Geology of the Stensgar Mountain Quadrangle, Stevens County, Washington

U.S. GEOLOGICAL SURVEY BULLETIN 1679

U.S. DEPARTMENT OF THE INTERIOR

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Geology of the Stensgar Mountain Quadrangle, Stevens County, Washington

By James G. Evans

Stratigraphy and structure
of the Proterozoic and
Lower Cambrian rocks

U.S. GEOLOGICAL SURVEY BULLETIN 1679

DEPARTMENT OF THE INTERIOR
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Geology of the Stensgar Mountain Quadrangle, Stevens County, Washington

By JAMES G. EVANS

Abstract

The Stensgar Mountain quadrangle contains a section of the Middle Proterozoic Deer Trail Group at least 4,000 m thick, a section of the Middle(?) and Late Proterozoic Windermere Group at least 1,600 m thick, and Lower Cambrian strata at least 2,600 m thick. The Deer Trail Group consists of the following formations, listed from oldest to youngest: Togo Formation, Edna Dolomite, McHale Slate, Stensgar Dolomite, and Buffalo Hump Formation. The overlying Windermere Group consists of the Huckleberry (conglomerate, metabasalt, metatuff) and Monk (slate, conglomerate, dolomite) Formations. Gabbro and metabasalt feeder dikes of the volcanic rocks member of the Huckleberry Formation intruded the Deer Trail Group and the conglomerate member of the Huckleberry Formation causing minor contact metamorphism. The Lower Cambrian Addy Quartzite and Old Dominion Limestone were deposited unconformably over the Proterozoic assemblage. After deposition of the Lower Cambrian units, the section consisting of the Togo Formation through the Addy Quartzite underwent low-grade regional metamorphism, probably as a result of deep burial. The entire Deer Trail Group is enclosed in a duplex thrust with the Lane Mountain thrust as the floor thrust and the Stensgar Mountain thrust as the roof thrust. Contemporaneously with thrusting, the rocks were subjected to low-grade dynamothermal metamorphism during which slaty cleavage formed in the predominantly argillaceous Deer Trail Group and folding occurred in the Proterozoic rocks and Addy Quartzite. During metamorphism, most sedimentary sections were thinned and bedding and older faults were rotated subparallel to the cleavage. Thrusting and dynamothermal metamorphism occurred in the Late Triassic to Cretaceous. The Deer Trail Group was intruded by felsic dikes in the Mesozoic or Tertiary and by andesite dikes, possibly in the Tertiary. Pleistocene and Holocene glacial deposits and alluvium cover the lower elevations of the quadrangle.

INTRODUCTION

The Stensgar Mountain quadrangle is 70 km north-northwest of Spokane, Wash., and 10 km west of the town of Valley in Stevens County (fig. 1). The study area, which is near the middle of the northeast-trending "magnesite belt," was named by Weaver (1920, p. 319) for its deposits of crystalline magnesite. This quadrangle was chosen for study because it contains a representative section of the Middle Proterozoic Deer Trail Group. These rocks have been correlated with the Priest River Group east of Metaline Falls near Metaline (Becraft and Weis, 1963, p. 16), and with the Belt Supergroup (Campbell and Loofbourow, 1962, p. F20; Miller and Clark, 1975, p. 18). However, the stratigraphic sequences of the Deer Trail Group and the Belt Supergroup are different, and the low-grade dynamothermal metamorphism undergone by the Deer Trail Group contrasts with the generally unmetamorphosed sedimentary character of most of the Belt Supergroup. A detailed study of the Deer Trail Group was undertaken to improve understanding of the stratigraphy and structure of the magnesite belt and to clarify the relation of these rocks to the Belt Supergroup.

In addition to the crystalline magnesite, the belt contains deposits of antimony, barite, gold, lead, silver, tungsten, uranium, and zinc. Small deposits of all but tungsten and uranium are present in the Stensgar Mountain quadrangle.

The first geologic map of the magnesite belt appeared in Weaver's (1920) report on the mineral resources of Stevens County. Progressively more detailed maps of the belt were made by Bennett (1941) and by Campbell and Loofbourow (1962). Reports accompanying these maps describe prior studies in detail. Parts of the belt were later remapped by Becraft and Weis (1963, Turtle Lake quadrangle), and by Campbell and Raup (1964, Hunters quadrangle). Miller and Clark (1975) studied the easternmost exposures of the Deer Trail Group in the Chewelah-Loon Lake area.

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I greatly appreciate the ideas and comments by my Geological Survey colleagues, K.F. Fox, Jr., F.K. Miller, and J.W. Whipple, and by R.F. Burmester, Western Washington University.

GEOLOGIC SUMMARY

The quadrangle contains a thick section (at least 4,000 m) of the Middle Proterozoic Deer Trail Group, which consists of, from oldest to youngest, the Togo Formation, Edna Dolomite, McHale Slate, Stensgar Dolomite, and Buffalo Hump Formation (pl. 1). The Deer Trail Group is overlain by the Middle(?) and Late Proterozoic Windermere Group (at least 1,600 m), consisting, in ascending order, of the conglomerate and volcanic rocks members of the Huckleberry Formation and the Monk Formation. Gabbro and metabasalt feeder dikes of the volcanic rocks member of the Huckleberry Formation intrude the Deer Trail Group and the conglomerate member of the Huckleberry resulting in minor low-grade contact metamorphism of the wall rocks.

