate in this small area of outcrop is highly brecciated dolomite—so brecciated that bedding is difficult to recognize.

These fossil beds were searched for without success. Exposures are good only locally, and stratigraphic relations are not clear in this area. *Kutorgina cingulata* (Billings) is found in the upper part of the Addy Quartzite, and Jones and Barnes may have collected the fossils from a fault slice of that unit. The lithologic descriptions by Jones and Branson are not detailed enough to test this hypothesis, although Branson (1931, p. 70) mentioned that "the limestone is reddish in color, dense, sandy, and bears thin lenses of impure argillite." Some of the fossiliferous beds in the Addy Quartzite east of Addy would fit this description.

The rocks from which the fossils came may also be part of the Old Dominion Limestone of Weaver (1920), which overlies the Addy Quartzite north of Colville (Bennett, in Mills, 1962, pl. 6). Although quite fossiliferous, the limestone there contains only *Archaeocyatha*, however.

R. H. B. Jones and J. P. Thompson collected Late Devonian fossils from a rather nondistinctive medium-gray limestone in the north-central part of sec. 19, T. 31 N., R. 41 E. The rock appears to be interbedded with light-gray and cream-colored dolomite, but exposures are not good enough to determine even local stratigraphic relationships. The limestone is punky in places owing to the presence of intricate solution cavities. In addition to the brachiopods, the rock contains solitary corals, which are poorly preserved, and pelmatozoan debris.

J. T. Dutro and Helen Duncan of the U.S. Geological Survey examined these fossils and identified the brachiopod *Cyrtospirifer* sp., which, they remarked, "is widely distributed in rocks of Late Devonian age in the United States and elsewhere" (written commun., 1958). Another collection was made from a locality a few hundred yards farther north by R. G. Yates, J. C. Moore, and F. K. Miller. The collection was examined by C. W. Merriam of the U.S. Geological Survey, who stated that the collection "probably is Upper Devonian as suggested by *Tenticospirifer* which resembles *Tenticospirifer utahensis* of the Devils Gate Limestone in Nevada and other species of this genus in the Late Devonian of Iowa. This fauna may be of about the same age as those from the Devonian examined by Dutro and Duncan" (written commun., 1965).

In 1927, R. H. B. Jones and J. P. Thomson collected fossils from a poorly exposed fine-grained medium- to light-gray dolomitic limestone in the NW¼ sec. 23, T. 30 N., R. 40 E. These fossils were identified in 1958 by J. T. Dutro and Helen Duncan as follows:

Large crinoid columnals, indet.
Fenestella sp.
Cystodictya sp.
Leptargonia cf. *L. analoga* (Phillips)
productoid brachiopod, genus indet.
Tetracamera? sp.
camarotoechid brachiopod, indet.
spiriferoid brachiopod, indet.
Syringothyris? sp.

Dutro and Duncan concluded that "The association of *Leptargonia* cf. *L. analoga* (Phillips), *Tetracamera*? sp., and *Syringothyris*? sp., together with productoid and spiriferoid brachiopods, indicates an Early Mississippian age. A reservation is placed on this assignment only because the poor preservation of the fossils make positive identifications difficult" (written commun., 1958). The unit from which these fossils were collected may be the same as the Mississippian limestone, but lithologic differences are too great to map it as that.

Embysk (1954, p. 15) collected the following fossils from limestone found in secs. 23, 26, and 27, T. 30 N., R. 40 E.:

Rhabdammina sp.
Ammobaculites? sp.
Endothyra sp.
Millerella sp.
Globovalvulina cf. *G. bulloides*
Trochammina sp.
Lophophyllidium cf. *L. proliferum*
Rhombopora nitidula
Spirifer aff. *S. rockymontanus*
? *Squamularia transversa*
Small *Composita*
Amphissites cf. *A. centronotus*
Amphissites cf. *A. simplicissimus*

She reported that the fauna is Pennsylvanian in age, but she did not indicate exactly where in those three sections the fossils were collected. Despite a concerted effort to recover the locality, no Pennsylvanian fossils were found. Unfortunately, four of the Paleozoic carbonate units in addition to the carbonate rocks mapped as undivided are found in sections 23, 26, and 27.

MESOZOIC PLUTONIC ROCKS

Plutonic rock underlies a total of about 150 square miles at the north and south ends of the report area and intrude the complexly faulted and folded Precambrian and Paleozoic rocks. Eight distinct plutonic bodies, ranging in composition from granodiorite to alkali-rich quartz monzonite, have been mapped. These plutons are part of the Kaniksu and Colville–Loon Lake batholiths, following the terminology of Yates, Becraft, Campbell, and Pearson (1966, p. 56).

Geochronologic work by Joan C. Engels shows that the plutons represent at least three periods of intrusive activity, although they are not easily grouped on the basis of mineralogical or physical characteristics. The plutons range in area from about 1 square mile to more than 60 square miles. Most are irregular in shape, and the configuration of only a few appears to have been controlled by preexisting structures.

Some plutons are not named because they are too small to be of regional importance or are obviously an early- or late-stage variant of a larger pluton. Four of the plutons considered to be of regional importance have been named.

The rock classification used is shown in figure 11. Approximately 150 specimens were stained using the method developed by Laniz, Stevens, and Norman (1964) and were modally analyzed. The number of specimens stained from any one pluton is a function of the size of the pluton and the variation of mineral composition within the pluton. The greatest number of modes were determined for the larger plutons and those that showed noticeable variations in mineral composition. In most specimens for which accuracy was checked, a total of 1,000–1,500 points insured no more than a 2 percent error in any constituent. Point counts were made with a glass plate imprinted with a grid of points spaced 0.06 inches apart. The plate was placed upon a stained slab, and the number of points falling on each mineral was counted under a binocular microscope.

Modes of some fine-grained rocks were obtained using stained thin sections and methods described by Chayes (1956). Almost all thin sections used for petrographic descriptions were stained for potassium feldspar. All

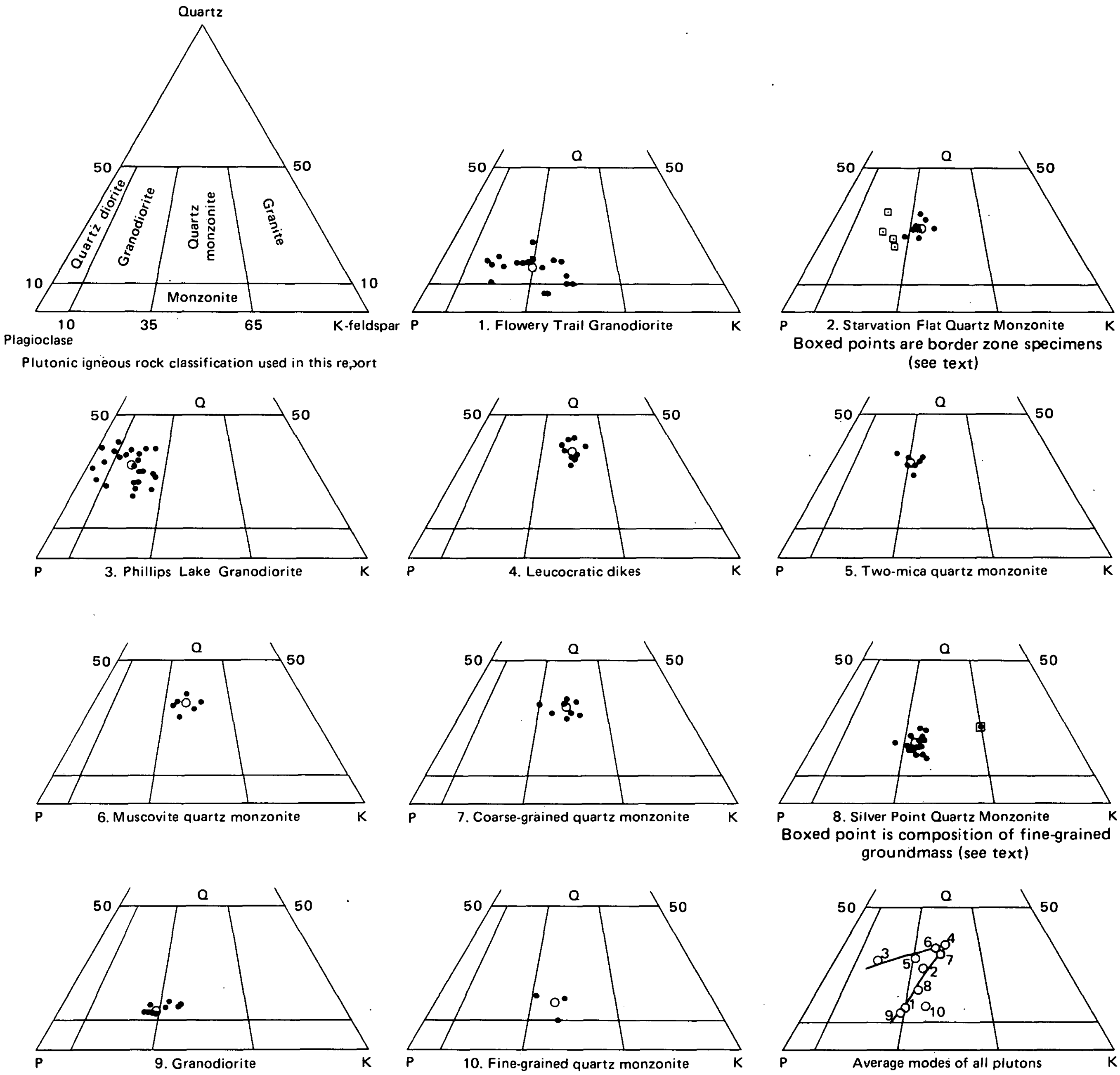


FIGURE 11.—Modes of plutonic rocks in the Chewelah-Loon Lake area. Circle is the average for each pluton. Modes recalculated to 100 percent quartz, plagioclase, and potassium feldspar.

mineral identifications are from hand specimen or thin section; no X-ray work has been done. Plagioclase compositions were determined by measuring refractive indices using oil-immersion methods. The refractive indices of amphiboles were measured in the same manner. Although a small amount of overlap of indices was noted, each pluton contains a hornblende of different refractive indices and presumably different composition.

Table 2 summarizes the mineralogy and texture of the major plutons in the report area.

FLOWERY TRAIL GRANODIORITE

LOCATION, EXTENT, AND TOPOGRAPHIC EXPRESSION

The Flowery Trail Granodiorite, named by Clark and Miller (1968, p. 3), is an elongate pluton about 10 square miles in area, one of the few entirely within the report area. The west end lies about 1 mile east of Chewelah, and the long axis extends east-northeast from there for about 7.5 miles. The width nowhere exceeds 2 miles. About 5 square miles is covered by

TABLE 2.—*Petrographic data on plutonic rocks*

Name of pluton	Extent within report area (sq mi)	Texture	Average mode	Plagioclase	Potassium feldspar	Hornblende	Biotite	Hornblende: biotite	Pyroxene	Accessories
Flower Trail Granodiorite.	10	Medium to coarse grained, hypidiomorphic-granular.	Plagioclase, 38; potassium feldspar, 20; quartz, 11; mafics, 31.	Zoned from An_{22} to An_{35}; some albite rims.	Microcline and orthoclase(?).	X=light tan, Y=dark olive green, Z=bluish green, ny=1.686, Z∧C=14°.	X=light tan, Y=Z=dark olive green; amount is highly variable.	≈100–≈2	As cores in hornblende only; $2V$=60°.	Epidote, sphene, apatite, zircon, magnetite, ilmenite(?), sericite, clinozoisite, calcite, tourmaline, pyrite.
Starvation Flat Quartz Monzonite.	35	Coarse grained, hypidiomorphic-granular.	Plagioclase, 37; potassium feldspar, 24; quartz, 26; mafics, 13.	Zoned from An_{20} to An_{35}.	Orthoclase(?) (microperthite); microcline.	X=light tan, Y=olive green, Z=light green, ny=1.670, Z∧C=17°–25°, $2V$=66°–69°.	X=tan, Y=Z=brown; euhedral.	(avg) .7	As cores in hornblende near contact with greenstone of Precambrian Huckleberry Formation.	Apatite, sphene, magnetite, zircon, allanite, epidote, tourmaline, muscovite, chlorite, calcite, hematite.
Phillips Lake Granodiorite.	60	Coarse grained. Very slightly foliated. Micas are interstitial to, and wrapped around feldspar and quartz.	Plagioclase, 46; potassium feldspar, 11; quartz, 28; mafics, 15. (includes muscovite).	Avg composition An_{22}; zoned from An_{16} to An_{37}.	Microcline	Absent	X=grayish tan, Y=Z=blackish brown; muscovite:biotite =0.3 avg.		Absent	Apatite, zircon, allanite, epidote, magnetite, garnet, tourmaline.
Leucocratic dikes (associated with the Phillips Lake Granodiorite).	See text section "Phillips Lake Granodiorite (and associated dikes)."	Medium to fine grained, hypidiomorphic- to xenomorphic-granular.	Plagioclase, 29; potassium feldspar, 28; quartz, 33; muscovite, 10. (includes muscovite).	Avg composition An_{7}.	Microcline	Absent	X=tan, Y=Z=blackish brown; muscovite:biotite =2.1 avg.		Absent	Zircon, apatite, garnet.
Two-mica quartz monzonite.	2	Coarse grained, hypidiomorphic-granular.	Plagioclase, 38; potassium feldspar 21; quartz, 30; mafics, 11. (includes muscovite).	Avg composition An_{18}.	Microcline	Absent	X=golden tan, Y=Z=reddish brown; muscovite:biotite =0.3 avg.		Absent	Zircon, apatite, opaque minerals.
Coarse-grained quartz monzonite.	15	Very coarse grained, porphyritic	Plagioclase, 33; potassium feldspar, 29; quartz, 31; biotite, 7.	Avg composition An_{20}.	Perthite and microperthite; also occurs as thin rims bordering quartz and plagioclase.	Absent	X=light tan, Y=Z=brown; much is chloritized.		Absent	Sphene, apatite, zircon, magnetite.
Muscovite quartz monzonite.	2	Coarse grained, hypidiomorphic-granular. Locally has texture that may be protoclastic.	Plagioclase, 34; potassium feldspar, 33; quartz, 27; muscovite, 6.	Avg composition An_{3}.	Microcline	Absent	Absent		Absent	Garnet, magnetite, apatite, zircon, limonite.
Silver Point Quartz Monzonite.	30	Medium to coarse grained, porphyritic.	Plagioclase, 40; potassium feldspar, 25; quartz, 18; mafics, 17.	Avg composition An_{20}; zoned from An_{20} to An_{35}.	Perthite and microperthite.	X=light tan, Y=olive green, Z=medium green, ny=1.662, Z∧C=17°.	X=tan, Y=Z=dark brown.	.9	Absent	Sphene, apatite, magnetite, zircon, allanite, epidote, muscovite, chlorite.
Granodiorite.	¾–1	Coarse grained, hypidiomorphic-granular.	Plagioclase, 40; potassium feldspar, 22; quartz, 10; mafics, 28.	Zoned from An_{20} to An_{35}.	Orthoclase(?)	X=light tan, Y=olive green, Z=medium green, ny=1.663, Z∧C=13°.	X=tan, Y=Z=dark brown.	1.1	As cores in hornblende only.	Sphene, apatite, zircon, magnetite.
Fine-grained quartz monzonite.	1½	Medium to fine grained, hypidiomorphic-granular.	Plagioclase, 38; potassium feldspar, 32; quartz, 12; mafics, 18.	Avg composition An_{20}.	Orthoclase(?)	X=pale tan, Y=pale olive green, Z=pale green, ny=1.651, Z∧C=22°.	X=tan, Y=Z=dark brown.	1.0	As cores in hornblende only.	Sphene, apatite, zircon, magnetite, allanite.

glacial and alluvial material. The Flowery Trail road traverses the length of the pluton and in the eastern part provides excellent roadcut exposures.

A major canyon is incised along the length of the pluton and owes its presence, at least in part, to the nonresistant nature of the rock. Natural outcrops, where present, are usually quite weathered. The best exposures are found where the granodiorite is in contact with more resistant metamorphic rocks.

INTERNAL FEATURES

The Flowery Trail Granodiorite is an even-grained hornblende-biotite granodiorite. Although relatively small, the pluton shows the largest compositional variation of any in the report area. Modal analyses indicate compositions ranging from granodiorite, through quartz monzonite, to monzonite.

In contrast to normal zonation in a plutonic body, the most quartz-rich specimens are found around the margins of this pluton. The quartz-rich rock at the margins, however, contains about 15 volume percent more mafic minerals than the rock in the center. This variation in mafic minerals may reflect primary mineralogical zoning, but the high quartz content of the border rock is probably due to assimilation of the host rocks—quartz-rich metasedimentary rocks of the Belt Supergroup. Some modes of rocks from the interior of the pluton do not indicate any zonal distribution of minerals.

If the zoning suggested by most of the modes is real, it is too complex to be outlined by the few samples whose modes have been determined. Unfortunately, incomplete exposure due to heavy forest cover and glacial deposits precludes sampling at the density needed to reliably determine whether the zonation is real or apparent.

CONTACT RELATIONS

Most of the contact is concealed by glacial and alluvial material or heavy forest cover. Even where the granodiorite is shown on the map, outcrops are somewhat patchy. The best exposures of the contact are in the west half of the pluton in the vicinity of the Jay Gould mine, the Juno Echo mine, and the Blue Star mine. At these three localities, which are on both sides of the pluton, the contact with the metamorphic rocks appears to be steeply dipping and is very irregular on a small scale. Abrupt changes in direction every 10–100 feet are complicated by abundant, irregularly oriented granodiorite dikes and sills which intrude the host rock.

The metamorphic aureole surrounding the Flowery Trail Granodiorite varies in width. Along most of the south border of the pluton, noticeable effects apparently do not extend more than half a mile into the host rocks. One of the outermost areas where discernible effects are found is on the northeast flank of Quartzite Mountain, about 1 mile south of the pluton. There, carbonate-bearing rocks of the upper Wallace Formation have been recrystallized to calc-silicate hornfels of the low albite-epidote-hornfels facies of Turner and Verhoogen (1960, p. 511). North of the pluton, the only place where the metamorphic effects can be distinguished from those of the Phillips Lake Granodiorite is on the south flank of Eagle Mountain. Highly recrystallized rocks near the contact of the Flowery Trail Granodiorite can be traced northward into relatively unmetamorphosed rocks near the summit of Eagle Mountain. Low on the north flank, the metamorphic recrystallization again becomes apparent and increases northward toward the Phillips Lake body.

On the north side of the pluton, mineral assemblages characteristic of the hornblende-hornfels facies of Turner and Verhoogen (1960, p. 515) extend to within a few hundred feet of the contact. Assemblages include diopside-quartz-calcite-microcline and quartz-muscovite-biotite-plagioclase-andalusite. More than a few hundred feet from the contact, host rocks at most places are completely recrystallized, often with the formation of large porphyroblasts, but with the formation of minerals characteristic of the albite-epidote hornfels facies. Common assemblages include quartz-albite-muscovite-chlorite-clinozosite-green biotite, quartz-muscovite-tourmaline, and quartz-albite-muscovite-actinolite-clinozoisite.

About half a mile east of the pluton, in the SW cor. sec. 32, T. 33 N., R. 42 E., kyanite-bearing schist is found in the upper part of the Prichard Formation. Although this locality is not far from the Phillips Lake Granodiorite, the kyanite probably crystallized owing to the intrusion of the Flowery Trail Granodiorite. The latter is considerably closer, and when it was intruded the amount of cover was probably thicker than when the Phillips Lake body was intruded. Abundant andalusite schist is found on the east and northeast flanks of Goddards Peak in rocks of about the same chemical composition as those forming the kyanite schist, but these rocks are closer to the Phillips Lake Granodiorite and were probably crystallized by that pluton.

PETROLOGY

Most of the Flowery Trail Granodiorite is medium to fine grained. Locally the border rock is slightly foliate, but most of it is structureless. Inclusions are common though not abundant. Mafic minerals are abnormally rich in irregular patches as much as several hundred square feet, and these contrast strikingly with the more leucocratic surrounding rock. If alpine exposures were available, the rock would appear mottled on a large scale, with the light and dark areas blending over a distance of a few feet or less. Its high color index distinguishes the Flowery Trail Granodiorite from all other plutonic rocks in the area except for a granodiorite south of Springdale.

Plagioclase, the most abundant constituent of this rock, occurs as subhedral to euhedral crystals that are both normally and reversely zoned. Most of it is oligoclase; some rims are calcic albite, and some cores intermediate andesine. In most sections the calcic cores are obvious owing to large concentrations of small euhedral clinozoisite crystals, calcite, and muscovite or sericite. These minerals, especially the clinozoisite and calcite, indicate that the plagioclase cores were probably more calcic before alteration and recrystallization.

The potassium feldspar is microcline, although some untwinned crystals may be orthoclase. All crystals are anhedral and appear to have filled spaces between other minerals. Where alteration of the granodiorite is obvious, especially near the Jay Gould mine, only the potassium feldspar has been notably affected. There, the mineral appears to have been preferentially removed from the rock during mineralization. This relationship is found only in the alteration zones around noticeably mineralized areas.

Clear anhedral quartz is present as small interstitial grains and less commonly as rims around plagioclase. Some grains contain hairlike crystals of what may be rutile, and some have undulatory extinction. Most grains are 0.04 inch or less across and are difficult to see in hand specimen. Because of this and the high mafic mineral content, the rock is easily mistaken for a diorite in the field.

Hornblende, the most abundant mafic mineral, is euhedral to subhedral, averages about 0.08 inch in length, and is easily identified in all hand specimens. Some crystals are corroded and embayed by biotite, both around the edges and in the interior. Rounded cores of monoclinic pyroxene ($2V{\sim}60°$) occupy the centers of a few hornblende crystals. This is the only known occurrence of pyroxene in the Flowery Trail Granodiorite.

Biotite is present in all thin sections, but is always less than half as abundant as hornblende. Crystals are usually subhedral and closely associated with the hornblende.

Sphene is by far the most abundant accessory mineral and makes up as much as half a percent of some specimens. Crystals range in size from 0.004 to about 0.12 inch. The characteristic wedge-shaped crystals are visible in all hand specimens. Apatite is the only other abundant accessory mineral in the Flowery Trail Granodiorite, although it is not as abundant as sphene. Zircon, magnetite, and ilmenite(?) are ubiquitous but not abundant.