The Proterozoic section is overlain by a thick (at least 2,600 m) Lower Cambrian section consisting of the Addy Quartzite and the overlying Old Dominion Limestone of Weaver (1920).

Sometime, possibly in the Paleozoic, the rocks underwent low-grade regional metamorphism, probably as a result of deep burial. Subsequently, the beds were tilted up to 45^{o} sometime before the Mesozoic deformation.

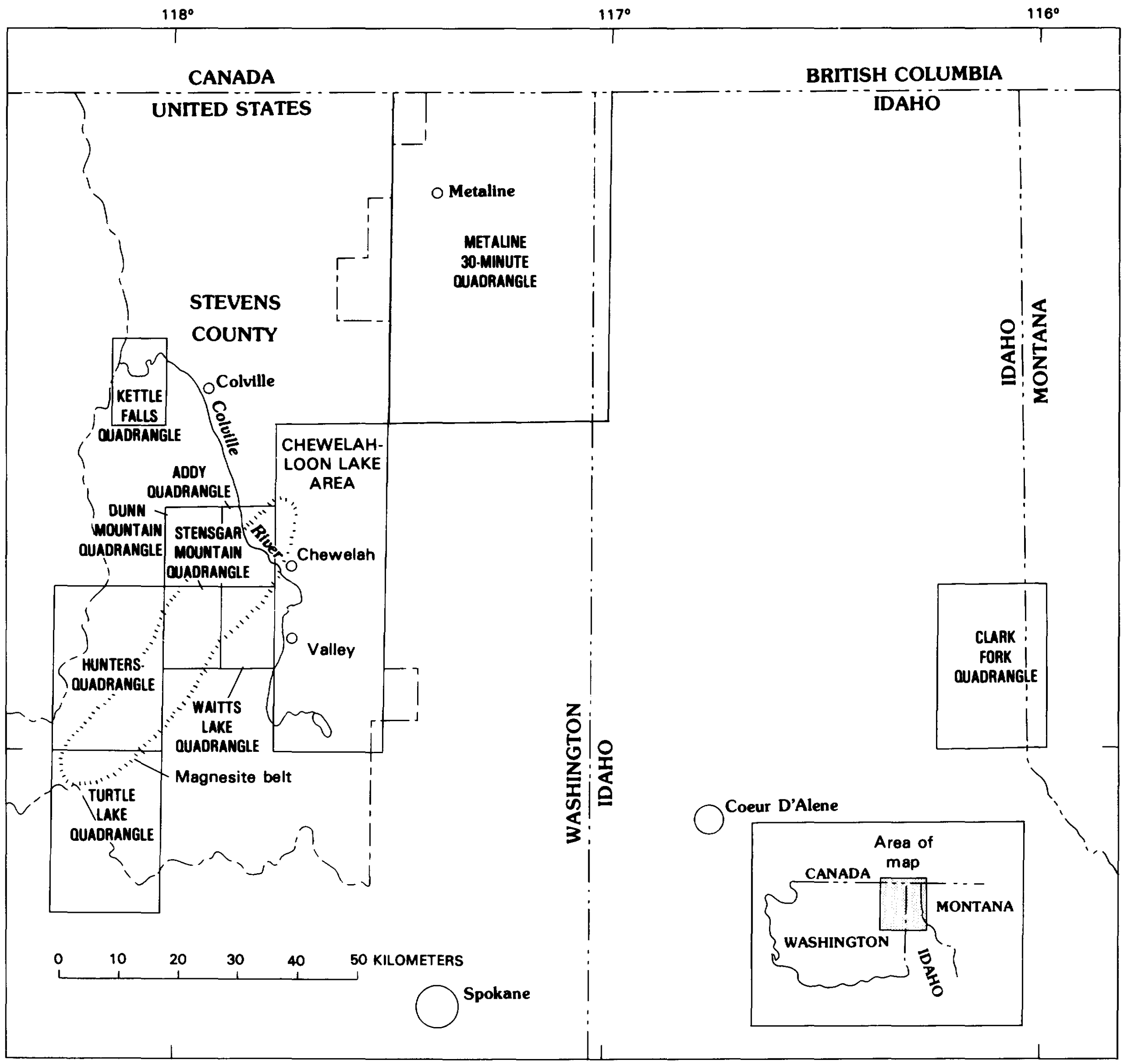


Figure 1. Index map showing location of Stensgar Mountain quadrangle and magnesite belt in Stevens County, Wash.

Superficially, the Deer Trail Group appears to be enclosed in a duplex thrust, having the Lane Mountain thrust as the floor thrust and the Stensgar Mountain fault as the roof thrust. The footwall of the Lane Mountain thrust, which is composed of the Addy Quartzite, indicates that the faulting must be post-Early Cambrian. The hanging wall of the Stensgar Mountain fault contains the Huckleberry Formation, Addy Quartzite, and Old Dominion Limestone. The thrusting, which may be as young as Cretaceous, approximately coincides with low-grade penetrative dynamothermal metamorphism in the Deer Trail Group.