Minerals of the epidote group are the most common alteration products in the granodiorite. Both epidote and clinozoisite are present; highly birefringent epidote ($\sim$0.35) is the most common. Epidote occurs as clusters of small anhedral to euhedral crystals around hornblende and biotite and within plagioclase crystals. Some are slightly pleochroic: Z′=very light green, X′=colorless. Pyrite, sericite or muscovite, tourmaline, and carbonate minerals are the only other secondary minerals in the granodiorite, and all but the sericite or muscovite are rare.

Changes in the characteristics of some minerals, evident only under the microscope, suggest a mild metamorphism that increases in an easterly direction. Highly discordant apparent potassium-argon ages on hornblende-biotite pairs support this conclusion. (See section "Potassium-Argon Ages of the Plutonic Rocks.") In thin section, plagioclase crystals are free of the cloudy haze of late-stage alteration products almost always found in normal plutonic plagioclase. In their place, scattered internally through the plagioclase, are anomalous-looking small euhedral crystals of clinozoisite, epidote, and muscovite which were probably formed by recrystallization of this late-stage alteration haze. Although metamorphism did not proceed far enough to erase zoning in the crystals, in plane-polarized light they look like the clear unaltered plagioclase found in metamorphic rocks. These relations are most noticeable at the east end of the pluton, suggesting the source of the metamorphism was in that direction.

Pleochroism and refractive indices of hornblende in rock from the eastern part of the pluton are markedly different from those in rock from the western part:

	X	Y	Z	n_y
West	Tan	Olive green	Blue green	1.679
East	Pale tan	Olive green	Green	1.691

Eskola (1952, p. 166), Shido (1958, p. 171), and Shido and Myashiro (1959, p. 86) reported a change in pleochroism of hornblende from blue green to green with advancing metamorphism. The changes they reported were in regionally metamorphosed rocks, but the comparison may be valid.

The metamorphism could have been caused by any of several younger bodies to the east. The younger Phillips Lake Granodiorite and leucocratic dikes lie to the north and east but may not be the sole cause of metamorphism, as the former shows some of the same metamorphic features found in the Flowery Trail Granodiorite. Younger intrusives are found about 10 miles southeast of the Flowery Trail Granodiorite. Although these intrusive bodies are probably not close enough to have affected the Flowery Trail, discordances in potassium-argon dates suggest that other younger plutons may be buried at shallow depth just east of the report area.

EMPLACEMENT

The location and shape of the Flowery Trail Granodiorite appear to have been controlled by two or more east-northeast-trending high-angle faults. Rocks and preexisting structures which strike towards the pluton are bent parallel to it near the contact and suggest that the emplacement was, at least in part, forceful. Abundant dikes and xenoliths and apparent contamination by country rock near the contact, however, hint that stoping and assimilation were at least as important as forceful intrusion.

STARVATION FLAT QUARTZ MONZONITE

LOCATION, EXTENT, AND TOPOGRAPHIC EXPRESSION

Starvation Flat, located in the northwest corner of the report area, is underlain almost entirely by a texturally and mineralogically uniform quartz monzonite. The rock was named by Clark and Miller (1968, p. 3) in the preliminary report on the Chewelah Mountain quadrangle, the north half of the report area. The pluton occupies about 20–25 square miles within an area extending on the south from the abandoned Cliff Ridge Lookout to the north and west boundaries of the quadrangle. Glacial debris covers much of the northwest corner of the report area. The east side of the pluton terminates at a high-angle fault of probably large displacement, which places it against the Phillips Lake Granodiorite. The quartz monzonite is found in many of the roadcuts along State Highway 294 north and west of the quadrangle and thus extends well outside the report area. Seventeen miles northeast of Calispell Peak, quartz monzonite that is texturally, mineralogically, and modally identical with the Starvation Flat Quartz Monzonite crops out in large roadcuts along State Highway 31, just north of Lost Creek.

The pluton forms a topographic low in the general landscape and attains significant relief only near contacts with the more resistant metamorphic rocks. The rather subdued topography around Starvation Flat is typical of that underlain by the Starvation Flat Quartz Monzonite. Although not mapped in detail, at least part of the contact between the quartz monzonite and the metamorphic rocks to the southwest can be inferred from topographic differences.

CONTACT RELATIONS

On Blacktail Mountain, near the north border of the area, intrusive relationships between the Starvation Flat Quartz Monzonite, the Phillips Lake Granodiorite, and highly metamorphosed sedimentary rocks are obscured because leucocratic quartz monzonite dikes and sills have intruded all three rock types. The configuration of the contact between the Starvation Flat Quartz Monzonite and the metamorphic rocks suggests a near-vertical relation. At localities far distanct from the quartz monzonite, the leucocratic dikes are closely associated with the granodiorite and genetically related to it. Since the quartz monzonite is intruded by the dikes, it must be older than the granodiorite and presumably is intruded by it.

South from Blacktail Mountain the inferred contact between the Starvation Flat Quartz Monzonite and the Phillips Lake Granodiorite is buried under glacial debris, but appears to be essentially straight. A contact this long and straight could be a fault, or it could be controlled by a preexisting fault. One well-located fault and two inferred faults are alined with the contact from the south and project toward it. Mylonite and cataclasite are developed for about 3 miles to the northeast and southwest along the contact from the point where it intersects Little Bear Creek.

Along the south border of the pluton, the quartz monzonite intrudes the Huckleberry Formation. However, the contact is almost entirely covered by glacial and alluvial sedmients, and its attitude is not known. Just outside the west boundary of the area, the contact swings south and roughly parallels the boundary of the area for 2 miles. Here the contact is fairly well exposed and quite irregular in configuration. The outcrop pattern and narrow width of the metamorphic aureole suggest that the contact is generally steeply dipping.

Pelitic, basic, and carbonate-bearing quartz-feldspathic rocks of the Monk Formation underlie the W½ sec. 27, T. 34 N., R. 40 E., at the west margin of the map area. About 400 feet from the quartz monzonite contact, the carbonate-bearing rocks are recrystallized to plagioclase-diopside-grossularite-quartz hornfels, and the basic rocks to plagioclase-diopside-hypersthene-quartz hornfels characterizing the pyroxene hornfels facies of Turner and Verhoogen (1960); in the interval from approximately 400 to 1,500 feet, actinolite replaces the hypersthene resulting in an assemblage characteristic of the hornblende hornfels facies. Beyond about 1,500 feet metamorphic effects diminish abruptly.

In contrast to the Monk Formation, noticeable recrystallization effects in the Huckleberry Formation extend only a few hundred feet from the quartz monzonite. At greater distances changes in the greenstone are microscopic. The metamorphic effects of the Starvation Flat Quartz Monzonite and the Phillips Lake Granodiorite on the host rocks between them cannot be differentiated by source.

PETROLOGY

The Starvation Flat pluton consists of a medium- to coarse-grained hypidiomorphic-granular hornblende-biotite quartz monzonite. The rock is extremely uniform in both mineralogy and texture. Modal analyses show that the composition is also uniform, except along the southwest border where the rock is contaminated apparently by assimilation of Huckleberry greenstone or by reaction with it. The average modal composition of the noncontaminated rock plots well within the quartz monzonite field, but toward the granodiorite side. (See fig. 11.)

In hand specimen the quartz monzonite is light gray and commonly speckled with pink feldspar. White plagioclase, pink or white potassium feldspar, clear or smoky quartz, shiny black biotite, greenish-black hornblende, and golden-brown crystals of sphene are easily seen without a hand lens. A distinguishing feature of this rock is the ubiquitous occurrence of biotite in euhedral pseudohexagonal tablets.

Plagioclase is the most abundant mineral in the rock. Crystals are subhedral to euhedral and average about 0.15 inch in length, although some reach 0.4 inch. Average composition is An_{25}. The most calcic plagioclase core measured is An_{35}, and the most sodic rim An_{20}. Some crystals appear to be unzoned, whereas others are highly complex and show both normal and reverse zoning.

Most of the potassium feldspar appears to be orthoclase. Perthitic intergrowths are common but extremely fine and difficult to see even under very high magnification. Carlsbad twinning is present but rare. Crystals are anhedral and occur interstitially to all other minerals of comparable grain size except quartz. Average size is about 0.2 inch; some grains are as large as 0.6 inch. No compositional zoning was noted, but small crystals of plagioclase, quartz, hornblende, biotite, and apatite are oriented parallel to crystallographic directions in some crystals.

Quartz also occurs as anhedral grains, filling interstices between plagioclase and the mafic minerals. Crystals average about 0.14 inch in size. Undulatory extinction is present in all grains but is not extreme. The quartz is clear in thin section and contains only a few almost submicroscopic inclusions.

Mafic minerals average 13 percent of the Starvation Flat Quartz Monzonite. The hornblende to biotite ratio averages about 0.7 except in the vicinity of the Huckleberry greenstone where hornblende is slightly more abundant. Biotite occurs as subhedral and euhedral crystals averaging about 0.12 inch across.

Hornblende is present in every specimen. Individual crystals are from about 0.04 to 0.16 inch long. Crystals are subhedral to euhedral and have reasonably consistent optical properties throughout the pluton. They are commonly partly altered to epidote or chlorite. Optical properties and accessory minerals are given in table 2.

Small rounded mafic inclusions composed mostly of hornblende, biotite, and plagioclase are common, though not numerous, throughout the pluton. Only along the south border of the body where the quartz monzonite intrudes greenstone of the Huckleberry Formation is the identity of these inclusions known. There they are many times more numerous than in the interior and are demonstrably xenoliths. In specimens collected within a few hundred yards of the contact, the average plagioclase composition is more calcic by about An_{10} than in the typical quartz monzonite, and mafic minerals are almost 50 percent more abundant. Hornblende predominates over biotite in the border rock, which is the reverse relationship of that in the bulk of the pluton. The average modal composition of "normal" and border-zone rock is as follows:

	Normal (nine specimens)	*Border zone (four specimens)*
Plagioclase	37	43
Potassium feldspar	24	15
Quartz	26	23
Mafic minerals	13	19

Mafic minerals in border specimens are concentrated in small clusters, along with sphene, apatite, and opaque minerals. In addition, clinopyroxene ($Z \wedge C = 37°$) is present in the cores of some hornblende crystals.

During emplacement, chemical and physical conditions in the magma apparently allowed rapid assimilation of the greenstone. This is reflected not only by the larger number of inclusions near the contact, but also by compositional and mineralogical differences between the contaminated marginal rock and the relatively uncontaminated interior. The contamination effects appear to extend less than 1 mile into the interior of the pluton, and diminish most rapidly within the first 500 feet from the contact with the greenstone.

OTHER ROCKS INCLUDED WITH THE STARVATION FLAT QUARTZ MONZONITE

A rock of unknown extent, which may be either a porphyritic phase of the Starvation Flat Quartz Monzonite or a separate plutonic mass, crops out due north of Starvation Lake and for several miles along State Highway 294 north of the area. Although its groundmass texture is similar to that of the Starvation Flat Quartz Monzonite, the rock contains phenocrysts of microcline as much as 3 inches long and almost no hornblende. In all other respects, including optical mineralogy and accessory minerals, the rock is similar to the Starvation Flat Quartz Monzonite. If this rock is not a phase of that pluton, the similarities suggest that it may be at least genetically related.

A leucocratic dikelike body intrudes the Starvation Flat Quartz Monzonite in secs. 34 and 35, T. 35 N., R. 40 E., and the two sections immediately to the south. In outcrop the rock resembles aplite and contains small pod-shaped bodies of coarser grained material. The coarser grained pods are less than 1 inch in diameter, but some are as much as 4 inches long and crudely tabular. In addition, the rock contains small miarolitic cavities. Most of the rock is pale pink to light gray. Biotite, the only dark mineral, makes up less than 1 percent of the body. The rock is unusual in that the fine-grained part is a mass of graphic intergrowths of quartz in albite and potassium feldspar (fig. 12). The graphic texture, however, is not easily detected in hand

FIGURE 12. — Photomicrograph of the fine-grained dike-form mass of graphically intergrown rock that intrudes the Starvation Flat Quartz Monzonite. The texture is typical of the entire rock, not just an isolated area. Long dimension of photograph is about 2.5 mm. Crossed nicols.

specimens. None of the rock has the xenomorphic texture typical of aplitic dikes. The average modal composition of four specimens is as follows:

Albite	28
Orthoclase (microperthite)	36
Quartz	36
Biotite	Trace

The homogeneity of the rock is indicated by the individual modes, none of which vary more than 2 percent from the average. Although an unknown, but probably small, amount of error is introduced, the mode can be used to approximate a norm because the plagioclase composition is only An_2, and the microperthitic potassium feldspar contains few lamellae of albite. The mode plots in the SiO_2-$NaAlSi_3O_8$-$KAlSi_3O_8$ minimum trough for 2,000 bars H_2O pressure (Tuttle and Bowen, 1958, p. 55), but falls about 10 percent off the ternary minimum point for that pressure in the direction away from the $NaAlSi_3O_8$ corner.

PHILLIPS LAKE GRANODIORITE (AND ASSOCIATED DIKES)

LOCATION, EXTENT, AND TOPOGRAPHIC EXPRESSION

The name Phillips Lake Granodiorite is here coined for the muscovite-biotite granodiorite exposed so well around Phillips Lake, its type locality, in sec. 34, T. 34 N., R. 41 E. This pluton underlies about 60 square miles in the northeast corner of the report area. A plutonic rock that is mineralogically, texturally, and modally similar has been found 17 miles north of the area on Huckleberry Mountain in the Spirit quadrangle (Yates and Engels, 1968, p. D245). To the east, the granodiorite is found in the Tacoma Creek drainage in the northwest quarter of the Newport 30-minute quadrangle. The full extent and configuration of this body in the areas to the north and east have not yet been determined, but the pluton appears to underlie a much larger area outside the quadrangle.

Unlike other plutons in the report area, the Phillips Lake Granodiorite underlies the highest elevations, and it appears to be considerably more resistant than any of the other plutons. However, this may be due largely to the more resistant remnants of metamorphosed Precambrian roof rocks and the leucocratic dikes which inject the pluton. The divide which runs partly outside the east boundary of the area from north of Calispell Peak south to Goddards Peak is capped by small but numerous roof remnants of metamorphic rock. Most of these appear to be less than 500 yards long and in this part of the area consist of amphibolite and mica schist.

INTERNAL FEATURES

The Phillips Lake Granodiorite is easily distinguishable from all other intrusive rocks in the report area. It is dominantly a biotite-muscovite granodiorite but varies greatly in composition (fig. 11). The interstitial arrangement of the muscovite and biotite flakes among the larger crystals of anhedral quartz and subhedral plagioclase is distinctive in hand specimen as well as in thin section (fig. 13). The micas in much of the rock impart a slight foliation, which was probably formed before the pluton totally solidified.

The granodiorite is intruded almost everywhere by leucocratic dikes which fall into three general groups. The most abundant are medium- to fine-grained equigranular quartz monzonite dikes from less than 1 inch to over 200 feet thick. Where most numerous, in the eastern part of the pluton, they average about 50 feet thick and appear to be randomly oriented. The dike walls are roughly parallel and relatively planar. The quartz monzonite also forms irregular bodies, the largest of which is a small cupola about 1 mile in diameter that crops out northwest of Lenhart Meadows (sec. 23,

FIGURE 13—Stained slab of Phillips Lake Granodiorite. Feldspar is white, quartz dark gray, and biotite black. Even though the specimen contains only about 5 percent potassium feldspar, it is not foliate as is typical of most potassium feldspar-deficient parts of this body. The specimen shows the interstitial occurrence of the biotite around the felsic minerals. Muscovite, although it does not show in the photograph, has the same relation to the quartz and feldspar as the biotite. Specimen is 3.5 inches long.

T. 34 N., R. 41 E). Concentrations of quartz monzonite float suggest similar bodies may be present in the northeastern part of the Phillips Lake Granodiorite. The general distribution of the quartz monzonite dikes is shown on the geologic map by an overprint. Where it intrudes sedimentary or metamorphic rocks, this rock invariably forms sills. Poor exposures prevent precise calculations of the total area underlain by the equigranular quartz monzonite, but the area should be at least a third of that mapped as Phillips Lake Granodiorite.

Dikes belonging to the other two groups are relatively few in number and underlie probably less than 1 percent of the area shown as Phillips Lake Granodiorite. They consist of aplites and pegmatites of the more typical varieties. Pure white aplite dikes, without biotite, occur throughout the pluton but are especially common around the margins. Although they have relatively planar walls, these dikes are considerably more irregular in shape than the quartz monzonite dikes. In hand specimen, microcline, plagioclase, quartz, muscovite, garnet, and tourmaline are visible.

Pegmatities are irregularly intermixed with the aplites and also, to about the same degree, with the quartz monzonite dikes. The association with the latter is largely restricted to the margin of the pluton. Most of the pegmatites consist only of perthite, sodic plagioclase, quartz, and muscovite, but some also contain biotite, garnet, tourmaline, and rarely beryl, or columbite. Where associated with the biotite-bearing quartz monzonite dikes, the pegmatites commonly contain intergrowths of muscovite and biotite in the same pseudohexagonal tablet. The three dike types are found together in only a few places, but where they are, the pegmatites and aplites both cut the quartz monzonite dikes.

From Phillips Lake eastward to the border of the report area, all three dike types become increasingly more numerous, and in the vicinity of Calispell Peak, the granodioritic country rock makes up probably less than half the rock exposed. This increase in the ratio of dikes to host rock is accompanied by a decrease in the potassium feldspar content of the granodiorite from an average of 15–20 percent near Phillips Lake to 0–5 percent along the east border. Also, the foliation in the granodiorite, barely perceptible at Phillips Lake, becomes increasingly more pronounced toward the east border.

The concentration of dikes thus may be related to the degree of foliation and to changes in composition of the host rocks. Several hypotheses were considered to explain these relations. The most probable explanation is that during the later stages of crystallization, when the composition of the remaining melt was similar to that of the dikes, the alkali- and volatile-rich melt was removed from the interstices of the already crystallized material, perhaps by filter pressing. The mobilized melt formed the dikes; thus, the granodiorite is most deficient in potassium feldspar where the dikes are most numerous. The foliate texture results from the collapse accompanying removal of the melt, which forced the micas and remaining melt into interstices between the larger quartz and plagioclase crystals.

The partitioning of potassium between microcline, muscovite, biotite, and plagioclase is also interpreted to mean that part of the melt was physically removed from the granodiorite. Neither the muscovite nor biotite, the only other two mineral phases in the granodiorite containing appreciable amounts of potassium, systematically changes in amount with the decrease in potassium feldspar content in the eastern part of the pluton. Plagioclase composition does not change systematically either.

The bulk composition of the original magma would have been considerably more alkali rich than the composition of the Phillips Lake Granodiorite indicates if all the dikes were originally derived from the Phillips Lake magma as suggested. The presence of two micas as characterizing minerals in the granodiorite supports this suggestion. Muscovite and biotite are not the characteristic minerals one would expect in a rock of this composition, especially along the east border of the area where the feldspar ratio of the host rock is that of a quartz diorite. However, a theoretical composition resulting from the combination of the magmas from which the granodiorite and the leucocratic dikes crystallized would be one from which muscovite and biotite might crystallize.

To arrive at a weighted-average mode for the granodiorite and dikes combined, the ratio of dike rock to pluton was estimated from surface exposure. As close as can be estimated, about 35 percent of the area shown as granodiorite on the map is underlain by dike rock. By using this estimate, the average modal compositions of the granodiorite and the dike rocks were weighted proportionately, and the following mode calculated:

Plagioclase	40
Potassium feldspar	17
Quartz	30
Mafics (including muscovite)	13

This mode still plots in the granodiorite field, but near the granodiorite-quartz monzonite boundary. If the volume of dike rock is actually greater than that of the granodiorite, an original magma from which they could have crystallized would have a quartz monzonite composition.

Except for the large remnants of roof rock in the eastern part of the pluton, almost no inclusions of any kind have been found within the Phillips Lake Granodiorite or the associated dike rocks. Even near the borders, the only perceptible difference from the rock in the interior, with respect to foreign material, is an increase in the amount of micas.

Brecciation and shearing resulting from faulting have been recognized only along the northwest border of the Phillips Lake Granodiorite and west and south of Bear Canyon, where the rock is highly fractured in places and recemented with quartz, chlorite, and epidote.

More than one set of joints is found in many outcrops of the Phillips Lake Granodiorite. Mineralized joints are cut by nonmineralized joints, suggesting more than one generation. Joint attitudes were not systematically measured and recorded on the map because most of the rounded outcrops into which the granodiorite characteristically weathers have obviously been rotated or moved by frost heaving.

CONTACT RELATIONS

The contact between the Phillips Lake Granodiorite and the metamorphosed host rocks is exposed only in a few roadcuts, and in some of these, injection of the leucocratic dikes has been so intense as to obliterate contact relations. A few generalizations can be made, however.

Possibly the most striking feature of the contact is its generally shallow dip. The outcrop pattern on the map shows a very low angle dip on the north flank of Wilson Mountain and around Bell Meadow. The southeasternmost 2 miles of the contact was drawn with very little control because of poor exposure and heavy forest cover. However, the presence of highly recrystallized schist with porphyroblasts of andalusite as much as 4 inches long in the saddle north of Goddards Peak suggests that the pluton underlies this area at shallow depth and therefore that this segment of the contact is indeed shallow. Almost all the roof remnants in the eastern part of the pluton also have shallow-dipping contacts. From the vicinity of Bear Canyon south to The Tinderbox, the contact appears to be steeper.

Where exposed, the contact is highly irregular owing to numerous dikes of both granodiorite and the leucocratic rock. Along most of the contact the host rocks appear to have reacted relatively passively; one exception is the large internal septum underlying McDonald and Brewer Mountains, which was probably rotated in a counterclockwise direction by the magma.

The Precambrian host rocks surrounding the Phillips Lake Granodiorite show contact metamorphic effects for several miles from the surface trace of the contact. Because of the shallow dip of the contact, none of the Precambrian rocks south of the pluton are very far from the source of metamorphism. In the area between the Phillips Lake Granodiorite and the Flowery Trail Granodiorite, calc-silicate minerals have formed in all the carbonate-bearing rocks except those on the west flank of Eagle Mountain and at higher elevations on the mountain. Contact metamorphic effects along the segment of the contact west of The Tinderbox but south of the northwest-bordering fault do not appear to extend as far into the host rocks as they do south of the pluton. In the McHale Slate, which is intruded by the granodiorite at The Tinderbox and on the hill immediately to the north, recrystallization is not apparent at distances greater than half a mile from the contact. This is probably because the contact is steep here but may be related to the greater distance from the Flowery Trail Granodiorite. The latter, which is an older intrusive, appears to have "prepared" the host rocks for the Phillips Lake Granodiorite. Whether this "preparation" consisted of preheating or was of a chemical nature is not known, as potassium-argon dates suggest that the Flowery Trail Granodiorite is about 100 m.y. (million years) older than the Phillips Lake Granodiorite.

Along the south boundary of the Phillips Lake Granodiorite, recrystallization of the host rocks at the contact has developed mineral assemblages indicative of the transition between the albite-epidote-hornfels facies and hornblende-hornfels facies of Turner and Verhoogen (1960, p. 511). Mineral assemblages in a single thin section of a metamorphosed carbonate-bearing rock typically include albite, epidote, clinozoisite, quartz, biotite, microcline, diopside (rare), and abundant medium-green amphibole that is highly birefringent. Highly quartzose pelitic rocks form coarse-grained quartz-mica schists that only rarely contain aluminum silicates. A typical assemblage includes quartz, albite, muscovite, biotite, tourmaline, and, in some specimens, andalusite. The assemblages developed in pelitic rocks are clearly indicative of the albite-epidote-hornfels facies, but the appearance of diopside and a hornblende-like amphibole in the carbonate-bearing rocks suggests the transition to the hornblende-hornfels facies.

In the roof remnants along the east border of the area, the rocks are well into the hornblende-hornfels facies. As most of the rocks were originally carbonate bearing, a typical assemblage includes diopside, hornblende, quartz, and plagioclase of intermediate composition. On the spur trending northwest from Calispell Peak, a small roof remnant contains the assemblage vesuvianite, scapolite, diopside, quartz, calcite, clinozoisite, and plagioclase of intermediate composition.

At distances greater than 2,000 feet from any contact of the Phillips Lake Granodiorite, all rocks are well down into the albite-epidote-hornfels facies, but almost all are, nevertheless, thoroughly recrystallized. The difficulties in separating the contact metamorphic effects of the Phillips Lake Granodiorite from those of the Flowery Trail Granodiorite to the south have already been discussed. See section "Flowery Trail Granodiorite, Contact Relations." Pronounced metamorphic effects do not extend more than about half a mile south of the Flowery Trail Granodiorite, and so most of the metamorphic effects more than 1 mile north of that pluton are probably due to the Phillips Lake Granodiorite.

PETROLOGY

Plagioclase is by far the most abundant mineral in the granodiorite. Crystals average 0.12–0.2 inch in length and are mostly subhedral. Composition averages about An_{22}. Normal zoning with minor reversals is present, although it is not obvious in all crystals. The most sodic zone measured is An_{16} and the most calcic An_{37}. The plagioclase is generally glass clear and free of the normal haze of sericite and other late-stage alteration products commonly found in plagioclase from plutonic rocks, but it does contain small euhedral to subhedral crystals of clinozoisite and epidote similar to those in the Flowery Trail Granodiorite. These inclusions and the highly discordant ages obtained from potassium-argon analyses of muscovite and biotite suggest that the granodiorite has been metamorphosed.

Potassium feldspar content averages about 11 percent but is highly variable, as the modes show (fig. 11). Only microcline has been identified, but orthoclase may also be present. The crystals vary in size but are about the same as the plagioclase. Locally, the rock is notably porphyritic, with microcline phenocrysts as much as 1 inch long. Smaller microcline crystals are mostly anhedral and appear to be interstitial to other minerals. As with the plagioclase, no alteration is apparent other than a slight clouding in some crystals. Small crystals of muscovite are common within potassium feldspar crystals in some thin sections and probably formed during metamorphism.

The quartz is a very distinctive pale gray violet and most commonly occurs in clusters of several grains which look like one large crystal. Clusters are from 0.2 to 0.4 inch across and are roughly oval in cross section. Inclusions are rare. Myrmekitic intergrowth of quartz and plagioclase are common.

Biotite and muscovite crystals average about 0.04 inch in length and occur in clusters as much as 0.4 inch across. Commonly both micas rim or mantle either quartz or feldspar. All crystals are subhedral and show almost no signs of alteration. Biotite is abundant and widespread, but muscovite is locally almost absent. The total amount of mica averages about 15 percent of the rock; the muscovite to biotite ratio averages about 0.3.

Epidote is found in all specimens but consistently amounts to less than 1 percent of the rock. Crystals are small, averaging about 0.008 inch. Some of the epidote may be primary, as many of the larger crystals have conspicuous cores of allanite.

Apatite, garnet, magnetite, zircon, and tourmaline occur as accessory minerals, but only apatite and zircon are widespread. Apatite is most abundant, and some crystals are as much as 1 mm long. Magnetite is rare or absent in almost all specimens. Garnet and tourmaline are present in only a few specimens and are probably the product of metasomatism accompanying instrusion of nearby pegmatites or aplites.

TWO-MICA QUARTZ MONZONITE

LOCATION, EXTENT, AND TOPOGRAPHIC EXPRESSION

The west edge of a two-mica quartz monzonite pluton crops out on the east flank of Nelson Peak. An area of about 2 square miles along the east border of the report area is underlain by the pluton. The total extent is not known, but anomalously gentle slopes continue for about 4 miles to the southeast in the adjacent Newport quadrangle and indicate a surface exposure considerably greater than that within the report area. This quartz monzonite, like most of the plutonic rocks in the area, forms a topographic low and gentle slopes except where in contact with more resistant metamorphic rocks.

INTERNAL FEATURES

Muscovite and biotite characterize this pluton. They were found everywhere that the rock was examined. The ratio of muscovite to biotite averages about 0.30 and varies only slightly from place to place. The Phillips Lake Granodiorite and its associated dikes are the only other rocks in the area that contain two micas.

The quartz monzonite has a hypidiomorphic-granular texture and is uniformly coarse grained. It is almost invariably deeply weathered and friable. The feldspars have a bleached appearance, and biotite is commonly surrounded by a brown stain of iron oxides.

Texturally and mineralogically the rock appears to be quite uniform, but exposures are so poor that internal variations could easily go unnoticed. No dikes or inclusions were found, and exposures are not such to show joint development.

Modal analyses of seven specimens from different localities scatter on both sides of the quartz monzonite-granodiorite boundary (fig. 11); however, the average composition, shown below, is on the quartz monzonite side and is close to the calculated average modal composition of the Phillips Lake Granodiorite and associated leucocratic dikes. (See section "Phillips Lake Granodiorite (and associated dikes), Internal Features.")

	1	*2*
Plagioclase	38	40
Potassium feldspar	21	17
Quartz	30	30
Mafics (including muscovite)	11	13

1. Average modal composition of two-mica quartz monzonite.
2. Calculated average modal composition of Phillips Lake Granodiorite and associated leucocratic dikes.

If not just coincidental, the similarity suggests that the two plutons are related or are disconnected parts of the same pluton.

In the Rattlesnake Hills, 15–20 miles south-southeast of Loon Lake, Griggs (1966) mapped another two-mica quartz monzonite which is almost identical in modal composition, texture, and appearance with the one in the report area. If all three two-mica bodies are genetically related, they represent a widespread plutonic entity in northeastern Washington.

CONTACT RELATIONS

The contact between the two-mica quartz monzonite and the Prichard Formation is nowhere exposed, but the relation of topography and inferred contact configuration suggests that it is steeply dipping.

Contact metamorphic effects are noticeable 1 mile southwest of the pluton but cannot confidently be distinguished from those caused by the coarse-grained quartz monzonite immediately to the south. Just east of Nelson Peak, argillite 200 feet from the quartz monzonite has been recrystallized to a quartz-muscovite-biotite hornfels with small porphyroblasts of chlorite and garnet. The chlorite shows signs of reaction and probably became unstable when the garnet began crystallizing. Although a contact metamorphic occurrence, the assemblage is most like the quartz-albite-epidote-almandine subfacies of the greenschist facies (Fyfe and others, 1958, p. 224).

PETROLOGY

The plagioclase in this rock is oligoclase. The crystals are subhedral to euhedral and commonly have sodic rims and calcic cores. Some plagioclase crystals contain crystals of biotite and muscovite. Large parts of some plagioclase crystals are replaced by irregular-shaped grains of microcline.

All potassium feldspar appears to be microcline. The characteristic grid twinning is well developed. The potassium feldspar does not occupy interstices between other minerals, as it would if it were a late-stage filling, but occurs as subhedral crystals with the same relation to surrounding minerals as the plagioclase. In this respect it is different from that in most of the plutonic rocks of the report area.

Large gray anhedral knots of quartz fill spaces between other minerals. The quartz contains inclusions of all other minerals in the rock and shows little undulatory extinction.

Muscovite and biotite are randomly oriented and not wrapped around felsic minerals as in the Phillips Lake Granodiorite. They most commonly occur in patches in which the two minerals are intergrown with one another. Biotite has the pleochroic formula X=golden tan, Y=Z=reddish brown.

Apatite and zircon are the only common accessory minerals in the rock. Zircon is unusually abundant and dots the biotite with pleochroic halos. Opaque minerals are unusually scarce.

COARSE-GRAINED QUARTZ MONZONITE

LOCATION, EXTENT, AND TOPOGRAPHIC EXPRESSION

An area of about 15 square miles in the southern part of the report area is underlain by coarse-grained biotite quartz monzonite. The pluton has a highly irregular outline and crops out in two separate areas divided by a septum of Belt rocks just east of Deer Lake. The easternmost of the two areas extends beyond the report area for an unknown distance.

The pluton is a perfect example of how easily the plutonic rocks in this region are eroded relative to the metamorphic rocks. The part of the pluton northeast of Deer Lake forms an open-ended basin, and a noticeable break in slope marks almost the entire length of the contact in this area. South of Deer Lake and northeast of Loon Lake, almost the entire axis of this part of the pluton is marked by prominent valleys. North and west of Loon Lake the pluton forms low hills or underlies flat alluviated areas.

Most of the rock is deeply weathered. In numerous 10–15-foot roadcuts, in all parts of the area, only loose friable rock can be found. Natural exposures are rare, and most of them are highly weathered. Only on State highway 292, 3½ miles east of Springdale, and on the road between Loon Lake and Deer Lake is the rock reasonably fresh.

INTERNAL FEATURES

The coarse-grained quartz monzonite is easily identified by grain size: The average size is greater than half an inch. Grains of quartz and feldspar are about equal in size, but grains of biotite are only about ⅛–¼ inch across.

Although most of the rock has a hypidiomorphic-granular texture and is not obviously porphyritic, phenocrysts of pink potassium feldspar as much as 2 inches long are locally common. Even where the rock is nonporphyritic, the abundant potassium feldspar imparts a pink cast.

Locally, the distribution of potassium feldspar is not uniform. In places, as much as a cubic yard of rock may consist of 50–60 percent potassium feldspar. Pods of potassium feldspar are irregular in shape and common throughout the pluton. Textural relations and grain size within these pods are the same as those in the rest of the quartz monzonite. The pods may represent crystal accumulates. Alternatively, they may have formed by some sort of localized filter pressing process or may be metasomatic.

The nonuniform mineral distribution makes it difficut to estimate the overall composition of the pluton with accuracy. The average composition shown in figure 11, however, is probably reasonably accurate because it represents modal analyses of several large samples from each locality.

Samples from a number of places in secs. 18 and 19, T. 30 N., R. 42 E., show a pronounced bimodal grain size. Crystals of potassium feldspar, quartz, plagioclase, and biotite averaging about 0.2 inch in size make up about 65 percent of the rock and are set in a groundmass of the same minerals with an average grain size of about 0.05 inch. This pronounced textural variation may be due to thermal or pressure quenching of a crystal-magma mixture. These textural variations could not be mapped, but they have been found only on and around the projections of preintrusion faults (pl. 2). As thin sections show no cataclasis, the texture is presumed to be a primary feature and is probably due to some deviation from the normal crystallization conditions of the pluton. The melt may have intruded host rock that was shattered enough so that it could not retain the fluid pressure maintained by the rest of the intrusive. Upon loss of fluid pressure, relatively rapid crystallization took place locally and possibly sealed the system from any further leakage.

Leucocratic muscovite-bearing bimodal rocks are present at the outer margins of the pluton, near the contact. These, however, are finer grained than the bulk of the pluton.

Inclusions of any sort are not abundant in this pluton, but dikes are common, especially around the margins. Some are normal-looking aplites and fine-grained muscovite-bearing rocks with graphic texture. These are probably genetically related to the quartz monzonite because of their consistent spatial relationship. Others, which consist of a dark-green rock, are highly altered. They also cut other plutons and may not be related to this pluton.

CONTACT RELATIONS

The attitude of the contact could not be measured directly, but the relationship of its configuration to topography suggests that it dips outward between 40° and 50° in most places.

Contact metamorphic effects of the quartz monzonite are pronounced within a few hundred feet of the contact but fade out rapidly beyond that. Good exposures of metamorphosed siltite and argillite on Jumpoff Joe Mountain, Deer Lake Mountain, and on the ridge east of Benson Peak furnish excellent control for determining the degree of metamorphism caused by the quartz monzonite. At all three localities the range in chemical composition of the host rocks is about the same. Other physical and chemical conditions must have been uniform also, because the aureoles are nearly the same width at all three localities and the minerals that crystallized are almost identical. Less than 200–300 feet from the contact, the most common mineral assemblage is andalusite-cordierite-biotite-muscovite-quartz, indicative of the hornblende-hornfels facies of Turner and Verhoogen (1960, p. 513). Beyond 200–300 feet, andalusite and cordierite are absent.

Quartz-albite-muscovite-biotite of the albite-epidote-hornfels facies is the typical assemblage in the interval from 200 or 300 to about 600 feet. All the rocks in this interval are thoroughly recrystallized but do not contain any of the higher grade minerals. Beyond 600 feet, chlorite porphyroblasts, fine-grained muscovite, incipient biotite, albite, and quartz are the characteristic minerals. Much of the rock beyond 600 feet does not appear to be thoroughly recrystallized. Beyond about 1,500 feet, few obvious metamorphic effects are observable.

PETROLOGY

The mineralogy of the coarse-grained quartz monzonite is relatively simple and uniform. Plagioclase, the most abundant mineral, averages An_{20} and occurs as white euhedral to subhedral crystals ranging in size from about 0.2 to 0.8 inch. Most crystals show some zoning, but it is not particularly pronounced. Locally the plagioclase is sericitized and is pale green in hand specimen.

Microperthitic orthoclase crystals are subhedral to anhedral and range in size from about 0.2 inch to more than 2 inches. One of the characteristics of the rock, the significance of which is not understood, is the presence of a thin hairline film of potassium feldspar around most plagioclase crystals. No evidence was seen to indicate whether this film resulted from exsolution or from crystallization of the late-stage potassium-rich solutions.

Biotite constitutes about 7 percent of the rock and is the only mafic mineral. It is generally finer grained than the other major minerals. The borders of most crystals are slightly chloritized, but the centers are unaltered.

Some sphene crystals are as large as a quarter of an inch across and easily seen in hand specimen. The other accessory minerals, given in table 2, are normally fine grained and less abundant.

MUSCOVITE QUARTZ MONZONITE

LOCATION, EXTENT, AND TOPOGRAPHIC EXPRESSION

Leucrocratic muscovite quartz monzonite is exposed in three areas in the southeastern part of the report area and underlies a total area of about 2 square miles. The largest exposure, in sec. 9, T. 30 N., R. 42 E., extends about 2.5 miles east of the report area and underlies as much as 3 square miles outside the area. A. B. Griggs located small outcrops of muscovite quartz monzonite at two localities just south of the report area. One is 4 miles south of Clayton, the other 3 miles southeast of Deer Park (oral commun., 1968). The rocks at both localities are identical in every respect and are undoubtedly genetically related to the muscovite quartz monzonite in the report area. In addition, Griggs reported that the same rock type crops out at various places in the mountainous part of the Clayton quadrangle. Although individual plutons appear to be relatively small, the rock type is apparently widespread in the region.

Like most of the other plutonic rocks in the area, the muscovite quartz monzonite forms topographic lows. The contact of the pluton with the more-resistant metamorphosed host rock generally is a break in slope. Although exposures are generally poor, contacts can be reliably mapped on the basis of this break in slope and the presence in the soil of mica from the decomposed plutonic rock.

CONTACT RELATIONS

Other than small variations in grain size, the muscovite quartz monzonite at the contact is identical with that in the interior. As with some of the other plutons, the metamorphic effects on the country rock are difficult to separate from those of other plutons nearby. On Blue Grouse Mountain, which is almost 1 mile from any any other pluton, the contact metamorphic effects are very slight. At the contact, siltite has been recrystallized into medium-grained quartz-muscovite schist. However, hand specimens of host rock from more than 50 feet away show almost no metamorphic effects, and thin sections of this rock show only slight recrystallization.

The south border of the largest muscovite quartz monzonite intrusion in the report area is cut by numerous huebnerite-bearing quartz veins, as is the adjacent host rock. Parts of the contact are greisenized, and the quartzite and siltite as much as several hundred feet from the contact are pock marked where limonite psuedomorphs have been leached out. These veins and the associated mineralization appear to be related to the quartz monzonite.

PETROLOGY

On the basis of mineralogy, the rock is quartz monzonite in composition and is referred to as such in this report. However, because the plagioclase is sodic albite, the rock could be classified as granite chemically. The average mode, a chemical analysis, and CIPW norm of an apparently representative sample are as follows:

Chemical analysis[1] *(weight percent)*		*CIPW norm (weight percent)*		*Average of 6 modes (volume percent)*	
SiO_2	75.1	Q	32.3	Plagioclase	34
Al_2O_3	14.0	C	1.4	Potassium feldspar	33
Fe_2O_3	.22	or	27.2	Quartz	27
FeO	.16	ab	34.7	Muscovite	6
MgO	.15	an	2.4		100
CaO	.53	en	.4		
Na_2O	4.1	fs	.2		
K_2O	4.6	mt	.3		
H_2O-	.06	il	.1		
H_2O+	.94	cc	.1		
TiO_2	.04		99.1		
P_2O_5	....				
MnO	.06				
CO_2	.05				
	100.01				

[1]Analyzed sample collected in NE¼, sec. 25, T. 30 N., R. 41 E. Analyzed by P. L. D. Elmore, S. D. Botts, Lowell Artis, James Kelsey, Gillison Chloe, James Glenn, and Hezekiah Smith.

The mineralogy of the rock varies little from place to place. Although the proportion of the minerals varies slightly, all specimens are made up of the same minerals. The only noticeable textural variations are slight differences in grain size. Near the margins and in the smaller intrusions, the grain size is only about 0.15 inch, compared with about 0.2 inch for the bulk of the rock. The texture of the entire quartz monzonite is hypidiomorphic-granular. Color is pink to cream, depending on how altered the potassium feldspar is. Small spots of limonite a few inches to a few feet apart are present in most of the quartz monzonite. No dikes or inclusions were found anywhere in the rock.

Average modal analyses of this rock indicate that potassium and sodium feldspar contents are almost equal. Miller (1969, p. 5) earlier reported the plagioclase composition to be about An_{10}, but additional work using immersion oils indicates an average composition of An_3. Normative An of plagioclase calculated from two analyses averages 3.5. The crystals are subhedral to euhedral, well twinned, and almost devoid of zoning. The rock does not appear to have been albitized.

Potassium feldspar is subhedral to anhedral pink or cream-colored microcline. In thin section most crystals show the characteristic grid twinning. Most of the microcline is microperthitic, but the sodic phase appears to have formed from normal exsolution rather than by albitization. Along with quartz, the microcline occupies interstices between other minerals.

Muscovite, the sole characterizing mineral, averages about 6 percent of the rock. No textural features in hand specimen or thin section suggest that any of the muscovite is secondary.

The rock contains no mafic minerals other than specks of magnetite, pyrite, or limonite. Accessory minerals, in order of abundance, are garnet, apatite, and zircon. The garnet is pale brownish red and ranges from a trace to 2 percent.

CENOZOIC PLUTONIC ROCKS

SILVER POINT QUARTZ MONZONITE

LOCATION, EXTENT, AND TOPOGRAPHIC EXPRESSION

The Silver Point Quartz Monzonite underlies about 30 square miles in the southernmost part of the report area. It was named by Miller (1969) for exposures at Silver Point on the west shore of Loon Lake. Although its total extent is not known, this pluton apparently underlies an extremely large area. The rock is exposed in roadcuts along U.S. Highway 2 that are 15 miles east-northeast of Loon Lake and may extend beyond there. South of the area, the pluton extends at least 2 miles into the northern part of the Clayton quadrangle. Rock similar in appearance to the Silver Point Quartz Monzonite is found about 20 miles southwest of Loon Lake on the west side of the Wellpinit quadrangle but may not belong to the same plutonic mass (A. B. Griggs, oral commun., 1968).

The Silver Point Quartz Monzonite forms areas of moderate to low relief. Almost all exposures are deeply weathered. Along U.S. Highway 395 near the south border of the area, some new roadcuts 20 feet deep do not penetrate unweathered rock. In the hills bordering Ahren Meadows, however, most of the quartz monzonite is relatively fresh, probably because of glacial erosion. The least weathered and best exposures are along the west and south shores of Loon Lake, where roadcuts have recently been blasted into very fresh rock.

INTERNAL FEATURES AND PETROLOGY

The Silver Point Quartz Monzonite is a porphyritic hornblende-biotite quartz monzonite. One of the most striking features of the rock is its lack of internal variation, either in composition or texture. Modal analyses (fig. 11) plot in a very small field that would probably be even smaller if slightly larger slabs had been counted. Several specimens collected 10–15 miles east of Loon Lake do not vary more than 2 percent in any mineral phase from specimens obtained around Loon Lake. Near the borders, the pluton shows no obvious contamination effects and no variability in modal composition. Color index ranges from 14 to 21, but for almost all specimens it falls in a relatively narrow range between 14 and 18. The ratio of hornblende to biotite is consistently between 0.75 and 1.0.

The texture of the Silver Point Quartz Monzonite is unlike that of any other plutonic rock in the report area. The distinctive feature is the groundmass, which is composed of two size groups of crystals (fig. 14): Crystals of hornblende, biotite, plagioclase, potassium feldspar, and quartz averaging 0.1–0.2 inch in size make up about 40 percent of the rock; the rest of the groundmass consists of crystals averaging only 0.02–0.06 inch in size and is composed of all mineral phases found in the rock. This bimodal texture, like the modal composition of the quartz monzonite, varies little from place to place. The greatest element of textural variability is the size and content of potassium feldspar phenocrysts, which range from 1/4 to 1 1/2 inches and from less than 1 to about 5 percent.