Basic and felsic dikes of probable Mesozoic and Cenozoic age intrude the Deer Trail Group. Quaternary alluvium and glacial deposits cover much of the quadrangle.

STRATIGRAPHY

Proterozoic Rocks

Deer Trail Group

Introduction

Weaver (1920, p. 54-61) named the metamorphosed sedimentary rocks of the magnesite belt the Deer Trail Argillite and differentiated the Stensgar Dolomite as a member. Bennett (1941, p. 7) called the rocks the Deer Trail Group and differentiated upper and lower quartzite members, Stensgar Dolomite (as a member), and a lower dolomite member. Campbell and Loofbourow (1962, p. F8) divided the Deer Trail Group into the following formaions listed from oldest to youngest: Togo Formation, Edna Dolomite, McHale Slate, Stensgar Dolomite, and Buffalo Hump Formation. This stratigraphic nomenclature is used in this report.

Togo Formation

The Togo Formation was named for good exposures in the vicinity of the Togo Mine, southwest corner sec. 6, T. 29 N., R. 38 E., Hunters quadrangle (Campbell and Loofbourow, 1962, p. F8-F9). Exposures of the Togo Formation in the Stensgar Mountain quadrangle are generally poor, especially in the southeastern part of the study area.

The formation consists largely of silvery gray to black thinly cleaved slate, gray and black siltite, and thinly interbedded siltite and slate (rhythmites). The siltite beds are usually lighter in color than the slate. Minor color variations in all lithologies include green and brown and white. The rocks in places are bleached pale gray adjacent to fractures. Some of the rocks contain disseminated pyrite. Some slate has alternating light and dark beds that are less than 2 mm thick. In the siltite and in sections of interbedded siltite and slate, the beds range from less than 1 mm to 4 cm. Some siltite has crossbeds and graded beds.

Muscovite, some of which may be sedimentary, and angular clasts of quartz and feldspar less than 0.05 mm make up most of the rock. Feldspar clasts have been altered to epidote and albite. Stilpnomelane and sphene, probably of metamorphic origin, are locally abundant, especially in zones parallel to bedding. Allanite, carbonate, iron oxides (hematite and magnetite) are accessory minerals, commonly concentrated along bedding surfaces or as a matrix of certain beds; these minerals are probably of sedimentary or diagenetic origin. Some magnetite and pyrite form porphyroblasts that postdate the principal phase of dynamothermal metamorphism.

Fine- to medium-grained quartzite and quartz siltite make up less than one percent of the unit, forming beds and lenses 1-90 m thick; most are less than 10 m. The characteristics of the quartzite varies widely even within an outcrop. It is white, gray, brown, green, pink, and red, and thin to thick bedded (1-1,000 mm). Most of the quartzite is pure; some of it has varying amounts of dark-gray, black, or green shaly partings and lenses as much as 6 cm thick. Ripple marks and mud cracks, observed by Miller and Yates (1976, p. 23) in Togo quartzites to the east of the study area, were not found in the Stensgar Mountain quadrangle.

In thin section, the rock is seen to be 50-70 percent medium sand-sized quartz (as much as 0.5 mm) in a matrix of very fine sand-sized quartz (as much as 0.1 mm) and muscovite. Locally, dolomite constitutes as much as 50 percent of some of the quartzite. Hematite, leucoxene, magnetite, and tourmaline clasts are minor constituents.

Dolomite, commonly weathering orange and rusty brown, is the least abundant rock type in the Togo Formation. Beds and lenses range from 15 to 90 m thick and are usually gray to black. The dolomite is thin to thick bedded (8-100 cm) or is massive. In places the dolomite is serpentinized. The rock is 80-97 percent silt-sized dolomite grains (less than 0.05 mm) and as much as 20 percent silt-sized clasts of quartz and feldspar. Stilpnomelane, hematite, and muscovite are minor constituents.

A partial section of the Togo Formation is 600 m thick in section 31, T. 31 N., R. 39 E., sec. 36, T. 31 N., R. 38 E., and sec. 6, T. 30 N., R. 38 E. Strata to the east and west of this section are folded. Although the quartzite is lithologically similar to a much thicker quartzite at the top of the Togo Formation in the Hunters (Campbell and Raup, 1964) and Turtle Lake (Becraft and Weis, 1963, pl. 1) quadrangles to the southwest, the quartzite in the Togo Formation of the Stensgar Mountain quadrangle appears to be much lower in the section.

The fine-grained evenly laminated character of the Togo suggests relatively quiet-water deposition. The mud cracks observed in the formation east of the study area (Miller and Clark, 1975, p. 23) do not clearly indicate a shallow-water environment because they also could be deep-water shrinkage cracks and may not be due to desiccation.