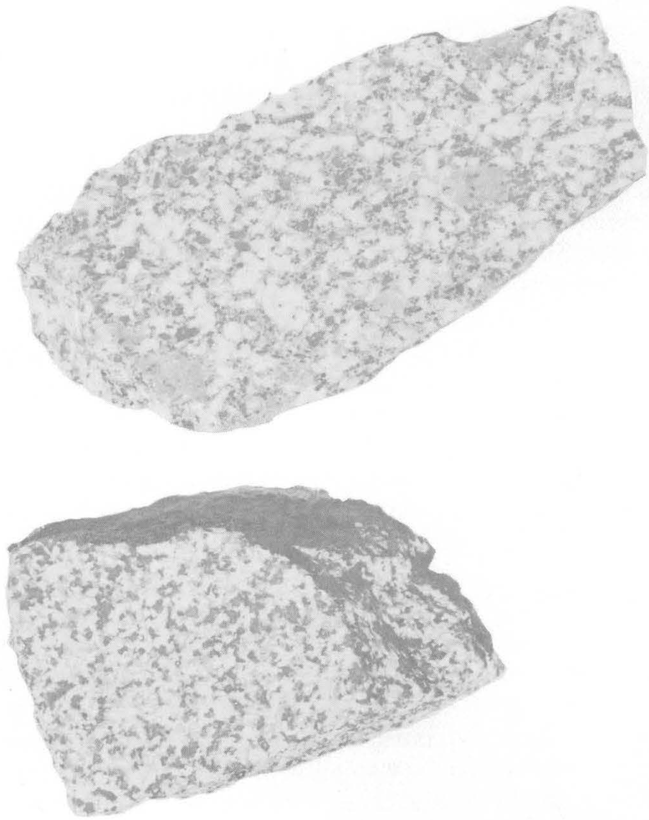


FIGURE 14.—Stained slab of Silver Point Quartz Monzonite (upper) and granodiorite from the small pluton southeast of Springdale (lower). The Silver Point specimen is a perfect example of the unusual texture found throughout that pluton. The large felsic and mafic minerals are "floating" in a finer grained interstitial mass of the same mineral types. Sparse potassium feldspar phenocrysts are typical of the pluton. The granodiorite specimen shows the high color index and equigranular texture typical of this body and contrasts sharply with the texture of the Silver Point specimen to which it is probably genetically related. Plagioclase is white, potassium feldspar gray, and hornblende and biotite black. Quartz is too fine grained to show well. The Silver Point specimen is 5 inches long, and the slabbed face of the granodiorite 3 inches long.

The table following compares the average modal composition of four specimens on the basis of the size range of crystals:

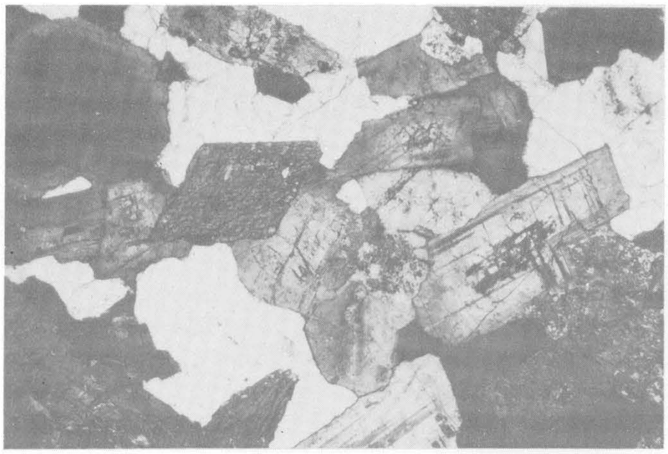
	Entire pluton	*Four specimens, entire rock*	*Four specimens, groundmass only*	*Four specimens, groundmass recalculated to 100 percent*	
	(*a*)	(*b*)	(*c*)	(*d*)	*d—b*
Plagioclase	40	42	13	21	−21
Potassium feldspar	25	26	22	36	+10
Quartz	18	18	15	25	+ 7
Mafics	17	14	11	18	+ 4
Large crystals		....	39	....	....

The specimens are from the southeast end of Loon Lake. As the last column shows, the proportion of plagioclase in the groundmass is considerably lower than it is in the rock as a whole, whereas the proportion of potassium feldspar, quartz, and mafic minerals is higher. If the larger crystals are considered to be earlier crystallization products, as they probably are, then the different proportion of minerals in the two size groups would suggest that a normal but pronounced differentiation occurred during crystallization. The average modal composition of the groundmass is plotted in figure 11 for comparison with the average composition of the rock as a whole. Some event, such as a rapid loss of heat or volatiles, caused approximately the last 60 percent of magma to crystallize so rapidly that it could not react with already crystallized minerals.

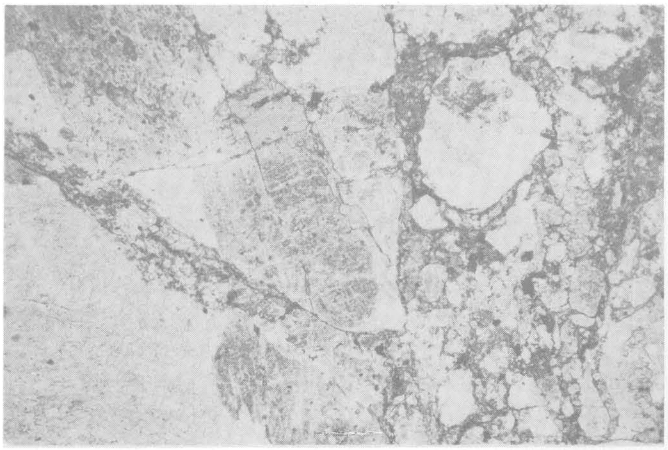
The Silver Point Quartz Monzonite shows none of the features interpreted as recrystallization effects in the Flowery Trail Granodiorite and Phillips Lake Granodiorite. Mafic minerals appear to have the same optical properties throughout the pluton, and the feldspars contain the haze of alteration products normally found in feldspar of plutonic igneous rocks.

Several north- to north-northwest-trending shear zones have been recognized west of Loon Lake. The largest zone, immediately west of the lake, is 500 feet wide at one place and was traced for more than 4 miles. In sec. 31, T. 30 N., R. 41 E., one of the zones intersects the contact between the Silver Point Quartz Monzonite and the fine-grained quartz monzonite, apparently without offsetting it. The exposures are so poor in this area, however, that the contact could be offset as much as half a mile in an apparent right-lateral sense.

Within these zones, the rock is highly sheared and displays cataclasis. Thin, anastomosing seams of silica and chlorite bond the rock and make it extremely strong. Small aplite dikes within the zones are highly broken and are offset in a right-lateral sense along individual shears. In thin section all minerals appear highly granulated and almost all mafic minerals appear to be ground up and chloritized (fig. 15). The highly



A



B

FIGURE 15.—Photomicrographs of Silver Point Quartz Monzonite. *A*, Typical undeformed Silver Point Quartz Monzonite. Crossed nicols. *B*, Cataclastic Silver Point Quartz Monzonite from the shear zone south and west of Loon Lake. Note that all the mafic minerals, so abundant in *A*, are not present; presumably they have been broken and altered to chlorite and opaque minerals. Plane–polarized light. Long dimension of both photographs is about 4 mm.

deformed rock in the shear zones grades into completely undeformed rock over a distance of 20 or 30 feet.

Plagioclase, the most abundant mineral, occurs as subhedral to euhedral crystals in both the fine- and coarse-grained fractions of the groundmass. Average composition is about An_{20}, with zoning ranging from An_{15} to An_{35}.

Potassium feldspar, which appears to be microperthitic orthoclase, is the only mineral that forms phenocrysts. It does not show the microcline grid twinning, but its crystal symmetry was not checked by X-ray. Phenocrysts are euhedral, but groundmass minerals are anhedral and fill intergranular spaces. In fresh specimens it is pink and easily distinguished from the plagioclase.

Quartz makes up about 18 percent of the rock but is difficult to see without a hand lens, as most grains are less than 0.04 inch long. It shows undulatory extinction and is interstitial to all other minerals.

Hornblende occurs as subhedral to euhedral crystals as much as 0.4 inch long. Average size is about 0.12 inch. Most shows no alteration. The optical properties shown in table 2 were checked in six samples from various localities, and little variation was noted.

Biotite, which is present throughout the pluton, is closely associated with hornblende. More biotite is present in the fine-grained groundmass than in the coarse. Only a few biotite crystals are as much as 0.12 inch in size, whereas most of the hornblende is this size or larger.

Sphene is the most obvious accessory mineral and is easily seen in hand specimen. Sphene, magnetite, and less conspicuous apatite occur in about equal proportions. Small zircon crystals are scattered throughout the rock but are not abundant. Allanite, relatively rare, occurs in crystals as much as 0.08 inch long.

CONTACT RELATIONS

Almost nowhere is the contact between the Silver Point Quartz Monzonite and any other rock exposed. Even where contacts are shown on the map, the actual trace is almost invariably covered. Because of this, no field evidence was found to ascertain the relative ages between the Silver Point Quartz Monzonite and the two smaller plutons adjacent to it west of Loon Lake. Along this part of the contact, the quartz monzonite exhibits no reduction in grain size, no foliation, and no increase in the number of inclusions normally found in the rock. Neither of the two adjacent plutons exhibit these features either, and no apophyses or dikes were found that might indicate relative age.

On the south flank of Loon Lake Mountain, the Silver Point Quartz Monzonite intrudes the Revett, Burke, and St. Regis Formations, and just east of Loon Lake Mountain, it intrudes the lower part of the Wallace Formation. The rocks are highly weathered here also, and none of the contacts are exposed. However, small but resistant outcrops of metamorphic rocks are found within 100 feet of the most uphill occurrence of soil containing decomposed granitic debris.

Most of the rock intruded ranges from silty argillite to quartzite. Within a zone approximately 400 yards wide next to the pluton, the argillitic rocks have been converted to quartz-albite-muscovite-biotite schist. Where the quartz monzonite intrudes the lower part of the Wallace Formation in sec. 32, T. 30 N., R. 42 E., the carbonate-bearing siltstone is recrystallized to a fine-grained quartz-albite-diopside-garnet-scapolite hornfels. The presence of albite in this group of minerals probably indicates a disequilibrium assemblage. Assuming an approximately constant pressure at the particular level in the host rocks now exposed, during the emplacement of the quartz monzonite the temperature could have increased through the stability range of the albite-epidote-hornfels facies and into that of the hornblende-hornfels facies. However, because the albite has not been converted to more calcic plagioclase, the temperature probably did not remain high long enough for a stable assemblage indicative of the hornblende-hornfels facies to form. At distances of more than half a mile from the quartz monzonite, the host rocks show almost no metamorphic effects.

GRANODIORITE

A small elongate body of hornblende-biotite granodiorite 2 miles southeast of Springdale underlies an area of about 1 square mile. The rock does not occur any other place in the report area and has not been reported outside the quadrangle. Although this pluton and the Flowery Trail Granodiorite are similar modally and mineralogically, their ages are considerably different, according to potassium-argon analyses.

INTERNAL FEATURES AND PETROLOGY

The granodiorite is easily distinguished from other plutonic rocks in the report area by its high color index and by the fact that it is equigranular. In addition, it exhibits hypidiomorphic-granular texture and is completely unfoliated. The pluton is nearly uniform internally, with no obvious variations in mineralogy, grain size, or texture.

The equigranular texture and high color index impart a salt-and-pepper appearance (fig. 14). Both feldspars are white, and quartz, although present, is not obvious. Abundant specks of honey-brown sphene averaging 0.05 inch in length occur in all specimens and make up as much as half a percent of some. Average grain size of the major minerals is 0.1–0.15 inch.

Small aplite dikes as much as 6 inches wide cut the pluton but are not common. Rounded mafic inclusions from 0.5 to 6 inches in diameter are scattered throughout the pluton. Irregular-shaped, wispy mafic segregations as much as 1 foot across are found from place to place but are not abundant.

Plagioclase, the most abundant mineral, forms euhedral to subhedral crystals, whose average composition is between An_{25} and An_{30}. Crystals are normally zoned from about An_{35} in the cores to about An_{20} at the rims. In most specimens, both feldspars show very little alteration.

Potassium feldspar appears to be microperthitic orthoclase. No microcline twinning was observed, but the crystals were not checked for symmetry by X-ray methods. The potassium feldspar, along with quartz, occupies interstices between other minerals.

Quartz is anhedral, clear of inclusions, and only slightly strained. The granodiorite has less quartz than any pluton in the report area.

The hornblende-to-biotite ratio is approximately 1:1. Hornblende forms euhedral crystals, some of which have partially altered pyroxene cores. Both mafic minerals occupy the same textural relationships to other minerals, and neither is segregated from the other.

Sphene, by far the most abundant accessory mineral, is obvious in all hand specimens. Apatite, zircon, and magnetite are abundant but visible only in thin section.

CONTACT RELATIONS

The granodiorite is in contact with the Addy Quartzite and three other plutonic rocks. Contacts are poorly exposed, like those of most of the other plutons. On the west and north sides of the body, the contacts with the coarse-grained and the fine-grained quartz monzonites are not exposed but were mapped on the basis of float. The southeast segment of the contact with the Silver Point Quartz Monzonite is locally exposed, but evidence bearing on the relative age of the two plutons is lacking. Because the configuration of the contact is highly irregular, the dip is hard to estimate from the relation of the surface trace to topography. Along a small segment of the contact, the granodiorite intrudes the Addy Quartzite, but because the quartzite contains relatively few impurities, the pluton produced no obvious recrystallization effects. In the SW cor. sec. 2, T. 29 N., R. 40 E., almost half a mile from the contact, the impure lowermost part of the Metaline Formation contains needles of tremolite. These are the only obvious contact metamorphic effects attributable to the granodiorite.

FINE-GRAINED QUARTZ MONZONITE

Fine-grained hornblende-biotite quartz monzonite underlies about 1.5 square miles between Springdale and Loon Lake. The pluton is irregular in shape but crudely elongate in an east-west direction. Rock of this type is not found at any other place in the report area and appears to be confined to this single pluton.

In general, the rock is deeply weathered and very poorly exposed. Roadcuts along State Highway 292 about 3.5 miles east of Springdale show that the rock is highly weathered more than 10 feet below the surface. Because of this deep weathering, part of the south border was mapped on the basis of sparse float and is not well located.

INTERNAL FEATURES AND PETROLOGY

The fine-grained quartz monzonite has a rather uniform hypidiomorphic-granular texture throughout but varies somewhat in grain size. In railroad cuts along the tracks of the Burlington Northern Railroad about 3.5 miles east of Springdale, dikes of this rock have chilled borders and cut the more extensively exposed coarse-grained quartz monzonite. In this area and along the nearby roadcuts on Highway 292, the grain size is 0.02–0.05 inch, but at the northwest end of Loon Lake, it is about 0.1–0.15 inch.

In hand specimen the rock is characterized by its relatively fine grain size, a salt-and-pepper appearance due to the interspersed light and dark minerals, and a splotchy pink cast due to coloring of the potassium feldspar.

Small aplitic dikes and dark-green aphanitic dikes are common but not abundant in the pluton. Large inclusions are rare, but the small inclusionlike clots are almost everywhere in the finer grained parts of the pluton.

The mineralogy appears to vary little in the few thin sections examined. Average plagioclase composition is about An_{20}, but many of the crystals have pronounced zoning. In all specimens examined, at least some of the plagioclase crystals are noticeably larger than all other crystals. Potassium feldspar is microperthitic orthoclase and together with quartz occurs as small irregular grains interstitial to the other minerals.

Hornblende is subhedral and commonly contains cores of partially altered pyroxene. Its indices of refraction are considerably lower than in other plutons, and its pleochroism is noticeably different: $X =$ pale tan, $Y =$ pale olive green, $Z =$ pale green. The hornblende-to-biotite ratio is about 1:1, as in the granodiorite and the Silver Point Quartz Monzonite. Biotite is similar to that found in the other hornblende-biotite-bearing plutonic rocks.

Hornblende and biotite occur in about equal proportions and make up about 18 percent of the rock. To a noticeable degree, they occur in clots. The clots are larger and more apparent at some places than at others and may be partially resorbed mafic inclusions derived from the Huckleberry greenstone or the Precambrian sills of the Prichard Formation. This may explain the highly discordant potassium-argon ages obtained from the hornblende.

Sphene, by far the most abundant accessory mineral, can be seen in most hand specimens. Apatite, magnetite, and zircon are common. Allanite is present but not as abundant as the other accessory minerals.

CONTACT RELATIONS

The fine-grained quartz monzonite is in contact only with other plutonic rocks. The southern contact is very poorly exposed, and the relation of the age of this pluton to that of the granodiorite and Silver Point Quartz Monzonite is not known, nor is the attitude of the contact.

On the other hand, the contact with the coarse-grained quartz monzonite is locally well exposed and appears to dip less than 50° N. The coarse-grained pluton is clearly older than the fine-grained one. No contact metamorphic effects resulting from the intrusion of the fine-grained quartz monzonite have been observed.

DIFFERENTIATION OF THE PLUTONIC ROCKS

In the preliminary report on the north half of the report area, the existence of an apparent differentiation sequence in the plutonic rocks was suggested (Clark and Miller, 1968, p. 3). A report on the south half (Miller, 1969) suggested that two separate differentiation sequences might exist and that it might be possible to separate rocks belonging to the Kaniksu and Colville–Loon Lake batholiths on the basis of which apparent differentiation sequence a particular pluton belonged to (Miller, 1969, p. 4 and fig. 2A). Potassium-argon ages since determined for most of the plutonic rocks show that the ages of plutons thought to belong to one apparent differentiation sequence span a period of 150 m.y. Unless a differentiating magma can exist for 150 m.y., it would appear that the previously proposed differentiation sequences are more apparent than real.

Most of the muscovite-bearing rocks appear to form a crude trend, and all appear to belong to a 100-m.y. period of plutonic activity. The Starvation Flat Quartz Monzonite also belongs to this age group even though it is a hornblende-biotite-bearing rock, and it plots closer to the apparent differentiation sequence characterized by hornblende and biotite. (See fig. 11.)

It seems more than just coincidental, however, that the average modal compositions of the plutons plot on such regular curves, that all the hornblende and biotite-bearing plutons plot on one curve, and that all the muscovite and two-mica plutons plot on another. To postulate a differentiating melt existing for over 150 m.y. strains the credulity of even the most imaginative geologist. However, a differentiation may have occurred even though the parent material was not continuously molten. It is possible that this parent material, regardless of whether it was derived from igneous, sedimentary, or metamorphic rocks, melted more than once without a change in the factors controlling differentiation. During the quiescent periods between melting, differentiation of the parent material, and emplacement of the individual plutons, chemical factors could have been essentially constant. Then, when a new period of melting occurred, differentiation could have begun where it had left off at the end of the preceding period.

POTASSIUM-ARGON AGES OF THE PLUTONIC ROCKS

By JOAN C. ENGELS

Seven of the eight major plutons in the report area have been dated by the potassium-argon method, in addition to several of the other plutonic rock types associated with them. Although the apparent absolute ages of two plutons have been determined and the relative age relations between others are known, the complete sequence of intrusion has not yet been established with certainty.

Potassium determinations were made in duplicate on Baird and Instrumentation Laboratories flame photometers with a lithium internal standard. Argon analyses were made on a 6-inch 60° Nier-type sector mass spectrometer, using standard isotope dilution techniques.

K^{40} decay constants:
$\lambda\varepsilon = 0.585 \times 10^{-10}\ yr^{-1}$
$\lambda\varepsilon = 4.72 \times 10^{-10}\ yr^{-1}$
$K^{40} = 1.19 \times 10^{-2}$ atom percent

Errors ($\pm$ values) have been assigned on the basis of both experience with duplicate analyses and uncertainties in the individual runs and represent 2σ.

Early in the study of the plutonic rocks by the potassium-argon method, it was found that a pair of minerals from a single rock often gave strikingly different ages. Efforts were then made to obtain rocks from which more than one mineral could be extracted for dating. By analyzing mineral pairs, an age disturbance could be detected if the two minerals yielded different ages. Of all the rocks dated, only two, the Starvation Flat Quartz Monzonite and the Silver Point Quartz Monzonite, appear to yield concordant ages on mineral pairs.

Table 3 gives ages of the rocks analyzed, and figure 16 their locations. Examination of the two together shows that the amount of discordance between individual mineral pairs within other plutons generally increases to the south and east. The discordances suggest a widespread thermal disturbance about 50 m.y. ago in the region just east of the report area. Although the Silver Point Quartz Monzonite lies in this direction and is among the youngest intrusive bodies in the area, it is not known for certain if the thermal effects of this pluton are widespread enough to have caused the discordances in the other plutons. If the Silver Point Quartz Monzonite is the source of the discordances, the effects extend several tens of miles from the surface outcrop, or else part of the pluton underlies at shallow depth the area north and northwest of where it is presently exposed.

Hart (1961, 1964), Doe and Hart (1963), and Hanson and Gast (1967) studied the effects of younger intrusions on host rocks, using a number of dating methods. They found that the apparent ages of the host rocks range from their true age to the age of the intrusion, and they plotted apparent age of the host rock against distance from the contact. In any given area, each mineral and each dating method employed should have a distinctively shaped age-retention curve which depends on a number of factors, including size and composition of the units and pressure-temperature conditions.

In general, hornblende has retained almost all argon within a relatively short distance of the contact, whereas biotite has retained argon only at much greater distances. Muscovite is intermediate but is generally closer to biotite than to hornblende. Figure 17 diagrams these qualitative observations, without regard to distances involved or size of intrusions. The expected distribution of apparent ages for each mineral type in an older pluton can be obtained by projecting the curve onto a line extending outward from the contact with the younger body (fig. 17). The areal distribution of apparent ages obtained by more than one method on several mineral types may initially lead to confusion. The scatter of apparent ages can range from the true age of the host rock to the age of the younger pluton. Thus, in an area where isotopic ages of plutonic rocks show many discordances and apparent anomalies because of a younger plutonic event, interpretation is simplified by initially comparing ages obtained using only one dating method on a single mineral type.

The unit which seems to best illustrate this qualitative treatment of the effects of a younger plutonic event is the Flowery Trail Granodiorite. Three biotite-hornblende pairs from this pluton show a progressive age loss in an easterly direction. The oldest age obtained for the pluton is 194 m.y. on hornblende from a sample (No. 1, fig. 16) collected in the western part of the body near the Jay Gould mine. Biotite from the same rock yields an age of only 98 m.y. Hornblende from another sample (No. 2) collected about 1.5 miles farther east gives an age of 183 m.y., and biotite an age of 84 m.y. Sample 3, collected about 3.3 miles east of the first one, yields ages of 143 m.y. for hornblende and 64 m.y. for biotite.

The apparent age of these three samples is compared with their location along an east-northeast–west-southwest line in figure 18. Although more data would be desirable, the graph at least suggests an approach to

TABLE 3.—*Potassium-argon data on plutonic rocks*

Sample No. (fig. 16)	Pluton or rock type	Mineral	K_2O (percent)	Radiogenic Ar^{40} (moles/g)	Atmospheric Ar^{40} (percent)	Age (m.y.)	Location
1	Flowery Trail Granodiorite	Hornblende	1.44	4.348×10^{-10}	11.2	194±7	NW¼ sec. 9, T. 32 N., R. 41 E.
		Biotite	8.71	1.283×10^{-9}	15.0	98±5	Do.
2	do	Hornblende	1.41	4.007×10^{-10}	9.4	183±6	Center sec. 3, T. 32 N., R. 41 E.
		Biotite	8.175	1.041×10^{-9}	22.5	84±3	Do.
3	do	Hornblende	1.39	3.042×10^{-10}	11.4	143±5	NW¼ sec. 1, T. 32 N., R. 41 E.
		Biotite	8.825	8.481×10^{-10}	69.9	64±3	Do.
4	Starvation Flat Quartz Monzonite	Hornblende	.6855	1.003×10^{-10}	20.1	97±3	SW cor. SW¼ sec. 3, T. 34 N., R. 40 E.
		Biotite	8.405	1.254×10^{-9}	7.8	98±3	Do.
5	Phillips Lake Granodiorite	Muscovite	10.765	1.092×10^{-9}	24.0	67±2	NE¼ sec. 34, T. 34 N., R. 41 E.
		Biotite	9.360	7.922×10^{-10}	10.0	56±2	Do.
6	do	Muscovite	10.64	1.356×10^{-9}	11.1	84±4	SW¼ sec. 8, T. 36 N., R. 42 E. (Colville 30-min quadrangle).
		Biotite	9.26	1.089×10^{-9}	9.5	79±3	Do.
7	do	Muscovite	10.665	9.244×10^{-10}	9.1	58±2	NE¼ sec. 12, T. 33 N., R. 41 E.
		Biotite	9.448	7.259×10^{-10}	11.4	52±2	Do.
8	Leucocratic dike	Muscovite	10.72	1.366×10^{-9}	11.1	84±2	NE¼ sec. 32, T. 35 N., R. 41 E.
		Biotite	7.895	6.996×10^{-10}	45.4	59±2	Do.
9	do	Muscovite	10.68	1.438×10^{-9}	13.8	89±6	SW¼ sec. 21, T. 35 N., R. 41 E. (Colville 30-min quadrangle).
		Biotite	8.695	9.568×10^{-10}	29.8	74±2	Do.
10	Silver Point Quartz Monzonite	Hornblende	.817	7.304×10^{-11}	32.7	60±2	NW¼ sec. 11, T. 29 N., R. 41 E.
		Biotite	8.70	6.502×10^{-10}	12.0	50±1	Do.
11	do	Hornblende	.6078	4.642×10^{-11}	37.5	51±5	NW¼ sec. 5, T. 30 N., R. 44 E. (Newport 30-min quad.).
		Biotite	8.45	6.016×10^{-10}	21.8	48±1	Do.
12	Granodiorite	Hornblende	.874	8.231×10^{-11}	26.6	63±2	NE¼ sec. 11, T. 29 N., R. 40 E.
		Biotite	8.88	6.493×10^{-10}	10.8	49±2	Do.
13	Fine-grained quartz monzonite	Hornblende	.338	1.745×10^{-10}	17.51	320±23	NW¼ sec. 32, T. 30 N., R. 41 E.
		Biotite	8.825	6.657×10^{-10}	11.8	50±2	Do.
14	do	Hornblende	.3555	1.671×10^{-10}	35.0	294±15	Do.
15	do	Hornblende	.374	1.136×10^{-10}	38.7	195±8	Do.
		Biotite	8.455	6.445×10^{-10}	48.3	51±2	Do.
16	Muscovite quartz monzonite	Muscovite	10.235	1.204×10^{-9}	8.0	78±2	NW¼ sec. 25, T. 30 N., R. 41 E.
17	Two-mica quartz monzonite	Muscovite	10.745	9.042×10^{-10}	21.7	56±2	SE¼ sec. 9, T. 31 N., R. 42 E.
		Biotite	9.12	7.568×10^{-10}	14.4	55±2	Do.
18	Aplite	Muscovite	11.00	1.033×10^{-9}	9.5	63±2	NE¼ sec. 34, T. 34 N., R. 41 E.

finding the true age of the Flowery Trail Granodiorite. The three biotites lie on a straight line which intersects the 50-m.y. level approximately 2 miles east of sample 3, while a smooth curve drawn through the three hornblende points intersects the 50-m.y. mark at about the same location. This construction is remarkably similar to the curves depicted by Hart (1964). The projected point of intersection may be the point at which an outcrop of the pluton causing the disturbance would be expected, or it may be the outer limit of complete argon loss in the rocks that the pluton intrudes. Extrapolated westward, the hornblende curve appears to flatten out, which may indicate that the "true age" is not much greater than the oldest age found for hornblende, 194 m.y.

These older hornblende ages do not appear to be due to excess argon in low-potassium minerals, because the hornblendes have a relatively high potassium content, averaging about 1.4 percent K_2O. Also, there is no correlation between higher apparent excess argon and lower potassium content, as might be expected in cases of excess argon; in fact, the potassium values for these three samples are remarkably similar (table 3).

Other plutons of early to middle Mesozoic age are known from northeastern Washington and southern British Columbia (Rinehart and Fox, 1972; Wanless and others, 1965, p. 14), although, like the Flowery Trail Granodiorite, each is associated with internal age discordances. Clark and Miller (1968) assigned the Flowery Trail Granodiorite a Mesozoic(?) age on the basis of geologic relations. This assignment is here refined to Late Triassic or Early Jurassic.

The next youngest pluton is the Starvation Flat Quartz Monzonite, which appears to be unaffected by the younger intrusive rock, at least where the dated sample was collected. Hornblende from a specimen collected on the west border of the area yields an age of 97 m.y., which agrees well with an age of 98 m.y. on

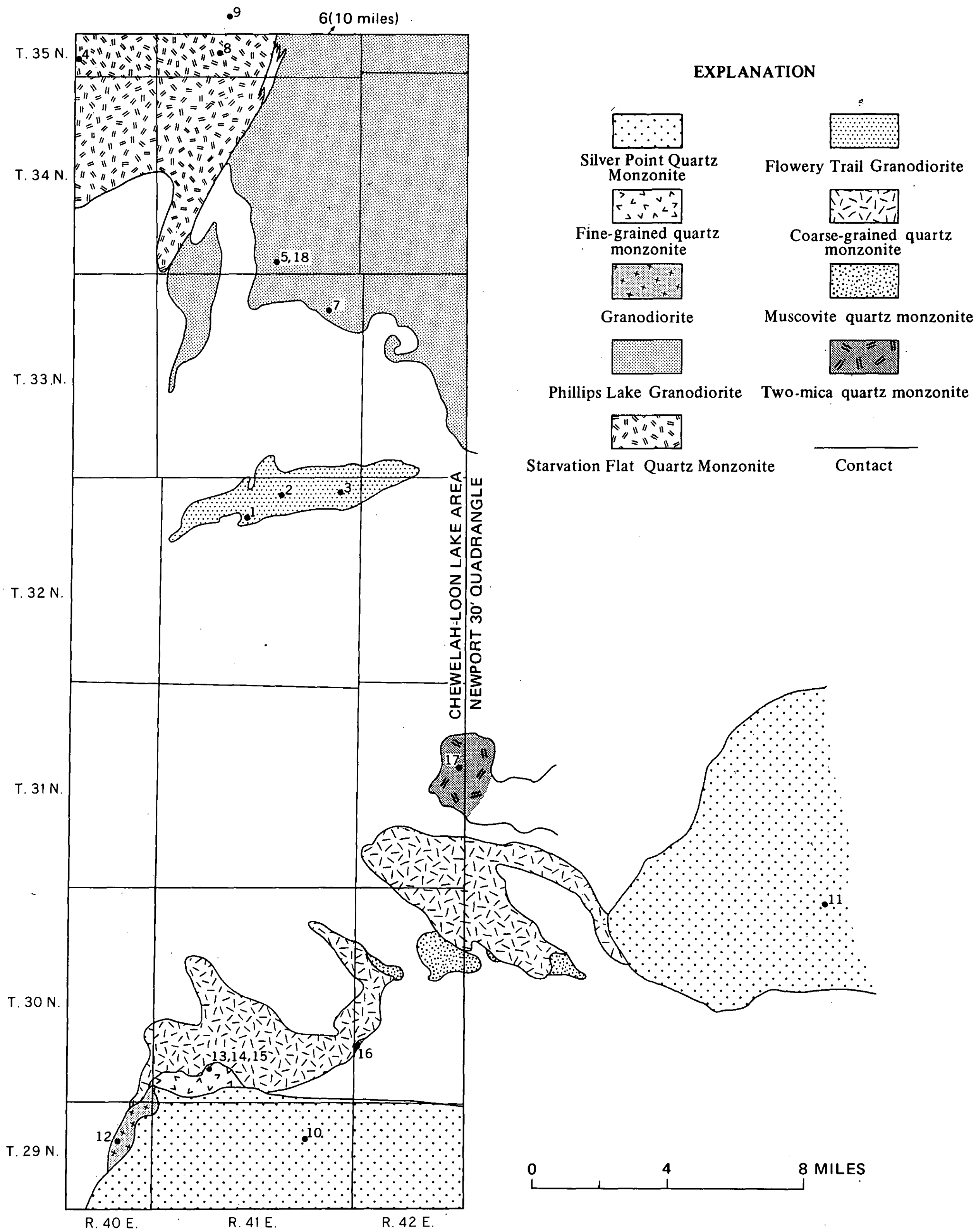


FIGURE 16.—Generalized map showing distribution of plutonic rocks in and around the Chewelah–Loon Lake area and the potassium-argon sample localities. Numbers correspond to sample numbers in table 3.

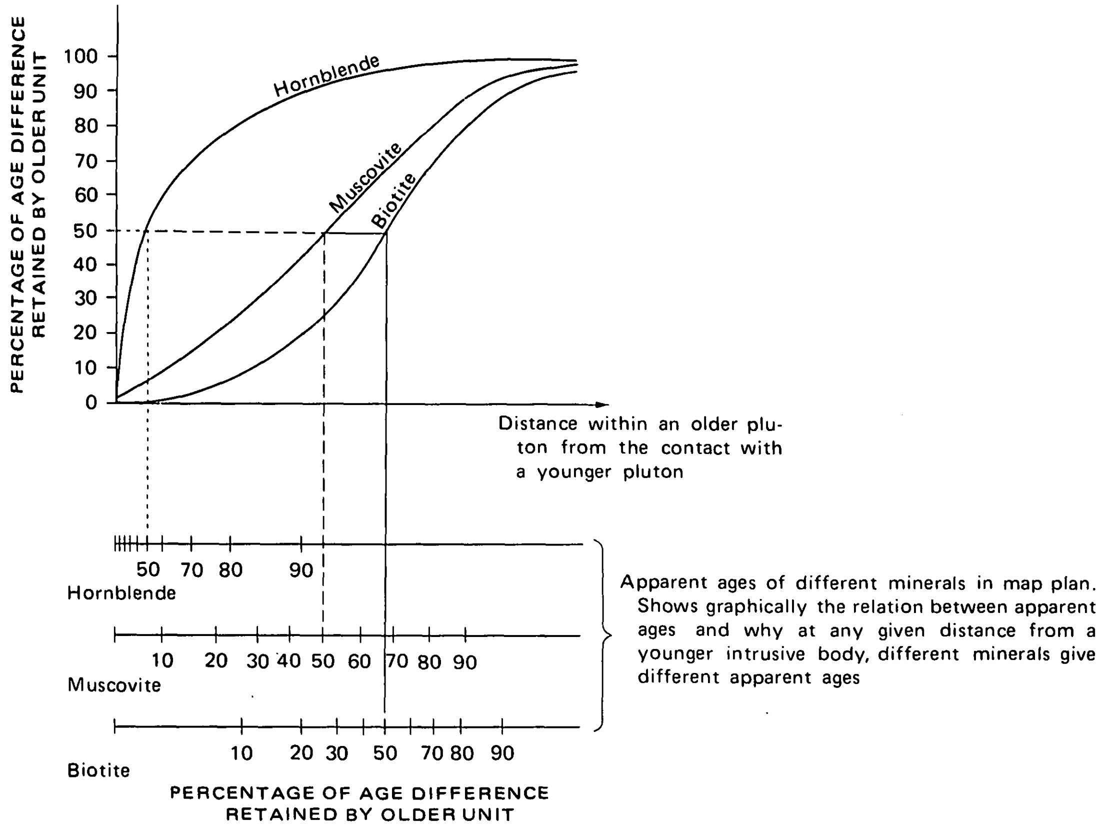


FIGURE 17.—Changes in apparent ages of hornblende, biotite, and muscovite in an older pluton intruded by a younger one.

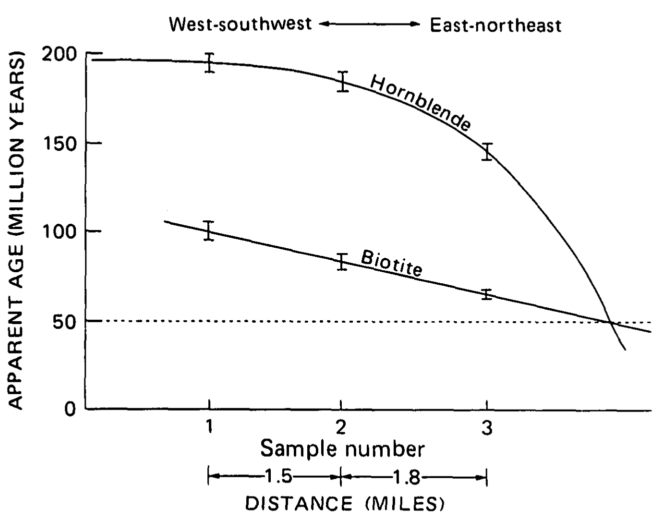


FIGURE 18.—Apparent ages of hornblende and biotite from the Flowery Trail Granodiorite plotted against the distances separating specimen localities.

biotite from the same specimen. The potassium-argon dates allow revision of the Mesozoic(?) age assigned by Clark and Miller (1968) to Cretaceous.

Muscovite-biotite pairs from two samples of the Phillips Lake Granodiorite collected within the area and one sample collected about 10 miles north of the area all yield discordant ages (table 9). Muscovite in the northernmost sample (No. 6) gives an age of 84 m.y., and biotite 79 m.y.; muscovite in the intermediate sample (No. 5), 15 miles south-southwest of sample 6, gives an age of 67 m.y., and biotite 56 m.y.; and muscovite in the southernmost sample (No. 7), 18 miles south-southwest of sample 6, gives an age of 58 m.y., and biotite 52 m.y.

Yates and Engels (1968, p. D245) reported similar but less pronounced discordances from probably correlative rocks in the Deep Creek area at a locality about 20 miles north of the report area. A rock which is modally, texturally, and mineralogically identical with the Phillips Lake Granodiorite was mapped as part of the Kaniksu batholith by Yates (1964) and gives a biotite age of 92±3 m.y. However, muscovite from a pegmatite associated with this rock gives an age of 99±3 m.y (Yates and Engels, 1968). About 2 miles north of these sample localities, the Spirit pluton, which Yates and Engels regarded as probably part of the Kaniksu batholith, gives a similar discordance (muscovite, 96±3 m.y.; hornblende, 94±3 m.y.; biotite, 91±3 m.y.), and biotite from a rock collected about 14 miles to the west gives a 100±3-m.y. age. They concluded that the biotite in the eastern samples lost part of its argon and that all these rocks were emplaced about 100 m.y. ago. If the correlation of the Phillips Lake Granodiorite with the Kaniksu rocks is correct, the probable age of the granodiorite is then 100 m.y., much older than any of the ages obtained so far for this rock in the report area.

Two of the leucocratic dikes that are associated with the Phillips Lake Granodiorite and cut it in places were also dated. A sample of the southernmost of the two (No. 8) was collected on the west flank of Blacktail Mountain where the dikes intrude the Starvation Flat Quartz Monzonite. Biotite from this rock gives a 59-m.y. age, and muscovite an 84-m.y. age. Sample 9, which was also from a dike intruding Starvation Flat Quartz Monzonite, was collected about 2 miles north of sample 8 and yields a biotite age of 74 m.y. and a muscovite age of 89 m.y. Since leucocratic dikes of this type cut the Phillips Lake Granodiorite, the maximum age obtained for the dike establishes a minimum age for the pluton. Thus, even though 84 m.y. is the oldest date obtained on the Phillips Lake body, the 89-m.y. apparent age of the dike rock more closely approaches the 100-m.y. age of the rocks to the north with which the pluton is believed to be correlative.

The relative discordances within the Phillips Lake Granodiorite and its associated leucocratic dikes point to the same area of disturbance as the Flowery Trail Granodiorite. A plot of the apparent muscovite and biotite ages of both the Phillips Lake Granodiorite and the associated leucocratic dikes versus distance between projected sample locations (fig. 19) shows a pattern similar to that predicted from figure 17.

A sample of the two-mica quartz monzonite northeast of Deer Lake gives a muscovite age of 56 m.y. and a biotite age of 55 m.y. Despite the seeming accordance, these ages are probably anomalous owing to the proximity of the samples to a younger plutonic event. This interpretation is supported by petrologic evidence that the two-mica quartz monzonite is the same as, or related to, the Phillips Lake Granodiorite.

Muscovite from a sample (No. 16) of the muscovite quartz monzonite collected 2 miles south-southeast of Deer Lake gives an age of 78 m.y. This sample was dated before it was known that the pre-Tertiary plutonic rocks had been disturbed by a later event. Because of the proximity of the sample locality to the apparent source of discordances noted in other units, the significance of the 78-m.y. date is questionable.

The coarse-grained quartz monzonite was not sampled for potassium-argon dating in this area, because it is highly susceptible to weathering and because the only datable mineral in the rock, biotite, is slightly

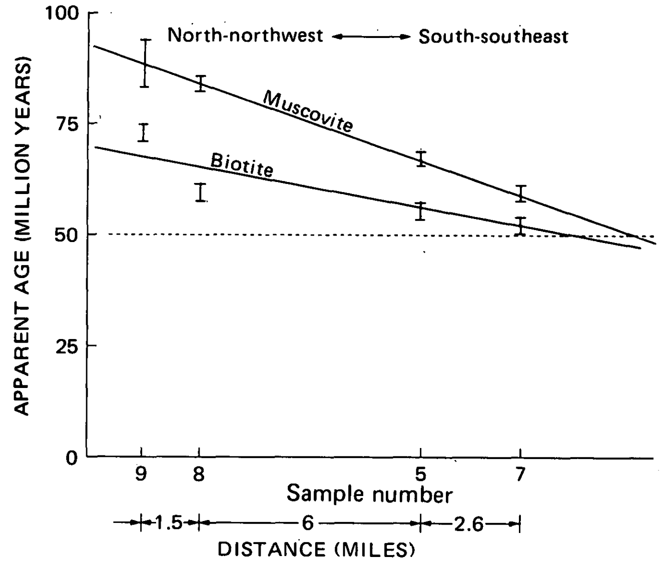


FIGURE 19.—Apparent ages of muscovite and biotite from the Phillips Lake Granodiorite and associated dikes plotted against the distances separating specimen localities.

chloritized. Even if a date that did not reflect the effects of alteration could be obtained, its significance would be difficult to evaluate because of the nearness of Tertiary plutonism. Field relations and spatial association, however, indicate that this unit and the muscovite quartz monzonite were probably emplaced during the same general event, but a little later than the Phillips Lake Granodiorite and its associated dikes and sills.

With the possible exception of the granodiorite and fine-grained quartz monzonite, the Silver Point Quartz Monzonite appears to be the youngest plutonic body in the area on the basis of both potassium-argon dating and crosscutting relationships found east of the area. Hornblende from a sample (No. 11) collected 15 miles east of Deer Lake in the Newport 30-minute quadrangle yields an age of 51 m.y., and biotite from the sample an age of 48 m.y. Biotite from a second sample from the Silver Point Quartz Monzonite (No. 10), collected near the southeast end of Loon Lake, yields an age of 50 m.y., whereas hornblende from this sample gives an age of 60 m.y.

A greater discordance was found in a sample from the small granodiorite pluton between Springdale and Loon Lake. Biotite from this sample yields an age of 49 m.y., and hornblende an age of 63 m.y. The most striking discordance, however, is from a sample of the fine-grained quartz monzonite (No. 13) collected between Springdale and Loon Lake. Biotite from this sample gives an age of 50 m.y., but the hornblende gives an age of 320 m.y. A reanalysis of a split of the hornblende from the same rock yields an age of 294 m.y. Another sample of the rock was collected and processed to check the large discordance, and a 195-m.y. age was obtained on the hornblende and a 51-m.y. age on the biotite.

Although potassium-argon dates appear to have been lowered by thermal events in many of the other rocks, the hornblende may give a misleading age for these three plutons. The granodiorite and the fine-grained quartz monzonite are thought to be genetically related to the Silver Point Quartz Monzonite on the basis of spatial associations and almost identical mineralogy. In addition, biotite from all three plutons gives an age of about 50 m.y., as does cogenetic hornblende from the Silver Point Quartz Monzonite east of the area.

Thin sections show that the hornblende in the granodiorite and in the fine-grained quartz monzonite, which yields ages in excess of 50 m.y., exhibits two characteristics that differ markedly from the hornblende in the sample of Silver Point Quartz Monzonite, collected east of the area. While some of the amphibole in the two smaller plutons occurs as isolated crystals, much of it is clustered together into clots of crystals. The other anomalous characteristic is the presence of pyroxene cores in many of the crystals. In much of the rock, the pyroxene has been partly altered to hornblende, and its optical properties are not normal for either mineral. Pyroxene cores and clotted hornblende have also been observed in thin sections of samples of the Silver Point collected near the southeast end of Loon Lake, but both characteristics are much more sparsely developed than in the granodiorite and fine-grained quartz monzonite. The hornblende clots may represent incompletely digested material picked up by the Tertiary plutons from either the Huckleberry Formation or the amphibolite sills in the Prichard Formation. (See "Fine-grained Quartz Monzonite, Internal Features and Petrology.")