Edna Dolomite

The Edna Dolomite was named by Campbell and Loofbourow (1962, p. F9) for outcrops in the vicinity of the Edna Mine (SE cor., sec. 9, T. 31 N., R. 39 E.). The least altered and most accessible exposures of the unit are along the Red Marble Road in sections 19, 20, and 29, T. 31 N., R. 39 E. This dolomite is commonly thick bedded (beds 15-300 cm), rusty brown, fine grained and impure. Zones of thin bedded (beds as much as 3 cm) and laminated dolomite are present between some thick beds. Some of the dolomite is massive. Parts of the formation that are thinly cleaved, usually parallel to bedding, consist of pale-gray dolomite with chlorite and muscovite along cleavages. Pencil cleavage is locally developed. In parts of the dolomite, especially where it is massive, white quartz veins as much as 6 cm thick intersect the bedding surfaces at large angles. In the vicinity of the Edna Mine, the dolomite is pale gray to white, generally thin bedded, recrystallized, and thinly cleaved in some zones. In sec. 16, T. 31 N., R. 39 E. the dolomite has slaty interbeds.

The carbonate is dominantly dolomite. In thin section, the dolomite appears very fine grained (0.01-0.05 mm) and is associated with locally abundant (as much as 10 percent) quartz, muscovite, and chlorite, and minor feldspar clasts, calcite, hematite, leucoxene, and pyrite. Rare beds of very fine grained limestone with as much as 10 percent quartz clasts are present.

Lenses of slate, quartzite, and quartz siltite are present in the Edna Dolomite: thin-bedded quartzite at least 3 m thick, in SW1/4 sec. 10, T. 31 N., R. 39 E.; a lens of black and silvery-green slate (60 m) and quartzite (15 m) in NW1/4 sec. 15, same township; thin-bedded to laminated gray quartzite at least 3 m thick overlain by 32 m of black slate, in SE1/4 sec. 20, same township. Quartzite immediately above the dolomite is included in the formation, following Campbell and Loofbourow (1962). This quartzite, light gray to white with beds 0.5-50 cm thick, is as much as 67 m thick. In places, the upper quartzite is missing; it may have been removed by faulting along the contact between the Edna Dolomite and the overlying McHale Slate.

The fine-grained quartzite and quartz siltite contain abundant feldspar clasts and as much as 20 percent dolomite cement. Minor amounts of muscovite and pyrite partly altered to hematite are present.

A composite section of the Edna Dolomite is shown in figure 2. Most of the formation is dolomite; siliceous lithologies are more common in the upper half of the formation. The thickness determined from cross sections of 695 m represents a minimum because the contact with the underlying Togo Formation is a fault.

No sedimentary structures indicative of conditions of deposition are preserved in the Edna Dolomite. The stratigraphic succession and fine clast size in most of the formation suggest quiet-water deposition.

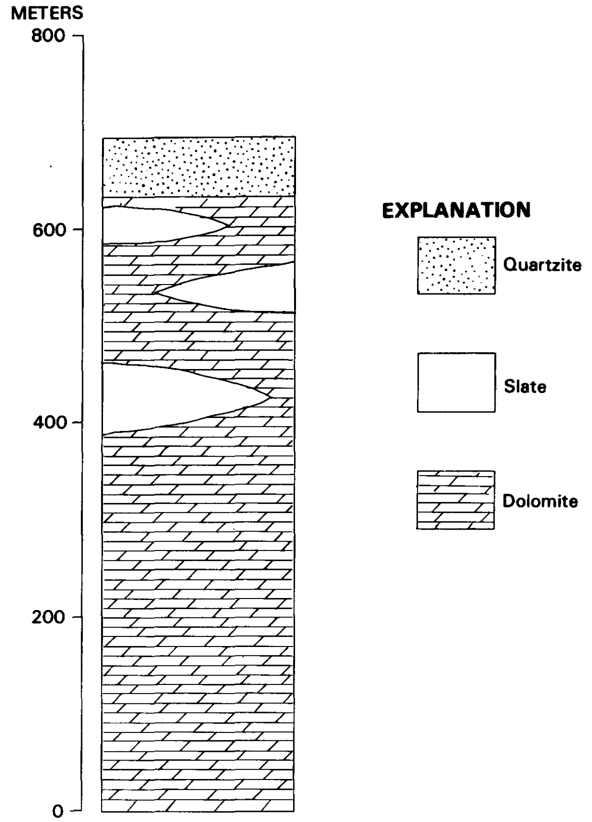


Figure 2. Composite section of the Edna Dolomite.

McHale Slate

The McHale Slate was named by Campbell and Loofbourow (1962, p. F11) for good exposures in McHale Canyon, sec. 19, T. 32 N., R. 40 E., Addy quadrangle (fig. 1). This canyon is named McKale Canyon on the topographic map. The formation is poorly exposed in the Stensgar Mountain quadrangle.

The rock, typically light to dark gray, is black slate and argillite (fig. 3). Local green and brown zones in the slate have no known stratigraphic significance. In places argillite is interbedded with light-gray siltite beds 2-20 mm thick (rhythmites), some of which are graded and contain minor siltite dikes as much as 3 mm long. Soft-sediment breccia in the McHale Slate, observed by Miller and Clark (1975, fig. 6) east of the study area, was not found in the Stensgar Mountain area. Lenses of pale-gray dolomite as much as 4 cm thick are interlayered with silvery-gray slate at a locality along the Red Marble Road in SW1/4 sec. 19, T. 31 N., R. 39 E.