Grain counts in immersion oils show that all three samples of hornblende from the fine-grained quartz monzonite consist mainly of pale-green crystals free of alteration and inclusions. The remainder of each sample consists of crystals that contain abundant opaque and semiopaque material and may represent incompletely altered pyroxene. Since all gradations between the two types are present, the counts are somewhat subjective, but a rough correlation between degree of purity and age does seem to exist (fig. 20 and following table). The higher proportion of crystals that appear to be altered in the samples yielding older hornblende ages suggests that these crystals may be the cause of the older ages owing to incomplete outgassing of argon.

The data are not conclusive but do suggest that the Silver Point Quartz Monzonite, along with its satellitic plutons, is about 50 m.y. old. Whether this pluton or others as yet unrecognized east of the area are responsi-

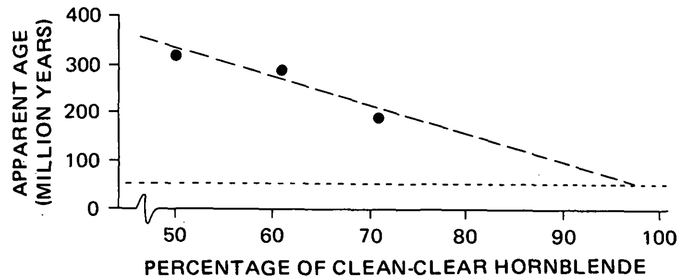


FIGURE 20.—Relation of apparent ages to concentration of clean-clear hornblende in the fine-grained quartz monzonite. Clean-clear taken to be all of clean + ½ intermediate.

Composition of hornblende separates used for potassium-argon dating

Apparent potassium-argon age of sample (m.y.)	*Clean hornblende*[1] *(percent)*	*Intermediate hornblende*[2] *(percent)*	*Impure hornblende*[3] *(percent)*
320	32	35	31 (also 2 percent biotite)
294	40	42	15 (also 3 percent biotite)
195	58	26	14 (also 2 percent biotite)

[1]Pale green and clear.
[2]Alteration noticeable; contains inclusions of other crystals; also contains fine-grained opaque material.
[3]Same as intermediate, but much more pronounced. Many crystals almost opaque owing to alteration and opaque minerals.

ble for the pronounced discordances shown by the older plutons is not yet known. Miller (1969) assigned the Silver Point Quartz Monzonite a Tertiary age, which is here referred to an Eocene age.

Hypabyssal dikes, some of which have petrologic affinities to the Silver Point Quartz Monzonite, are found throughout the report area and are especially abundant in the area just east of the Flowery Trail Granodiorite. Yates and Engels (1968, p. D246) reported ages of about 50 m.y. on lamprophyre dikes, shonkinitic sills and dikes, and volcanic rocks from the north half of the Colville 30-minute quadrangle. The lamprophyre dikes, in particular, strongly resemble some of the hypabyssal dikes in the report area and may be related to the same period of igneous activity as the Silver Point Quart Monzonite. If so, the intrusive event represented by these 50-m.y.-old rocks may be widespread in northeastern Washington.

The varied potassium-argon dates reported here can be explained by as few as three main periods of plutonic activity, if the dates are considered in conjunction with the petrographic and field relations of the plutonic rocks. The oldest, apparently involving only the emplacement of the Flowery Trail Granodiorite, occurred in Late Triassic or Early Jurassic time. The Starvation Flat Quartz Monzonite and the Phillips Lake Granodiorite and its associated dikes and sills were emplaced in mid-Cretaceous time. The absolute age of the muscovite quartz monzonite and coarse-grained quartz monzonite is uncertain. Field relations and spatial associations suggest that they were probably emplaced during the mid-Cretaceous event, but later than the Phillips Lake Granodiorite and its associated dikes and sills. Plutonic activity in the area apparently climaxed in the Eocene with the intrusion of the Silver Point Quartz Monzonite, the granodiorite, and the fine-grained quartz monzonite.

CENOZOIC HYPABYSSAL, VOLCANIC, AND SEDIMENTARY ROCKS

MAFIC DIKES

Fine-grained dikes or dike-form bodies, rich in mafic minerals, are scattered throughout the report area. Almost all cut across bedding, but some are locally concordant where they intrude Precambrian metasedimentary rocks. Although they could be called lamprophyres, perhaps sodium-rich minette or vogesite, they are referred to here simply as mafic dikes because they do not fit well into any of the lamprophyre categories.

Although the dikes do not form an appreciable percentage of the bedrock anywhere, they are obviously concentrated in some places. Three such places are in and around the Silver Point Quartz Monzonite, the Phillips Lake Granodiorite, and in the vicinity of Jay Gould Ridge. Dikes are numerous though widely scattered throughout the area underlain by the lower part of the Belt Supergroup, but almost none were found in the Deer Trail Group, Huckleberry Formation, or the Paleozoic rocks.

Most of the dikes are between 10 and 30 feet wide and too small to map. The largest of the few that were mapped are immediately south of Cottonwood Creek. One is just over 1 mile long, and the other, although not exposed along its entire length, is about 1.5 miles long. They average about 300 feet in width, but the shorter one bulges to about 1,000 feet near its north end. Another large dike, but one considerably different in appearance, crosses Jay Gould Ridge in sec. 9, T. 32 N., R. 41 E. It averages about 150 feet in width and is about half a mile long.

No preferred orientation common to even a majority of the dikes is evident on a regional scale. Their orientation appears to be controlled almost entirely by local structural and sedimentary features such as foliation, faults, joints, and bedding. The emplacement of the two large dikes south of Cottonwood Creek was obviously controlled by bedding in the host rocks. At least one dike, too small to be mapped, intrudes part of the fault in sec. 9, T. 33 N., R. 41 E. If the dikes were better exposed and more data available on the orientation of the smaller bodies, a regional trend might become apparent.

The dikes vary greatly in appearance, even though all are fine grained and most have approximately the same mineral composition. Three dikes coarse grained enough for modal analyses are of monzonite and quartz monzonite composition (see following table and fig. 21); most of the others are too fine grained for modal

Modes of three mafic dikes

Specimen No.	*2*	*8*	*9*
Plagioclase:			
Phenocrysts	19	20	12
Groundmass	25	23	28
Potassium feldspar	25	26	28
Biotite	9	6	14
Hornblende	11	11	7
Pyroxene	4	1	0
Quartz	6	10	10
Apatite	Trace	1	Trace
Opaque minerals	1	1	1
Calcite	Trace	1	Trace
Sphene	Trace	Trace	Trace
Total	100	100	100

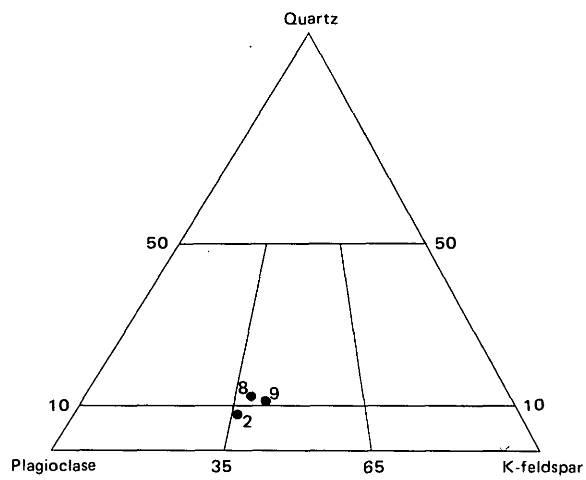


FIGURE 21.—Plots of mafic dike modes. Felsic minerals recalculated to 100 percent.

analyses. Almost all the dikes are unusually fresh and show little alteration. Because of this, they commonly crop out even where the host rock is completely covered by a thick layer of forest debris and soil.

Minerals found in the dikes are, in order of decreasing abundance, plagioclase, potassium feldspar, hornblende, quartz, biotite, magnetite (or ilmenite), apatite, sphene, chlorite, epidote, calcite, hematite, zircon, pyrite, allanite, and actinolite(?). Not all these are found in every dike. Table 4 gives the minerals of a representative group from throughout the report area.

Phenocrysts are generally euhedral or subhedral and consist of only plagioclase, potassium feldspar, quartz, biotite, hornblende, and pyroxene. Plagioclase phenocrysts are found in most of the dikes and average 0.12–0.2 inch in size. Their average composition is An_{34},

TABLE 4.—*Mineral composition of selected mafic dikes*

[X=mineral present; A=mineral present, but altered]

				Phenocrysts						Groundmass																
No.	Sec.	T. (N.)	R. (E.)	Plagioclase	Potassium feldspar	Quartz	Biotite	Hornblende	Pyroxene	Plagioclase	Potassium feldspar	Quartz	Biotite	Hornblende	Magnetite[1]	Apatite	Sphene	Zircon	Allanite	Epidote	Chlorite	Hematite	Pyrite	Carbonate	Actinolite(?)	Remarks
1	1	82	41	X			A	X		X	X	X	X		X	X										Aphanitic groundmass.
2	29	84	41	X			A	X		X	X	X	X	X	X	X	X	X			X	X		X		See mode (table, p. 58).
3[2]	29	84	41				A	X	A	X	X	X		X	X	X	X	X			X	X		X		Color index over 40.
4	9	82	41	X			A	X	A	X	X	X	X	X	X	X		X			X			X		Groundmass is high in potassium feldspar.
5	28	84	41	X			X			X	X	X	X		X	X		X			X			X		Even-grained large hornblende phenocrysts.
6	12	82	41	X			X	A		X	X	X	X	X	X	X	X				X			X		Groundmass is high in potassium feldspar.
7	18	88	41	X				X		X	X	X	X	X	X	X		X			X			X	X	
8	18	88	41				X	X		X	X	X	X		X	X	X	X			X	X	X	X	X	See mode (table, p. 58).
9	24	82	41	X			X	X		X	X	X	X		X	X	X	X		X	X			X		Do.
10	28	80	41	X	X	X	X			X	X	X	X		X	X		X			X					Gray aphanitic groundmass, pink potassium feldspar phenocrysts.
11	8	29	41	X	X	X		X		X	X	X		X	X	X	X	X		X	X					Low color index, large potassium feldspar phenocrysts.
12	8	29	41	X	X	X		X		X	X	X			X	X	X	X	X	X	X			X		
13	8	29	41	X	X	X		X		X	X	X		X	X	X	X	X	X	X				X		
14	9	29	41	X	X	X	X	A		X	X	X	X		X	X	X			X	X					Same as 10 in appearance.
15	17	29	41	X	X	X		X		X	X	X		X	X	X	X		X	X	X			X		Same as 11 in appearance, but no potassium feldspar phenocrysts.
16	1	29	40	X	X		A	X		X	X	X		X	X	X	X	X		X	X					Same as 11 in appearance.
17	82	80	41	X	X	X	X	A		X	X	X			X	X	X	X	X	X	X			X		Dark-gray even-grained phenocrysts are small.
18	16	29	41	X	X	X	A	X		X	X	X			X	X	X	X		X	X					Same as 11 in appearance, but no potassium feldspar phenocrysts.
19	1	29	40	X	X		A	X		X	X	X	A	X	X	X	X	X	X		X					Rounded potassium feldspar phenocrysts.
20	6	29	41	X		X	A	X		X	X	X	X	X	X	X	X	X			X					Large potassium feldspar phenocrysts. Strong resemblance to Silver Point Quartz Monzonite.
21	29	80	41	X			X			X	X	X			X	X		X	X		X	X		X		Dull-pink groundmass.
22	29	80	41	X			A	X		X	X	X	X	X	X	X	X	X		X	X			X		Same as 17 in appearance.
23	6	81	42					X		X	X	X		X	X	X	X				X					High color index.
24	8	81	42	X			A	X		X	X	X		X	X	X					X				X	
25	15	81	41	X			X			X	X	X	X		X	X					X			X		Dull-pink groundmass.
26	12	81	41	X			A	X	A	X	X	X		X	X	X	X	X		X	X			X		Pink groundmass.
27	15	81	41	X			A	X	A	X	X	X			X	X	X	X	X		X			X		Dull-pink groundmass.
28	15	81	41	X			A	X	A	X	X	X	X	X	X	X	X	X			X			X		Do.
29	8	81	42				A	X		X	X	X		X	X	X					X					Same as 23 in appearance.

[1]Probably includes ilmenite.
[2]Margin of same dike as No. 2.

whereas average groundmass composition is An_{15}. In addition, potassium feldspar phenocrysts as much as 0.8 inch long are found in dikes in and around the Silver Point Quartz Monzonite that may be related to that pluton. Hornblende, the most common mafic phenocryst, is found in about 90 percent of the dikes. Biotite is found in about 80 percent, but in many of these it is altered partly or completely to chlorite. In dikes that contain phenocrysts of both biotite and hornblende, one of the two minerals is always altered. In most, only one of the mafic minerals is found in the groundmass, and it corresponds to the unaltered mafic phenocryst. Almost all the biotite in the dikes associated with the Silver Point Quartz Monzonite is severely altered. These crystals may not have been stable in the physical-chemical conditions which produced the potassium feldspar phenocrysts found in all these rocks. Pyroxene phenocrysts are found only in the large dikes south of Cottonwood Creek and in the chilled margins of an unusually mafic dike about 2 miles west of Phillips Lake. The borders of all pyroxene crystals examined are altered to hornblende, biotite, or chlorite.

Groundmass grains are euhedral to anhedral and range from 0.001 to 0.005 inch in size. The variation in grain size is somewhat a function of the proximity to the borders, although chilled margins are not obvious in the field. Width of dike and grain size of groundmass are generally related.

The groundmass is some shade of gray in all dikes except the two large ones south of Cottonwood Creek. The darker dikes contain a higher concentration of mafic minerals in the groundmass than do the lighter ones. The color of the dikes south of Cottonwood Creek is a distinctive dull pink, which might be the result of an abnormally large amount of small potassium crystals in the groundmass. As with the feldspar-rich dikes associated with the Silver Point Quartz Monzonite, the biotite in all the pink rocks is highly altered.

The dikes are younger than any of the plutonic rocks. Engels used potassium-argon methods to date several dikes in the northeast quarter of the Colville 30-minute quadrangle (Yates, 1964). Two lamprophyre dikes which are similar to some of the mafic dikes in the report area and may be related to them were dated as 49.9±1.5 m.y. and 51.7±1.5 m.y. (Yates and Engels, 1968, p. D243).

ANDESITE

Relatively well exposed black andesite underlies an area of about a quarter of a square mile, 3 miles southwest of Chewelah. It is not found any other place in the report area, but somewhat similar rocks have been mapped in the Turtle Lake quadrangle to the southwest by Becraft and Weis (1963). They referred these rocks to the Gerome Andesite (Weaver, 1920) and assigned them to the Oligocene on the basis of plant fossils in interbedded tuffaceous rocks (Becraft and Weis, 1963, p. 37). The Gerome Andesite has also been mapped in the west half of the Hunters quadrangle, just 22 miles west of Valley, by Campbell and Raup (1964), and in the adjacent Wilmont Creek quadrangle by Becraft (1966). Becraft renamed the unit the Gerome Volcanics.

The lithologic descriptions given by Becraft and Weis partly fit the andesite in the report area, but they differ in some important aspects. The andesite in the Turtle Lake quadrangle is higher in SiO_2 and Al_2O_3 but is noticeably lower in FeO and MgO than the andesite described here. Other constituents are about the same.

The andesite appears to rest on the Stensgar Dolomite, but the exposures do not show whether the contact is extrusive or intrusive. The top is not preserved, and no flow structures were observed. If the andesite is extrusive, and assumed to be near horizontal, it must be at least 240 feet thick.

Phenocrysts of olivine, hornblende, and, to a lesser degree, biotite are easily seen in hand specimen. Plagioclase and pyroxene crystals larger than the groundmass, but smaller than the phenocrysts, are abundant although obvious only in thin section. The mode in the table that follows is representative of the andesite. Hornblende is the most abundant mafic mineral. Almost all of it is highly oxidized and has thick rims of magnetite or ilmenite finely disseminated in a semiopaque material. Inside the alteration rim the remaining hornblende is slightly altered and highly zoned. It is probably basaltic hornblende because of the oxidation products, medium to high birefringence, and pleochromism, X=light tan, Y=yellow brown, and Z=deep golden brown. Olivine crystals are as much as 0.08 inch in size and are surrounded by fine-grained olive-green to golden alteration products that have moderate birefringence. The olivine has a $2V_x$ of about 80°–90°. Biotite is much less abundant than the hornblende but is surrounded by the same rims of fine-grained opaque minerals. Its pleochroism is X=light yellow brown and Y=Z=red brown.

Chemical analysis, CIPW norm, and mode of andesite southwest of Chewelah

[Analysts: P. L. D. Elmore, Hezekiah Smith, James Kelsey, Lowell Artis, and James Glenn]

Chemical analysis (weight percent)		CIPW norm (weight percent)		Mode (volume percent)	
SiO_2	58.3	Q	8.0	Plagioclase	8
Al_2O_3	14.7	or	19.0	Hornblende	20
Fe_2O_3	2.5	ab	25.6	Pyroxene	3
FeO	3.8	an	17.0	Olivine	1
MgO	5.8	wo	3.3	Biotite	1
CaO	5.8	en	14.5	Groundmass	67
Na_2O	3.0	fs	3.7		100
K_2O	3.2	mt	3.6		
H_2O+	1.1	il	1.7		
TiO_2	.89	ap	1.2		
P_2O_5	.51	cc	.2		
MnO	.11		99.7		
CO_2	.10				
	99.61				

Pyroxene is the second most abundant mafic mineral but is not easily identifiable in hand specimen. It ranges in size from about 0.12 inch down to microlites in the groundmass and is the only mafic mineral that appears to be relatively free of any alteration. The $2V_z$ is about 45°–50° in some crystals, but about 0°–25° in others. At least some of the pyroxene appears to be pigeonite, and some augite. Plagioclase ranges in size from about 0.12 inch down to microlites in the groundmass. It appears to be andesine but is highly zoned.

The finer grained plagioclase and pyroxene, along with opaque minerals and partially devitrified brown glass, form an aphanitic groundmass with a pilotaxitic texture. Much of the very fine grained material may be devitrified glass. Although it is too fine grained to identify in thin section and has not been X-rayed, its alkali content (see preceding table) suggests that much of it is quartz, potassium feldspar, and albite.

BASALT

About 10 square miles in the southwest part of the area is underlain by basalt, but more than half of this is masked by younger glacial and alluvial debris. The basalt appears to be confined chiefly to elevations below 2,500 feet, except on the northwest slope of Jumpoff Joe Mountain, where it is found as high as 2,900 feet. Individual flows and flow thicknesses could not be discerned, and so the attitude of the flows could not be measured directly. Since the base is found at about the same elevation in most places in the area, the flows must be nearly horizontal, except around irregularities in the underlying surface. In the NE¼ sec. 30, T. 31 N., R. 41 E., Empire Explorations Inc. drilled a test well which penetrated 139 feet of basalt. Assuming that the basalt is nearly horizontal, this is the most accurate measurement of a partial section but probably does not represent a maximum thickness in the area.

The rock is black to dark gray, nonporphyritic, and locally vesicular. Columnar jointing is well developed in places but is not obvious because exposures are poor. Petrographically, the rock has a hyalo-ophitic texture. Plagioclase, pyroxene, and olivine crystals generally constitute about 45 percent of the rock. Dark-brown almost opaque glass with abundant disseminated magnetite and (or) ilmenite makes up the rest of the rock. Several thin sections of rocks from different localities showed little variation in mineralogy or texture. The only notable variation was found in a specimen collected in the NW¼ sec. 17, T. 31 N., R. 41 E., which contained about a third more plagioclase and pyroxene than other specimens and about a third less glass and olivine. The following table gives the modes, in volume percent, for specimens of this basalt:

	Specimens from three localities[1]	*One specimen from sec. 17, T. 31 N., R. 41 E.*
Plagioclase	26	35
Pyroxene	13	17
Olivine	6	3
Glass	45	35
Opaque minerals	10	10
Total	100	100

[1]Average of three modes; less than 3 percent variation in any constituent.

Weaver (1920, p. 99) and Jones (1929, p. 50) applied the name Camas Basalt to all the Tertiary volcanic rocks in the report area but recognized that the basalt is largely continuous with the Columbia River basalts to the south. They erroneously included the andesite 3 miles southwest of Chewelah with the basalt, however. Griggs (1966) mapped basalt of the Columbia River Group within 2 miles of the southwest corner of the report area. Becraft and Weis (1963, p. 39–40) reported at least 900 feet of Columbia River Basalt in the Turtle Lake quadrangle, at least part of which appears to be "Late Yakima flows." D. A. Swanson, of the U.S. Geological Survey, examined several thin sections of basalt from the report area and reported that they strongly resemble Yakima Basalt, which is of Miocene and Pliocene age (oral commun., 1966).

LATAH FORMATION

Hosterman (1969, p. 56) assigned the clay-bearing strata found in two small claypits in sec. 4, T. 29 N., R. 42 E., and sec. 32, T. 30 N., R. 42 E., to the Latah Formation of Miocene age. In an earlier report, Miller (1969, p. 4) erroneously included these clay beds in the Quaternary.

The formation was originally named by Pardee and Bryan (1926, p. 1), who thought that the unit was overlain by the Columbia River Basalt. Kirkham and Johnson (1929, p. 483) showed that the Latah Formation actually interfingers with the basalt. The contact between the basalt and Latah Formation is not exposed in the report area.

In the two claypits the formation consists of sandstone, siltstone, silty clay, and what Glover (1941, p. 282) described as "bog iron." The so-called bog iron consists of rust-colored cohesive material which resembles limonite and which cements clastic material ranging in grain size from coarse sand to clay. The surface of the bog iron is botryoidal in many places. The bog iron may be confined to a single bed at both pits and if so may prove an aid to prospecting for clay. (See Miller, 1969, p. 6 and fig. 3.) The clay is unconformably overlain by glacial material.

CONGLOMERATE

Small isolated patches of well-indurated unsorted pebble to boulder conglomerate are found on the hill southwest of Chewelah and on the south and east flanks of Cliff Ridge. At both localities, the clasts are composed of the unit on which the conglomerate is lying or of nearby lithologies.

Southwest of Chewelah, the conglomerate appears to cling to the side of the hill. Stratification is not obvious in the rock, and so it is not known if this is a buttress relationship. At Cliff Ridge, the conglomerate is unconformably overlain by glacial material. Many of the clasts in the conglomerate southwest of Chewelah are fairly well rounded, but most of those in the unit at Cliff Ridge are angular.

The age of the rock is not known. The conglomerate at one locality may not even be the same age as at the other. The unconformity at the base of the conglomerate and the fact that the rock is well indurated lead us to suspect a Tertiary age.

GLACIAL, ALLUVIAL, AND TALUS DEPOSITS, UNDIFFERENTIATED

Glacial, alluvial, and talus deposits were mapped as one unit. Alluvial material is confined to the beds and flood plains of modern streams. Glacial debris consists chiefly of sand and gravel and locally fine-grained lacustrine deposits. The debris forms moraines, outwash plains, terraces, and thin mantles on hillsides.

The distribution of glacial erratics, striae, and moraines suggests that ice covered the western two-thirds of the report area and that a lobe moved eastward from the main body in Burnt Valley. The presence of glacial debris on top of Quartzite Mountain led Clark and Miller (1968, p. 3) to surmise that the glacial ice in the Colville River valley must have been more than 2,000 feet thick. Chewelah Mountain does not appear to be glaciated above the 5,000-foot elevation, although bedrock is deeply frost riven. Similarly, that part of Calispell Peak above this elevation must have protruded above the continental glaciers, but two cirques on the north side of Calispell Peak indicate local mountain glaciation, presumably during the waning phase of glaciation.

Moraines are common in the area, but many have been eroded almost beyond recognition. A prominent moraine is responsible for the sharp switchback in the road in the NE¼ sec. 24, T. 33 N., R. 41 E. Additional small moraines are found at about the 3,600-foot level on the slopes north of Burnt Valley.

The erosionally resistant north-south belt of Addy Quartzite, Striped Peak Formation, and upper Wallace Formation appears to have dammed westward-flowing glacial streams and caused deposition of large volumes of glacial debris in basins east of it. The largest of these basins is now drained by Cottonwood Creek. Others to the north are now drained by Sherwood Creek, Thomason Creek, South Fork of Chewelah Creek, North Fork of Chewelah Creek, and Bear Creek.

Poorly indurated lacustrine materials, probably deposited in ponded water behind ice dams or moraines, are found at several places in the area. An excellent example is exposed in small roadcuts at the top of the divide which lies on the line between secs. 15 and 16, T. 31 N., R. 41 E. The sediments there are fine grained, light gray to white, and thin bedded. Almost none of the lacustrine deposits crop out in natural exposures. Although they contain some clay minerals, these deposits do not have the high proportion of clay minerals found in the lake sediments previously described by Miller (1969, p. 4).

Talus deposits are found chiefly below bold outcrops, formed mainly by the quartzite units.

STRUCTURE

Structural and stratigraphic interpretations are closely interrelated in the report area because interpretations involving one are almost invariably based on the other. Much of the structure shown on the map and cross sections is inferred or at best based in part on indirect evidence because of the generally poor exposure. This is particularly true for much of western part of the area, where the structural complexity is greatest. To a certain extent, the degree of interpretive freedom is inversely proportional to the amount and quality of exposure. During the mapping, as stratigraphic information increased and interpretations became more refined, the complexity of the structural history became increasingly apparent. Because the amount of interpretation is so great and the structure so complex, the authors have tried their best to distinguish observed from interpreted structures in the following section and, as much as possible, to point out the basis for interpretations.

Two distinct structural blocks are evident in the report area. (See fig. 8) The eastern block, which underlies most of the area, consists largely of Belt rocks that form the west limb of a large anticline. Numerous faults cut this fold, and the northern part of the fold is overturned toward the west. The western block is underlain primarily by the Deer Trail Group, which here forms the north end of the magnesite belt. Significant differences in stratigraphy and structures within the blocks suggest that a major fault separates them and that a large but unknown amount of lateral movement has occurred. The Jumpoff Joe fault may be this fault, but this possibly cannot be demonstrated conclusively. (See section "Structure Separating the Two Blocks.") At the few localities where the fault is exposed, it appears to dip shallowly to the west and places Precambrian Deer Trail rocks on the west side against Paleozoic rocks on the east side. Most of the fault, however, is concealed beneath the surficial debris filling the Colville Valley, and so it is not known for certain whether the various segments mapped as the Jumpoff Joe fault are in fact parts of a single, continuous fault.

STRUCTURES IN THE BELT SUPERGROUP BLOCK

FOLDS

The northerly trending west limb of a large anticline is evident from the outcrop pattern and attitudes of Belt rocks in the east-central part of the report area. (See pls. 1, 2; figs. 8, 22). Southeast of Deer Lake the lower part of the Belt Supergroup strikes northwest and dips at moderate angles to the west. North of Deer Lake the strike of the section swings progressively more northward, and the dip becomes increasingly steeper. At about the latitude of Grouse Creek, the section strikes north-northeast and is overturned towards the west. Northward it becomes progressively more overturned, but the strike remains fairly constant up to the latitude of Quartzite Mountain. There, it begins to swing farther eastward until in the vicinity of Goddards Peak the section strikes east-northeast, and locally east-west, and is then cut off by plutonic rocks.

The east limb of this anticline, east of the report area, is not nearly as well defined. Attitudes of the Belt rocks between Fan and Horseshoe Lakes (fig. 2), about 7 miles east of Deer Lake, indicate that the rocks are near the axis of the fold, as the strike swings to east-west, and in places east-northeast. In this area all the strata are right-side-up and dip gently to the south. If the strike of the section continues to swing to the northeast, the strata will form the east limb of the anticline. However, plutonic rocks may interrupt this continuation.

This anticline, here named the Nelson Peak anticline, and the partially exposed syncline to the west, here named the Chewelah syncline, are shown in figure 22. They are the westernmost in a series of rather evenly spaced generally north-south-trending folds extending eastward almost to Pend Oreille Lake in Idaho. Barnes (1965, pl. 1) mapped two synclines, the Priest River and the Peewee, in the Priest River valley in Idaho. The two folds project toward one another and may be a single structure. About 5 miles west of them, he mapped an anticline which Schroeder (1952, p. 26) had previously named the Snow Valley anticline. This anticline is en echelon with an unnamed anticline that Barnes mapped near the southwest end of Priest Lake, but is separated from it by about 8 miles of intervening younger granodiorite. All the folds are rather open, and the strike of the axes varies somewhat. The general strike, however, is about due north.

In the Bead Lake area, Schroeder (1952, p. 25) mapped only the east limb of a fold that he called the Newport syncline and suggested that its axis lay to the west, approximately in the position of the Pend Oreille River. From the strike of the beds of the east limb, the

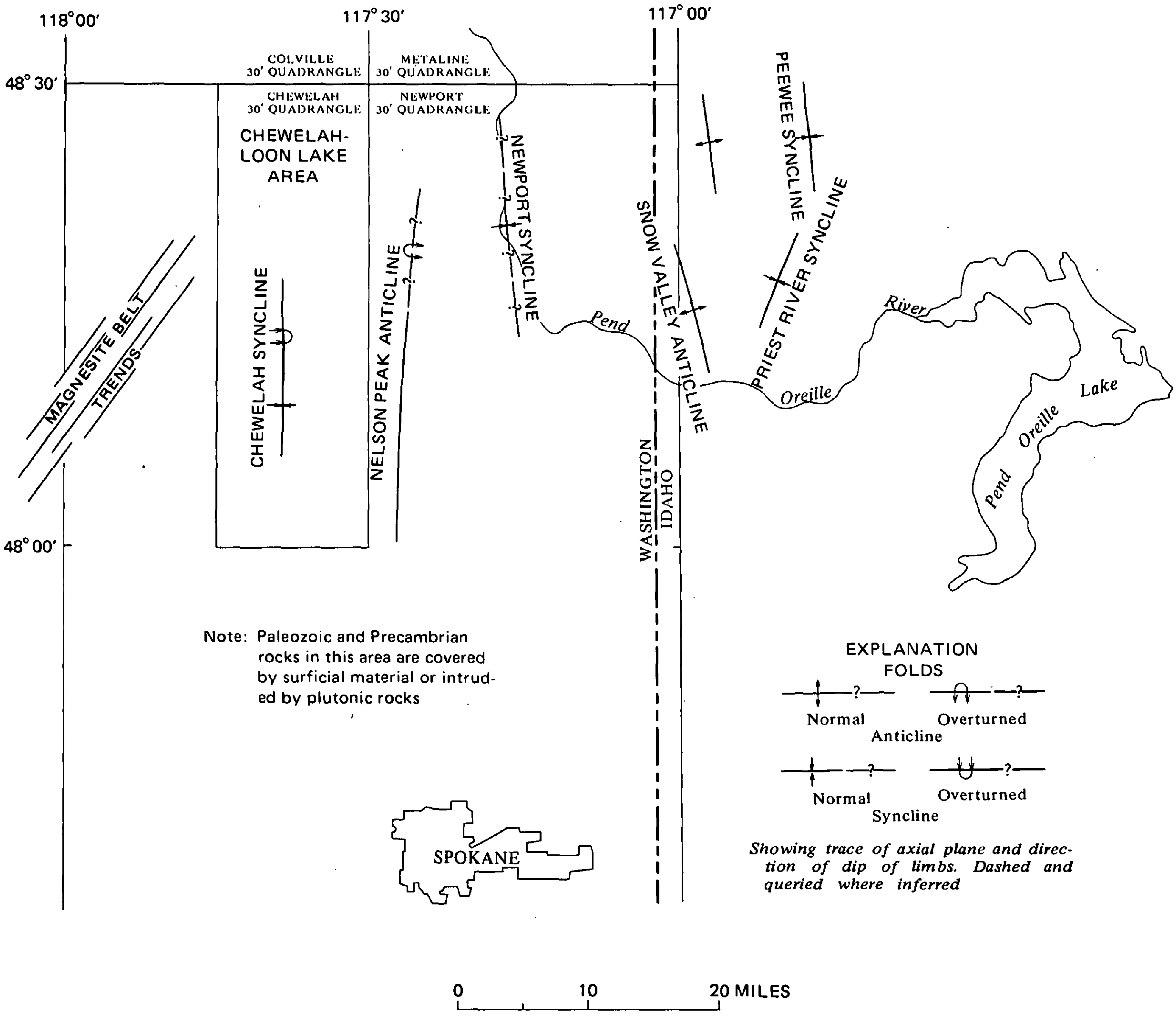


FIGURE 22.—Location of major fold axes in the Belt Supergroup block.

axis should have about the same strike as the Snow Valley anticline.

The Priest River syncline, Snow Valley anticline, and Newport syncline are all rather open folds with gentle to moderately dipping flanks. The Nelson Peak anticline and Chewelah syncline are noticeably tighter than these three and are in part strongly overturned. There is an equally noticeable increase in the intensity of folding from south to north within the report area. Both limbs of the Nelson Peak anticline at about the latitude of Deer Lake are normal, but to the north, at the latitude of Chewelah Mountain, the west limb is strongly overturned to the west. The Chewelah syncline, although ill defined and complicated by faulting in the southern part of the area, is reasonably well developed from the Jay Gould Ridge area north to Johnson Mountain. Comparison of cross sections *B–B′* and *A–A′* (pl. 1) shows the change between these two areas. Although strongly overturned, the syncline at the latitude of *B–B′* is rather regular. Along *A–A′* however, the limbs of the fold are attenuated, and the nose is considerably thickened. The structural change in thickness is due at least in part to the extreme development of a closely spaced slip cleavage in this area. The cleavage appears to be about parallel to the axial plane of the fold and is so pervasive that it has given the rocks a highly sheared appearance. In the vicinity of *B–B′* the slip cleavage is relatively low dipping and roughly parallel to the axial plane of the syncline but is not nearly as well developed as it is to the north. The right-side-up part of the syncline is exposed on Johnson Mountain, in Ten Mile Creek southeast of the Flowery Trail Granodiorite, and in the vicinity of Deer Lake.

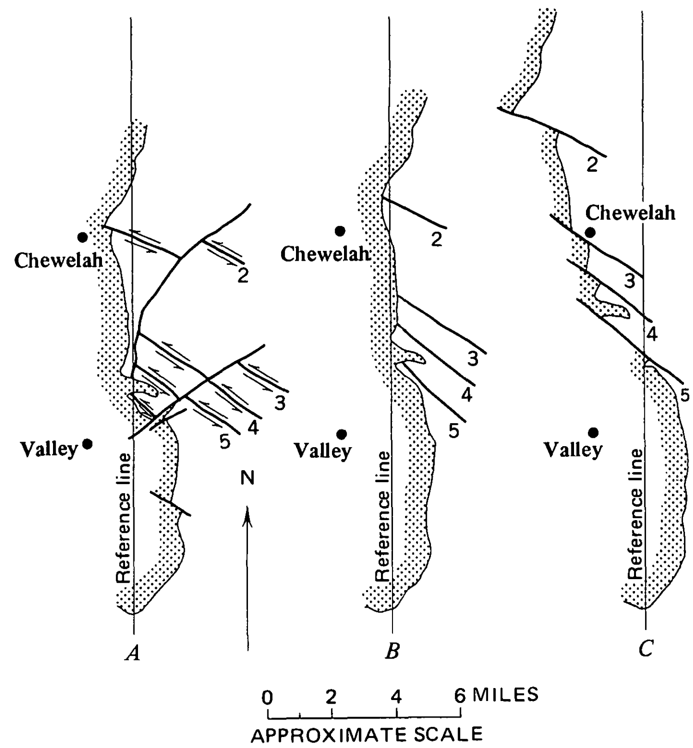


FIGURE 23.—Sketch maps showing alternate interpretations of the relation of the Addy Quartzite to the northwest-trending faults between Eagle Mountain and Jumpoff Joe Mountain. *A*, Present mapped relation. *B*, Lower contact of Addy Quartzite restored to position undisturbed by faults. Inferred northeast-striking fault has been removed. *C*, Probable configuration of lower contact of Addy Quartzite if movement on northwest faults had all been post-Addy. Arrows indicate apparent offsets only.

FAULTS

A group of northwest-trending faults offset the Belt Supergroup rocks with rather large apparent displacement, all but one in a left-lateral sense. Parts of the traces of all of them are masked by glacial debris, and so their position can only be inferred in many places. The apparent movement along these faults could have been strike slip or dip slip, probably the latter, and is thought to be Precambrian in age. These faults are considered to predate the development of the major folds in the Belt Series, as they do not appear to cut the Cambrian Addy Quartzite, which is involved in the folding (fig. 23). The fault east of Quartzite Mountain (No. 2, pl. 1) has a minimum apparent displacement of 10,000 feet in an apparent left-lateral sense. At the west end the displacement of the lower contact of the Addy Quartzite is only about 1,000 feet and is interpreted as renewed post-Cambrian movement along a Precambrian fault. Faults 3 and 4 do not cut the Cambrian quartzite, although they appear to be truncated by a north-northeast-trending, perhaps roughly contemporaneous, fault before they reach the quartzite. Fault 5, which has a minimum apparent offset of about 11,500 feet, passes through an area covered by glacial drift, across which the Addy Quartzite has a roughly comparable offset. The fault relations in this covered area are largely interpretive and are complicated by a well-documented large northeast-trending fault and a small thrust fault, the existence of which is based largely on interpretation. Most of the apparent left-lateral offset of the Addy Quartzite in this area is probably caused by dip slip on the large northeast-trending fault just mentioned. This fault clearly offsets the lower contact of the Addy Quartzite and also offsets the inferred contact between the quartzite and overlying carbonate rocks south of where it cuts fault 5.

The trace of fault 6 is covered for its entire length, but the existence of the fault is inferred in order to explain the juxtaposition of unlike Belt units to the north and south of it. Because of the dips of the strata involved, a prohibitively large dip slip would be required to bring the rocks here from their original location.

Although broken by other faults, fault 6 appears to strike northwest. The best reconstruction of premovement paleogeology necessitates about 7 miles of right-lateral slip along this fault. However, this distance may bear little resemblance to the actual amount of movement along the fault because of complications by other faults in the immediate area and uncertainties in correlation of offset features across the fault. The best "match" across the fault appears to be the syncline on Deer Lake Mountain with a similar syncline on the south flank of Blue Grouse Mountain. Both folds have the same axial strike, are rather open and involve the same stratigraphic units. If they are offset parts of the same structure and are related to the Nelson Peak anticline and Chewelah syncline, then fault 6 must have formed after the folding and is much younger than faults 1 through 5 are considered to be.

Faults with approximately the same northwest strike as fault 6, and with right-lateral offset as well, are abundant and well documented in the Coeur d'Alene district and in the Clark Fork quadrangle. The two largest and most widely known of these are the Hope fault in the Clark Fork and Libby areas, and the Osburn fault in the Coeur d'Alene district. Harrison and Jobin (1963, p. K28) reported 16 miles of apparent right-lateral offset on the Hope fault but estimated that some of this is due to vertical movements. They suggested that the major strike-slip movement took place in the Precambrian and that the dip-slip movement occurred in Laramide or post-Laramide to pre-Pleistocene time.

Hobbs and his colleagues estimated as much as 16 miles of right-lateral strike-slip movement along one segment of the Osburn fault. Detailed structural analysis suggests that the major strike-slip movements along this fault occurred after the intrusion of the 100-m.y.-old monzonitic stocks in the vicinity but before the extrusion of the Miocene Columbia River basalts (Hobbs and others, 1965, pl. 10, p. 128). Even though the major movement along this fault is probably Cretaceous, a zone of weakness approximately coincidental with the present position of the fault may have existed there much earlier (Hobbs and others, 1965).

The probable strike-slip movement along fault 6 in the report area may have occurred during the Cretaceous, when the Osburn fault was most active.

If the northwest-trending faults are folded, as the cross sections indicate, their trace on plates 1 and 2 should be sinuous rather than straight, especially where they pass from a slightly overturned to a more tightly overturned part of the section and where they cross areas of rugged topography. The faults probably do have sinuous trace, but exposure is simply not adequate to map actual fault traces. Locally, offset features are exposed well enough that short segments of a fault can be drawn. Most of the segments whose strike can be determined with some reliability trend northwest. Gross offsets such as those east of Quartzite Mountain and in the vicinity of Cottonwood Creek are obvious, but actual fault traces could be quite irregular there as exposure is poor even in areas shown as bedrock. To minimize interpretive prejudice, separated segments along a fault have been connected by relatively straight lines through areas of sparse outcrop or glacial cover, even though the trace may actually be quite irregular.

Previously published cross sections (Miller, 1969) show the faults unfolded. Neither interpretation can be conclusively proved or disproved, but as the Addy Quartzite is involved in the folding, the faults, which apparently predate the quartzite, also must be folded.

Differential movement along preexisting faults during folding, in addition to the later small-scale thrusting, has created a locally complicated structural knot where the Cottonwood Road intersects the Addy Quartzite (sec. 5, T. 31 N., R. 41 E.). Much of the rock in this small area is highly crushed, broken, or tightly folded. If exposures were more complete, detailed study here could help to better define the relations between the various structures.

STRUCTURES IN THE DEER TRAIL GROUP BLOCK

The Deer Trail Group block and most of the structures within it strike almost unvaryingly northeast-southwest. Beds, faults, and, to a lesser degree, folds diverge little from this trend. The trend is not well displayed in the relatively limited outcrops of the Deer Trail Group in the report area but is excellently developed southwest and northeast of the area. North-south and northwest-southeast trends within the Belt Supergroup block appear to be truncated by the Deer Trail Group. The differences in trend and the truncation distinguish one block from the other. No large structures in the Deer Trail Group block strike northwest-southeast. A few faults strike northwest-southeast, but none have large displacements or great extent (Campbell and Loofbourow, 1962, pl. 1).

The rocks north of Chewelah form the northeast end of the magnesite belt. Most of the rocks here are covered by glacial and alluvial material, but the northeast-southwest trends that characterize the magnesite belt are better developed here than elsewhere in the report area. Because exposure is so sparse in this part of the report area, a small amount of reconnaissance mapping was done just outside the west boundary, in the Iron Mountains, where the rocks are fairly well exposed. Structure and stratigraphy can be projected from there into the report area. A series of four northeast-southwest-trending moderate-angle reverse faults, which project into the area, are well developed in the Iron Mountains. Two more, not well defined in the Iron Mountains, are found in the report area. Although they appear to be steeper, almost all the faults that Campbell and Loofbourow (1962, pl. 1) mapped in the magnesite belt are reverse faults and have a similar strike and trend.

In the north half of the magnesite belt, Campbell and Loofbourow showed an inferred northwest-southeast fault folded by a locally developed north-south-striking anticline and syncline. Another anticline and another syncline with the same axial strike are found on Gold Hill and Deer Mountain 3 miles northwest of Chewelah. There the folds appear to involve the Addy Quartzite and are therefore post-Cambrian in age. Campbell and Loofbourow suggest the sequence of structural events in the magnesite belt began with development of the northeast-southwest faults, followed by local folding with subsequent development of the few northwest-southeast faults. Presumably the northeast-southwest strike of the entire magnesite belt was established during or before development of the faults with that strike.

OTHER DIFFERENCES BETWEEN THE TWO BLOCKS

The primary lithologic differences between the Deer Trail Group and the upper part of the Belt Supergroup are exaggerated by differences in degree of dynamic metamorphism. Throughout most of the magnesite belt, the argillites of the Deer Trail Group are to some degree phyllitic and (or) slaty. This is especially true of the Togo Formation and the McHale Slate. The Edna Dolomite contains argillite zones as much as 600 feet thick and also bedding-plane partings of argillite. Almost all the argillite is phyllitic; in places it even imparts a phyllitic look to the dolomite. From reconnaissance in the magnesite belt, the authors gained the impression that the Deer Trail Group is noticeably and consistently more phyllitic than the Belt Supergroup east of the Colville Valley. Deer Trail Group rocks are relatively undeformed only in isolated pockets —for example, the Stensgar Dolomite 4 miles northwest of Chewelah on the north side of U.S. 395 and the argillite in the SW cor. sec. 22, T. 33 N., R. 40 E. (fig. 6). The McHale Slate within 20 feet of the Stensgar Dolomite at the locality just mentioned, however, is converted almost to a phyllonite, as is argillite within 100 feet of the undeformed argillite in sec. 22. In contrast, the Belt Supergroup rocks east of the Colville Valley are only locally phyllitic, and where they are it is due to the extreme development of a slip cleavage, particularly in the northern part of the area. In general, however, the difference in dynamic metamorphism between the two sections is real and noticeable.