The rock is composed mostly of angular quartz and feldspar clasts as much as 0.05 mm and muscovite. Some quartz and feldspar grains are elongate 2:1 parallel to cleavage. Chlorite, pumpellyite, pyrite, sphene, and stilpnomelane are locally abundant. Calcite, carbonaceous matter, hematite, leucoxene, magnetite, and tourmaline (clasts) are minor constituents.

The greatest apparent thickness of the slate, determined from exposures in sec. 20, T. 31 N., R. 39 E., is 1,510 m.

Relatively quiet water characterized the environment of deposition for most of the McHale Slate. The soft-sediment breccias observed by Miller and Clark (1975, fig. 6) are apparently the result of slumping.

Stensgar Dolomite

The Stensgar Dolomite was named by Weaver (1920, p. 57-59), who considered it a member of his Deer Trail Argillite. Bennett (1941, p. 7-8) recognized the dolomite as a member of the Deer Trail Group. Campbell and Loofbourow (1962, p. F8) raised the unit to formational rank. Outcrops of the unit are generally poor because they are concealed by regolith from the stratigraphically and topographically higher Buffalo Hump Formation. Some outcrops on ridge tops and in deeply entrenched canyons are excellent. The Stensgar Dolomite is also exposed in roadcuts, prospect pits, quarries, and small outcrops where it usually weathers from gray to brown.

The unit typically consists of white to light-gray laminated to thin-bedded (beds 1-40 mm) pure dolomite (fig. 4). Rare graded beds, as much as 3 mm thick, with abundant silt-sized dolomite clasts have what appear to be ripple marks less than 1 mm in amplitude. Some of the unit has silty and argillaceous interbeds. A few slate and conglomerate lenses are also present. Gray laminated argillite as much as 85 m thick is present near the middle of the 460-m-thick section of the Stensgar Dolomite in sec. 35, T. 31 N., R. 38 E. In the Red Marble Quarry two slate beds are present, one 2 m and the other 15 m thick. The thinner bed is black with scattered fragments of gray dolomite, possibly from brecciation (tectonic?) of pure dolomite layers. The thicker bed, possibly higher in the section than the thinner slate bed, is purple to silvery gray and grades laterally into dolomite. A 3-m-thick bed of dolomitic conglomerate, which is also present, may lie near the middle of the Red Marble section. This conglomerate contains angular to subrounded clasts of light-gray siltstone as long as 3 cm in a matrix of fine- to medium-grained black carbonate rocks, siltstone, and argillite. The dolomitic conglomerate bed may be a channel deposit. At the Keystone Mine a 3-m thick, dark-gray slate bed is present 376 m above the base of the section.

In thin section, the rock appears to be 80-99 percent carbonate; grains range from nearly cryptocrystalline to 0.5 mm. The carbonate minerals, identified by X-ray diffraction, are mostly dolomite and minor calcite. Some beds are composed of as much as 20 percent silt-sized quartz clasts. In some rocks, the clasts are rounded; in others they are elongate 3:1 parallel to a cleavage that parallels bedding.

Micaceous minerals, concentrated along the bedding surfaces, are chiefly muscovite and chlorite, but include minor biotite, stilpnomelane, and talc. Minor amounts of albite (metamorphic), carbonaceous matter, epidote, garnet, pumpellyite, pyrite, magnetite partly altered to hematite, sphene partly altered to leucoxene, tourmaline (clasts) and tremolite (metamorphic) are present.

A possible solution breccia was found in the Red Marble Quarry (fig. 5). The breccia body, about 15 m in length, consists of irregular fragments of very fine-grained, light-gray dolomite in a matrix of creamy-white medium- to coarse-grained dolomite. The contacts of the fragments with the matrix are lined with pyrite that forms a zone as much as 1 mm thick around the gray dolomite fragments.

Figure 3. The McHale Slate near top of hill 3496 near center of sec. 2, T. 31 N., R. 39 E. Handle of hammer is 30 cm. Cleavage dips steeply left. Bedding (not visible) strikes parallel to photograph and dips 50° into outcrop.

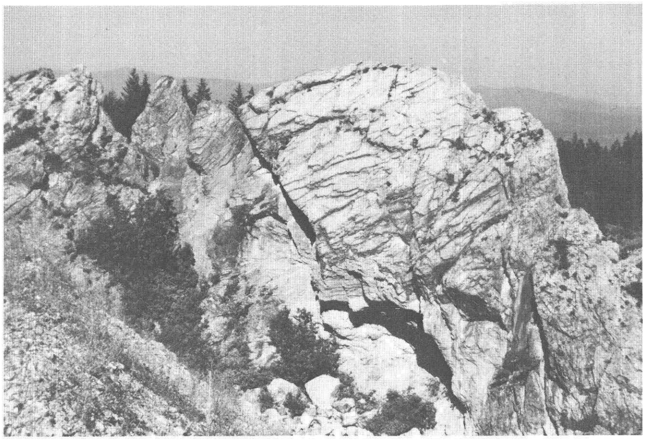


Figure 4. The Stensgar Dolomite at Keystone Mine, SW1/4 sec. 9, T. 31 N., R. 39 E. Top of outcrop is 20 m above bottom of canyon in foreground. Bedding dips to left. View east.

Figure 5. Possible solution breccia in the Stensgar Dolomite, Red Marble Quarry. Pen is 13 cm.

An estimate of the thickness of the Stensgar Dolomite is based on the thickest unfaulted section in the Keystone Mine area. Geometric construction indicates that the dolomite exposed between faults is at least 500 m thick. Other sections of the dolomite in the study area are much thinner, probably because of tectonic thinning.

Parts of the Stensgar Dolomite have been replaced by fine- to coarse-grained gray, black, and red magnesite (identified by X-ray diffraction pattern) in zones as much as 100 m wide, lying both parallel to and at angles to the bedding. During replacement the bedding was variously preserved, obscured, mimicked, or obliterated. Lack of disruption of bedding laminae in some rock altered to magnesite and absence of voids in replacement bodies suggest that there was no volume change during replacement. In many deposits, the magnesite-rich zones contain few silicic or pelitic impurities and the rock has zebroid structure; the white and gray bands in the rock appear to parallel and to lie at a low angle to bedding in nearby outcrops. The gray color of some bands is due to microscopic impurities, bands of which cut across magnesite grain boundaries. The composition of the impurities was not determined because of the very small size of the particles. In some deposits, gray magnesite porphyroblasts as much as 1 cm wide grew in fine-grained white magnesite rock.

Previous work summarized by Campbell and Loofbourow (1962, p. F36-F38), favors a hydrothermal origin for the deposits. Campbell and Loofbourow proposed that magnesitization accompanied or followed dedolomitization of the Edna or parts of the Stensgar Dolomite. The magnesium derived in this way was transported by hydrothermal solutions to parts of the Stensgar Dolomite and deposited. The low content of pelitic and silicic impurities in the host dolomite has been suggested as a controlling factor in localization of the deposits (Schroeder, 1948). Replacement features observed in the Stensgar Dolomite (Jenkins, 1918; Weaver, 1920, p. 322; Bain, 1924, p. 422; Willis, 1927) have been cited to support a hydrothermal origin for the magnesite. Campbell and Loofbourow pointed out that the magnesite must have been deposited after the regional deformation, because the magnesite does not appear to be deformed like the surrounding rocks (magnesite lacks folds and major faults), and, therefore, must be hydrothermal in origin. The study of joints in the magnesite of the Red Marble Quarry by Thole and Mills (1979, p. 154) suggests that the magnesite-bearing veins there developed during or subsequent to emplacement of the nearby Loon Lake batholith, dated at 93-63 Ma, or Cretaceous, by Ludwig and others (1981, p. 96-98, Pb/U, Pb/Th, and fission track methods) and is consistent with a hydrothermal origin.

However, the evidence cited above may also be interpreted another way. If the magnesite-bearing veins and other replacement features could have formed by late hydrothermal redistribution of the magnesite, the veins, then, provide little information about the origin of the magnesite. The strain history of the magnesite is partly related to its competence and provides no clear information regarding the relative age of the magnesite and regional deformation, contrary to what Campbell and Loofbourow (1962) suggested. Parts of the Edna Dolomite are as free of pelitic and silicic impurities as the Stensgar Dolomite, and moderately pure dolomite is present in places in the Togo and Monk Formations. If purity of the host dolomite is a major ore control, it is not clear why the Stensgar Dolomite is the only dolomitic unit to contain magnesite deposits. The veins of quartz, barite, and copper-bearing minerals (chalcopyrite, azurite, malachite) in the Edna Dolomite indicate that hydrothermal solutions did have access to this unit, but magnesite was not deposited in these rocks.

Kinsman (1967), when describing an occurrence of huntite $CaMg_3(CO_3)_4$ from a carbonate-evaporite environment along the Persian Gulf, suggested that some magnesite may have altered from huntite precipitated during early diagenesis. If his suggestion applies to magnesite in the Stensgar Dolomite, it would help explain this unique occurrence of magnesite in the Deer Trail Group. Fox and Rinehart (1968), who studied magnesite deposits in Triassic rocks of northern Okanogan County, concluded that the deposits were of sedimentary origin because of their wide distribution, their restriction to a narrow stratigraphic interval, and their interlayering with other metasedimentary rocks. Similar reasoning applied to the Deer Trail Group suggests that the magnesite deposits of Stevens County may also have had a sedimentary (or diagenetic) origin; however, until a detailed study of the magnesite deposits is made, a hydrothermal origin for the magnesite cannot be ruled out.

The pervasive thin-to-laminated relict bedding in the dolomite indicates largely quiet-water deposition. The formation, however, probably contains erosional unconformities within it as suggested by the presence of possible channel conglomerate and by bedding attitudes in dolomite at an angle to contacts with slate layers in the Stensgar (see sec. 35, T. 31 N., R. 38 E.). The general lack of pelitic and silicic sedimentary materials in most of the formation suggests distance or isolation from terrigenous sediment sources. The possible erosional unconformities suggest shallow-water deposition for the unit. The possible solution breccia and magnesite deposits would be consistent with deposition in an evaporative basin.

Buffalo Hump Formation

The Buffalo Hump Formation was named by Campbell and Loofbourow (1962, p. F17) for slates and quartzites exposed on a hill in SW1/4 sec. 3, T. 31 N., R. 39 E.. Although informally named the Buffalo Hump, it is unnamed on the Stensgar Mountain quadrangle. The formation consists of strata above the Stensgar Dolomite and below the conglomerate member of the Huckleberry Formation.