In addition to structural and stratigraphic contrasts, a significantly different group of rocks overlie the Deer Trail Group and Belt Supergroup blocks. The Addy Quartzite rests on the Huckleberry and Monk Formations in the Deer Trail Group block, whereas it directly overlies the Belt Supergroup in the other block.

There are only two localities in the report area where the Huckleberry or Monk Formations possibly overlie Belt rocks. In sec. 7, T. 33 N., R. 41 E., and in sec. 29, T. 34 N., R. 41 E., small areas are underlain by greenstone and amphibolite injected by many leucocratic dikes. These localities are just east of what may be the main structure separating the Deer Trail and Belt blocks at this latitude. However, another fault which also may be the main break passes just east of the greenstone and amphibolite.

Across the international boundary, and on strike with the Priest River Group, Rice (1941, map 603A) showed a thick section of the Purcell Series (the Canadian correlative of the Belt Supergroup) overlain by the Irene Volcanics and Toby Conglomerate (the Canadian correlative of the Huckleberry Formation). Although the descriptions by Rice (1941, p. 8–13) fit the Belt equivalent Purcell Series much better than the Priest River Group, there is some doubt as to whether the specific section in the southwest corner of the map area fits the descriptions. Furthermore, it is not known if the structure separating the Deer Trail and Belt blocks in the report area extends this far north.

Along its strike length the Huckleberry Formation is intermittently continuous for at least 90 miles. Near the center of the magnesite belt, it is 4,500 feet thick (Campbell and Loofbourow, 1962, p. F22, F23). Only 10 miles east, however, across the strike of the magnesite belt, it is not found between the Cambrian and Belt rocks. The basin in which the Huckleberry Formation and its equivalents were deposited may, therefore, have been elongate in a northeast-southwest direction. Even so, a basin this long might well have greater lateral extent, and the narrow configuration preserved today may be the result of structural shortening across the basin. Erosion prior to the deposition of the Addy Quartzite is known to have removed parts of both Precambrian sections and could have removed all or most of the Huckleberry and Monk Formations from above the Belt Supergroup now preserved east of the Jumpoff Joe fault.

There is a consistent difference in trend between the Deer Trail Group and the Belt Supergroup block. North of Chewelah the northeast-trending Deer Trail rocks abut younger plutonic rocks but continue with the same trend on the other side of the batholith, in the

Metaline quadrangle. In the Metaline quadrangle they are called the Priest River Group (Park and Cannon, 1943, p. 6), but are correlated by Becraft and Weis (1963, p. 16) with the Deer Trail Group.

Just south of the Metaline quadrangle, in the Bead Lake area, Schroeder (1952, pl. 1) showed the Newport Group (Belt Supergroup) with a generally north-south strike. Although the Newport and Priest River Groups are separated over a distance of 15–20 miles by younger plutonic rocks, their structural relationship appears to be the same as that of the Belt Supergroup and Deer Trail Group, respectively, in the report area. In both areas the Deer Trail Group or its correlatives are confined to the northeast-trending block, which appears to truncate the less well defined north-south trend of the Belt Supergroup and its correlatives. Only locally in the northeastern part of Washington are the Belt rocks known to have this northeast regional trend, and only locally does the Deer Trail Group diverge from it.

STRUCTURE SEPARATING THE TWO BLOCKS

Stratigraphic differences between the Belt Supergroup and Deer Trail Group, structural differences between the two blocks to which they are confined, and differences in the sections which overlie the two lead to the conclusion that they are separated by a thrust fault along which an undetermined, but possibly large, amount of movement has occurred. This fault is thought to be the Jumpoff Joe fault. However, subsequent faulting, possibly regional tilting, and concealment by glacial and alluvial debris combine to make study of the fault contact difficult.

The Jumpoff Joe fault is exposed at several localities in the report area. It separates rocks of the Deer Trail Group from those of the Belt Supergroup, but whether it juxtaposes the two sections or merely downdrops the major structure is uncertain. The fault is well located but poorly exposed on the hill immediately west of Jumpoff Joe Lake; here Paleozoic carbonate rocks dip under argillite probably belonging to the Deer Trail Group. South of there the fault is covered by younger deposits for a distance of 2 miles but crops out on the hill north of Springdale, where it again places presumed Deer Trail argillite against Paleozoic carbonate rock. Here the fault dips at a moderate to shallow angle under the Deer Trail Group and is clearly a low-angle reverse or thrust fault. The fault is covered from Jumpoff Joe Lake to Chewelah. It is exposed on Embry Hill northeast of Chewelah and north of there on the west flank of Eagle Mountain. The actual fault plane or fault zone is not exposed on Embry Hill but can be located precisely enough to confidently show that the fault dips shallowly to the west under the Deer Trail Group. The fault is steeper on Eagle Mountain but may be slightly downdropped by a younger, high-angle fault. North of Eagle Mountain relations are obscured by glacial cover.

The fault appears to bend to the west at the north end of Eagle Mountain and then continue northward through The Tinderbox. The rocks in that area are extremely deformed and resemble those adjacent to the fault on Eagle Mountain. From The Tinderbox the fault must follow Bayley Creek for a short distance and then continue to the northeast, forming the boundary between the Starvation Flat Quartz Monzonite and the Phillips Lake Granodiorite. Where it cuts through the plutonic rocks, the fault forms a prominent mylonite and cataclastic zone as much as 500 feet wide. However, from the south boundary of sec. 31, T. 34 N., R. 41 E., to the north edge of the area the identity of this fault is uncertain. It may be the Jumpoff Joe fault, one of the northeast-striking faults which project from the southwest, or a combination of both. The mylonite and cataclastic zone is thought to be part of the Jumpoff Joe fault because it dips from 30° to 45° NW. In addition, no extension of the Jumpoff Joe fault is known on the northwest side of the zone.

An inferred fault passes just east of The Tinderbox, across the north fork of Chewelah Creek, and up Bear Canyon (pl. 1). It merges with the Jumpoff Joe fault to the south and with the mylonite and cataclastic zone to the north. The fault is drawn through a number of isolated but extremely contorted and brecciated outcrops and may represent the main trace of the Jumpoff Joe fault.

If the Jumpoff Joe fault is the major structure that juxtaposes the two Precambrian sections, then the Deer Trail Group has been thrust over the Belt Supergroup; but if it merely downdrops the major structure, then, because of the distribution of the two sections, the Belt Supergroup would be thrust over the Deer Trail Group (fig. 24). Although the field relations permit either conclusion, the former is favored because it is simpler and because the trends of the Deer Trail block appear to truncate those of the Belt Supergroup block. The fact that it is not even known with certainty which section has overidden which, or for that matter if one section is even thrust over the other, illustrates how equivocal the data concerning this problem are. Regardless of the exact form of the structure, however, the obvious differences in trend, internal structures, and stratigraphy of the two blocks indicate the existence of some sort of large-scale break.

Although there are significant facies changes, fairly good stratigraphic correlations can be made between the Belt Supergroup in the Coeur d'Alene district, at Clark Fork, and in the report area. The facies changes are not unreasonable, considering the distances in-

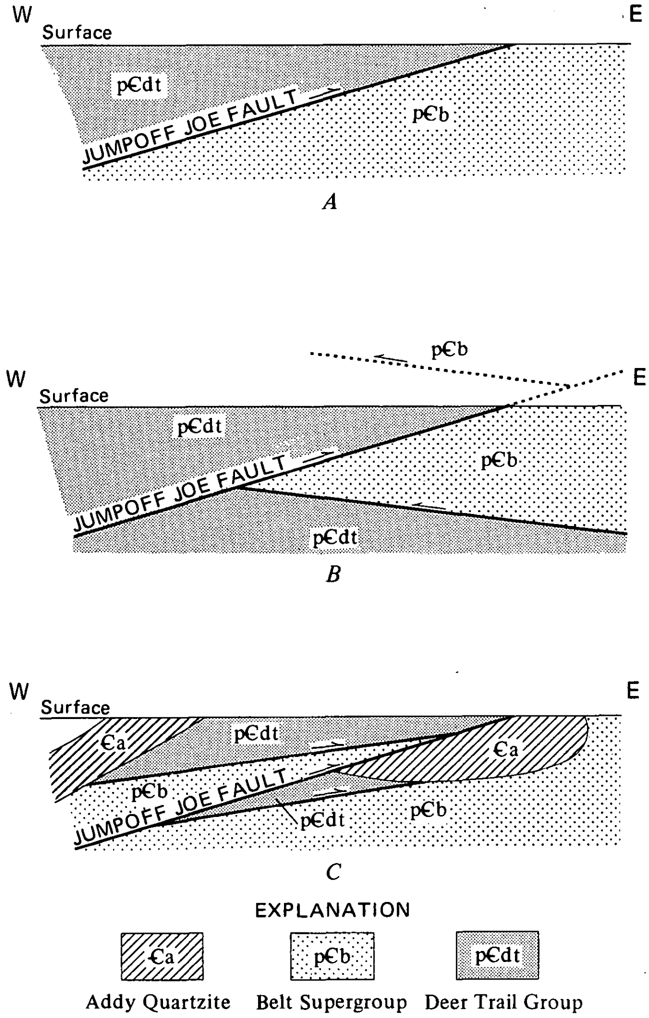


FIGURE 24.—Diagrammatic cross sections showing possible interpretations of thrust relations between the Belt Supergroup and Deer Trail Group. *A*, Simple thrust relationship. Deer Trail Group thrust eastward over the Belt Supergroup. This is considered the most likely relation and is the one shown on the geologic map and sections. *B*, Belt Supergroup thrust westward over the Deer Trail Group. Thrust later displaced by the Jumpoff Joe fault and upper part removed by erosion. *C*, Deer Trail Group thrust eastward over the Belt Supergroup. Thrust is in subsurface only and is displaced by the Jumpoff Joe fault. This interpretation assumes the thrusting is Precambrian in age.

volved. No large-scale thrust faults are known in the area between these localities, and so if the Belt section of the report area had been thrust over the Deer Trail Group, the rest of the Belt rocks east to the Clark Fork and Coeur d'Alene areas would have had to move with it. The thrust sheet would be tremendously large and would contain no known imbricate thrusts, and the thrust plane would be nowhere exposed at the surface. If this is the case, however, the upper plate would be thrust from east to west, as no section resembling the Deer Trail Group is known east of Chewelah.

For the same reason, if the Deer Trail Group is thrust over the Belt Series, as is suspected, the upper plate must be thrust from west to east. Unfortunately there are no exposures of the Deer Trail Group or Belt Supergroup west of the relatively narrow magnesite belt that would allow an estimate of how much movement has occurred. The facies changes in the Belt section between Clark Fork and the report area are more pronounced than those between the report area and the magnesite belt. This suggests that the amount of thrusting may not be too great.

Unfortunately, a complete section of Addy Quartzite resting on the Belt Supergroup is nowhere exposed in the area. Because systematic differences in thickness of the basal Cambrian quartzite are known to exist in the region, comparison of the thickness of the quartzite on the Belt Supergroup with that above the Deer Trail Group might furnish information on how far the two sections were originally separated. Addy Quartzite overlying the Belt rocks east of the area should be examined with this in mind.

Paleozoic rocks on Addy Quartzite appear to be different above the two Precambrian sections, but the Paleozoic carbonate rocks above the Belt Supergroup are not well enough nor extensively enough exposed to make meaningful comparisons with those above the Deer Trail Group. In the Hunters quadrangle, which includes the southern part of the magnesite belt, Campbell and Raup (1964) showed several thousand feet of limestone, slate, and chert of Ordovician age or older above the quartzite, where it overlies the Deer Trail Group. Yates (1964) also showed several thousand feet of carbonate, phyllite, and slate of Ordovician age or older above the quartzite, all of which apparently overlies the Deer Trail or Priest River Group, but no rocks corresponding to either of these thick sections are found above the quartzite resting on the Belt Supergroup. Throughout the report area, Mississippian carbonate rocks are found above the Addy Quartzite where it rests on the Belt rocks, but they have not been found anywhere else in northeastern Washington.

The Jumpoff Joe fault is thus presumably younger than the Mississippian carbonate rocks and is probably younger than the Starvation Flat Quartz Monzonite and the Phillips Lake Granodiorite. This would indicate that the thrusting took place less than 100 m.y. ago.

Although it has been suggested here that the Jumpoff Joe fault is the major structure separating the Deer Trail and Belt blocks, it is possible that the large northeast-striking faults that pass a few miles northwest of Chewelah mark the division between the blocks and that one of the faults in that system is significantly larger than the others. If this were the case, several interpretations that have been proposed would have to be altered: (1) The rocks west of the Jumpoff Joe fault that have been assigned to the Deer Trail Group, primarily on the basis of their association with the greenstone of the Huckleberry Formation, are actually part of the Belt Supergroup, (2) the greenstone, although thin, is present within the eastern structural block, and (3) the Jumpoff Joe fault, although a large fault, has not had a significant amount of lateral movement along it.

There are two significant reasons for suggesting that this other fault system, and not the Jumpoff Joe fault, is the major structure separating the two structural blocks: (1) All rocks and structures with the northeast trend so consistently found in the Deer Trail block would be restricted to that block. Most of the rocks in the report area west of the Jumpoff Joe fault and south of Chewelah do not have this northeast strike. To suggest stratigraphic assignments on the basis of structural evidence in this manner is ordinarily not justified. However, the consistency of the northeast trends within the Deer Trail block is so striking and is developed over such a large area that any rocks assigned to this block not having the northeast trend should be regarded with some suspicion. (2) The topographic lineament formed by faults of this zone within the report area continues to the southwest outside the area. Campbell and Loofbourow (1962, pl. 1) mapped a fault along part of this lineament, but the fault passes beyond the limits of their mapping a short distance to the south.

Considering stratigraphic and structural relations jointly, the following correlations and sequence of events are tentatively proposed to explain the relation between the Deer Trail Group and Belt Supergroup: (1) The Deer Trail Group is roughly equivalent to the upper Wallace Formation and the Striped Peak Formation; (2) the strata of the two blocks were deposited at localities more widely separated than at present, although information is not sufficient to accurately estimate how far apart they were; (3) the two groups underwent separate, though possibly related, deformation at their respective sites of deposition; and (4) sometime after the extrusion of the Huckleberry volcanics and deposition of the Addy Quartzite, and probably after intrusion of the Starvation Flat Quartz Monzonite, the two sections were brought together by thrust faulting.

OTHER FAULTS

Several of the faults that have a general north-south strike appear to be older than the Flowery Trail Granodiorite because the brecciated rocks in one of the fault zones near the pluton are thoroughly recrystallized. Most of these faults are not exposed, but are inferred because of anomalous relationships across alluvium-filled valleys. The faults immediately east and west of the Jumpoff Joe fault appear to belong to this group. These faults may have been active at the same time as the large northeast-striking faults, some of which are interpreted as cutting the north-south faults and some as being cut by them. The fault separating the Addy Quartzite from the Paleozoic carbonate rocks southeast of Chewelah probably belongs to this group.

Two sets of northeast-striking faults are found in the Belt Supergroup block, but one set has a consistently different strike (N. 50°–60°E.) from the faults in the Deer Trail Group block. The large fault that passes through the Upper Cottonwood Road area is representative of that group. The other set, which approximates the N.30°–40° E. strike of the major faults in the Deer Trail Group block, is not represented in the report area. Three large faults with this trend are found south of Horseshoe Lake, about 4 miles east of Blue Grouse Mountain.

West of Loon Lake, three shear zones are found in the 50-m.y.-old Silver Point Quartz Monzonite. These zones strike about north-south to north-northwest. Within the zones the rock is highly brecciated and has been recemented by chloritic material and quartz. Aplite dikes in the zones are both twisted and broken, as if the process which formed the shear zones may have been active when the dikes were intruded and continued active for some time after the dikes solidified.

Numerous smaller faults, some inferred, some observed, are shown on the geologic maps. Most cannot be dated by crosscutting relationships with closely dated rocks or other structures.

MINERAL DEPOSITS

Commodities of economic interest in the report area include copper, silver, lead, tungsten, barite, clay, silica sand, and feldspar. Individual mine workings and prospects that were accessible between 1963 and 1968 were described in detail by Clark and Miller (1968, p. 4–5 and pl. 2) and Miller (1969, p. 6). Clark and Miller also described from old company mine maps some workings in the Eagle Mountain area that were not accessible during the study period. Only general mine areas will be discussed here. The reader is referred to earlier reports by Weaver (1920), Patty (1921), Huntting (1956), Clark and Miller (1968), and Miller (1969) for individual mine descriptions.

In terms of total value of ore produced, the copper-silver mines around Eagle Mountain are by far the most important in the area. Of the 10,551,098 pounds of recorded copper production from the report area, about 94 percent is accredited to the mines around Eagle Mountain (Fulkerson and Kingston, 1958, p. 27–28). Almost 100 percent of the recorded silver production also comes from these mines.

The copper-silver ore is found in quartz-carbonate veins which range in thickness from a few inches to over 25 feet. Most are between 10–25 feet wide and are within or near well-developed steeply dipping north-northeast shear zones confined to the upper and lower parts of the Wallace Formation. Sulfides are sparsely disseminated in the veins and host rock and include pyrite, chalcopyrite, tetrahedrite, pyrrhotite, covellite, and chalcocite or digenite. Malachite and hematite are also present. Most of the mines in the north half of the report area, including those around Eagle Mountain, are in country rock peripheral to the Flowery Trail Granodiorite, and the mineralization may be related to that pluton.

The sections accompanying the preliminary geologic map by Clark and Miller (1968, pl. 1) show Eagle Mountain to be underlain by a nearly horizontal thrust fault. However, additional work since the publication of that map and the delineation of individual Belt formations suggest that the mountain probably is not floored by a thrust fault and that the shear zones which appear to control the location of the veins are probably continuous at depth. Both the thrust and nonthrust interpretations are based on poor exposure in critical areas, however, and should be reevaluated if additional data become available.

Only minor exploration has ever been attempted below the main adit level in any of the mines in the Eagle Mountain area. Regardless of whether the area is floored by a thrust fault or not, the vein system should be explored at depth by drilling. Patty (1921, fig. 8) showed a rich streak of tetrahedrite continuing 350 feet below the main adit level in one mine and widening with depth. In addition, old company maps and records indicate an increasing silver-copper ratio at depth. The ever-increasing demand for silver possibly warrants exploratory drilling of the lower parts of the vein system.

Smaller deposits have been mined on Embry Hill, about 1 mile southwest of the Eagle Mountain area. These are also on the north side of the Flowery Trail Granodiorite. They are roughly on strike with the Eagle Mountain vein system, but whether the two are continuous is uncertain because faults and an area of poor exposure separate the two localities. Only 4,011 pounds of copper and 22 ounces of silver have been recovered from the ore of the Embry Hill mines (Fulkerson and Kingston, 1958, p. 28). The deposits differ from those on Eagle Mountain in that they have a much higher copper-silver ratio and also contain small amounts of molybdenite. The Embry Hill veins may have been emplaced within the southern continuation of the Eagle Mountain shear zone or formed in subsidiary breaks related to the Jumpoff Joe fault; the latter may border the west margin of the mineral deposits at this locality.

A few deposits of lesser importance are found on the south side of the Flowery Trail Granodiorite. Only the Jay Gould mine has had recorded production. The veins in the Jay Gould area are typified by lead and silver-bearing minerals and thus differ from the Eagle Mountain and Embry Hill ores. Argentiferous galena, the major ore mineral, is found in quartz veins as much as 10 feet wide.

The distribution of base metal deposits around the Flowery Trail Granodiorite appears to exhibit a zonation of sorts. The lead-silver deposits of the Jay Gould area and the Blue Star mine of the Eagle Mountain area occur within the margins of the pluton and in the adjacent sedimentary rocks. Next and relatively near the contact are the silver-copper deposits of Embry Hill, in which the silver-copper ratio is relatively low. The Eagle Mountain deposits, which have a high silver-copper ratio, are farthest from the pluton.

The only other base metal deposit in the report area which has had any significant production is the Loon Lake copper mine, in the NE¼ sec. 33, T. 31 N., R. 41 E. It has a production record of 622,555 pounds of copper and 532 ounces of silver (Fulkerson and Kingston, 1958, p. 45). The production was recorded principally for the period between 1916 and 1919 and came from an individual ore shoot in a quartz vein from 4 to 20 feet wide. Most of the ore came from a secondary zone of azurite, malachite, and cuprite. No new shoots were found by additional exploration. No other base metal mines with any production are located within several miles of the Loon Lake copper mine, although numerous large quartz veins similar to the one on which the property is developed are found in the mountains to the east and southeast.

Several tungsten deposits have been found on Blue Grouse Mountain east of Deer Lake. They consist of sparsely disseminated huebnerite crystals in quartz veins, greisen, and pegmatite segregations around the periphery of the muscovite quartz monzonite. The mineralization appears to be related to this quartz monzonite pluton.

Barite deposits have been found on Eagle Mountain, the hills north and east of Valley, and a few hundred feet southwest of the Loon Lake copper mine. Moen (1964) described all the deposits, gave production figures, and estimated reserves. All the barite occurs in veins or a series of veins which range in width from less than 1 inch to several tens of feet. Moen reported 634 tons of barite shipped, reserves of 2,300 tons measured and indicated, and 24,750 tons inferred. The veins are scattered and appear to be unrelated, but all are restricted to the Striped Peak Formation or its probable equivalent, the Deer Trail Group.

Two claypits in the Latah Formation are located southeast of Deer Lake, near the south edge of the map area. Similar deposits are found less than 1 mile east of the report area at the same latitude as those southeast of Deer Lake and just south of the area at the town of Clayton. Hosterman (1969, p. 55, 56) studied the clay deposits of Spokane County in detail. He described the deposit just east of the area and gave analyses of the clay. Miller (1969, pl. 1) prepared a highly generalized map which shows the possible location of additional clay in nearby areas.

The Addy Quartzite may in part be pure enough to be a source for silica sand. Above the purple and pink zones at the base of the formation, the sand is generally white or light gray and probably almost pure quartz. An extensive sampling program would be necessary to block out an area of sufficient purity. On Lane Mountain, a few miles west of the area, the quartzite is being mined for the manufacture of glass. There, however, the silica cement has been removed or was never deposited, and the rock is extremely friable. All the Addy Quartzite in the report area would require considerable crushing.

Feldspar for use as a flux in glass manufacture may be available from the muscovite quartz monzonite. Analyses of this unit (see section "Muscovite Quartz Monzonite, Petrology") show that it contains very little iron or manganese, two of the chief contaminants in glass making. Abundant, easy-to-reach reserves are available in the southeastern part of the area and just east of the area.

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INDEX

[Italic page numbers indicate major references]