The lowermost part of the Buffalo Hump is typically thinly cleaved black, gray, brown, and green slate. However, in NW1/4 sec. 4, T. 31 N., R. 39 E. and in W1/2 sec. 1, T. 30 N., R. 38 E., quartzite overlies the Stensgar Dolomite (fig. 6). These relations suggest that an erosional unconformity is present in the lower part of the Buffalo Hump that extends down to the Stensgar Dolomite. Other unconformities probably exist in the Buffalo Hump where quartzite

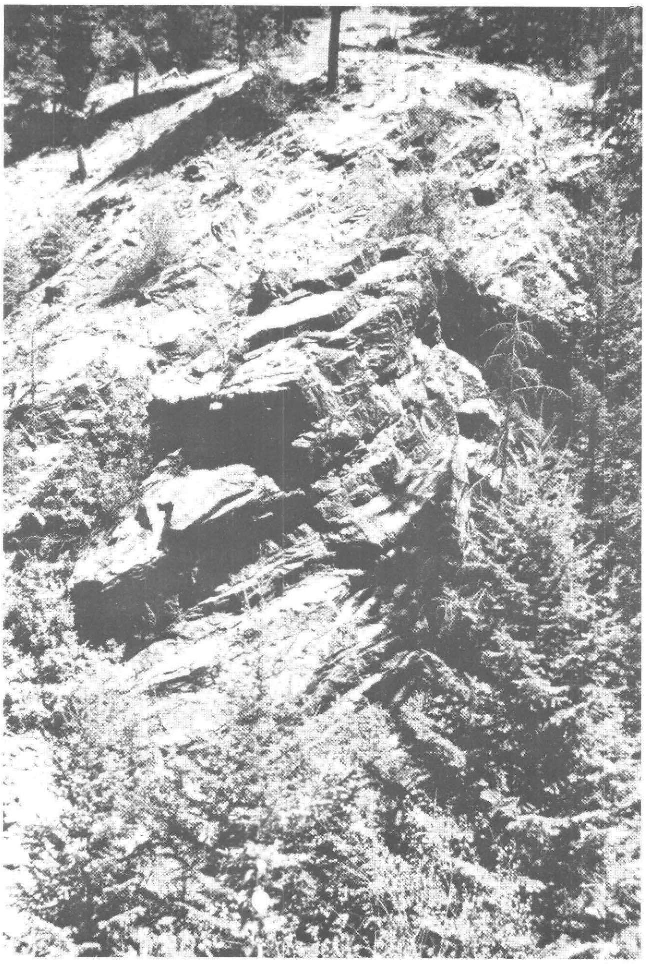


Figure 6. Bedded quartzite in the Buffalo Hump Formation, NW1/4 sec. 4, T. 31 N., R. 39 E. Tree trunk at top of hill in center is 30 cm in diameter. White sunlit bedding surfaces and vegetation are covered with volcanic ash from Mount St. Helens, Wash., (ash fall of June 23, 1980).

Figure 7. Stretched pebble conglomerate in the Buffalo Hump Formation, NE1/4 sec. 3, T. 30 N., R. 38 E. Head of hammer is 10 cm. Handle is aligned with elongation of clasts.

and slate beds vary greatly in thickness. At the places where quartzite and slate interfinger, they are clearly facies equivalents.

Most of the Buffalo Hump Formation is slate. The thinly cleaved slate is black, gray, brown, and green. Bedding is usually absent. Where visible, beds are 0.5-4 cm thick. In a few outcrops, the slate has light-brown diagenetic reduction spots as much as 7 cm long. Chlorite augen as long as 5 mm are present in slate in a quarry in NW1/4 sec. 19, T. 31 N., R. 39 E. In places, the slate contains pyrite cubes as wide as 1 cm. Veins of white quartz, 4-30 cm thick, occur parallel to cleavage and white quartz is present in lenses as much as 2 m long and 30 cm thick.

The slate is composed mostly of quartz and feldspar grains as much as 0.03 mm long and muscovite. Other minerals are present in minor quantities: allanite, chlorite, garnet, hematite, leucoxene, magnetite, and sphene. Some rocks contain euhedral magnetite porphyroblasts as much as 0.3 mm long.

White, pale-gray, light-brown, reddish, maroon, and black quartzite, and quartz siltite are present in layers that range from 1 to 195 m thick; some of the layers have slaty interbeds. Typically, the quartzite is fine to medium grained. In places, it grades to medium- and coarse-grained quartzite, quartz grit, and quartz cobble conglomerate with angular to subrounded clasts as much as 30 cm. In NE1/4 sec. 3, T. 30 N., R. 38 E., the rock is a stretched pebble conglomerate with clasts elongate as much as 4:1 (fig. 7). Clasts are flattened parallel to a pervasive cleavage; long axes of clasts plunge north-northeast at low angles.

The quartzite is 85-95 percent quartz with minor hematite, leucoxene, magnetite, white mica, stilpnomelane, and tourmaline (clasts). Most quartz grains exhibit evidence of strain: undulatory extinction, polygonization, recrystallization, flattening of grains, pressure shadows, and deformation lamellae. Sedimentary grain shapes are commonly distorted by deformation. Apparently, the grains were originally angular to subrounded. The matrix of very fine grained quartz, feldspar, and micaceous minerals constitutes 5-10 percent of the rock. Feldspar grains, making up 1-2 percent of some samples, are nearly completely sericitized.

The thickest section of the Buffalo Hump Formation, in sec. 17, T. 31 N., R. 39 E., determined by geometric methods to be 745 m thick, consists of alternating slate and quartzite (fig. 8). Thicknesses of quartzite layers vary greatly. For example, the stratigraphic section across sec. 24, T. 31 N., R. 38 E. is 390 m thick and contains a quartzite layer 55 m thick. This quartzite layer thins abruptly to 25 m to the southwest and thickens to 165 m within 0.5 km to the northeast. The thickest quartzite layer is more than 300 m thick in E1/2 sec. 5, T. 31 N., R. 39 E.

The pelitic and quartzose sedimentary rocks of the Buffalo Hump Formation contrast strongly with the underlying carbonate rocks of the Stensgar Dolomite; the difference indicates a drastic change in sedimentary environment from the underlying rock unit. Most of the Stensgar in the study area is overlain by slate of the Buffalo Hump; the slate indicates a continuation of quiet-water deposition and also an opening up of the basin to outside sediment sources. The quartzite, grit, and conglomerate beds of the Buffalo Hump, some with coarse festoon bedding, indicate a much higher energy depositional environment than one that prevailed during deposition of most of the underlying Deer Trail Group. The source of the quartz-rich detritus is not known.

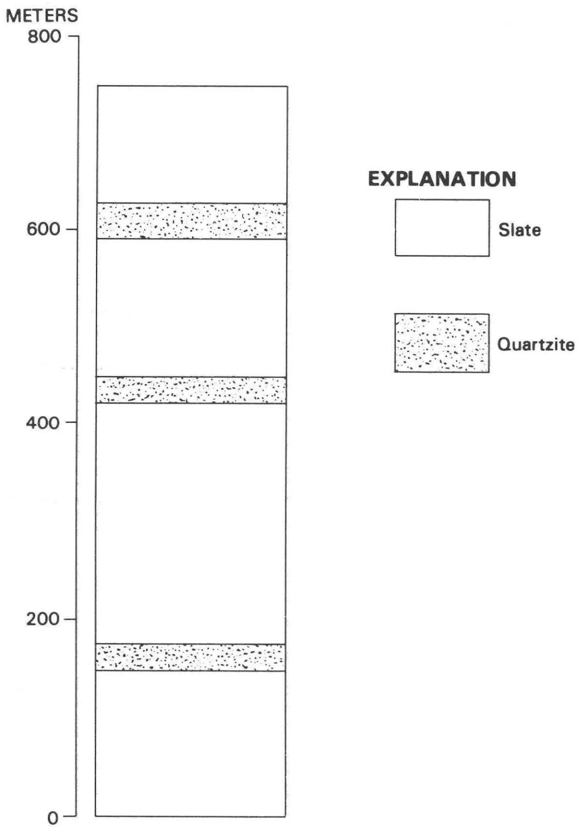


Figure 8. Stratigraphic section of the Buffalo Hump Formation in sec. 17, T. 31 N., R. 39 E.

Age and Correlation

Weaver (1920, p. 50) and Jones (1928) considered the Deer Trail Group to be Paleozoic. Bennett (1941, p. 8) placed the Deer Trail Group in the Proterozoic, on the basis of Early Cambrian fossils found in the limestone above the Addy Quartzite. Subsequent workers (Becraft and Weis, 1963; Campbell and Raup, 1964) similarly assigned the unit a Proterozoic age. Such an age assignment is consistent with the Proterozoic radiometric dates for the volcanic rocks member of the Huckleberry Formation (Miller and others, 1973), which overlies the Deer Trail Group. By this evidence, the Deer Trail Group is older than 929-734 Ma. (See section on age of Windermere Group). Therefore, in this report, these rocks are considered Middle Proterozoic in age following the usage of Harrison and Peterman (1982).

Campbell and Loofbourow (1962, p. F28) first correlated the Deer Trail Group with the Priest River Group of Park and Cannon (1943, p. 6) in the Metaline area to the northeast. Weis (1959) suggested correlation of all these rocks with the Belt Series (Belt Supergroup). Becraft and Weis (1963, p. 15-17) used both of these correlations. Miller and Clark (1975, p. 18-24) discussed the relation between the Belt Supergroup and the Deer Trail Group in the Chewelah-Loon Lake area and attempted to correlate the formations unit by unit. Another detailed correlation of the formations of the Deer Trail Group and the units in the Belt Supergroup of the Clark Fork quadrangle, Idaho and Montana (fig. 1), was suggested by McMechan (1981, fig. 10). The correlation schemes are shown in figure 9.

Miller and Clark (1975, fig. 7) made unit correlations between the Deer Trail Group and the Belt Supergroup of the nearby Chewelah-Loon Lake area. They presented two correlation schemes owing to the uncertainties in the stratigraphy of the Striped Peak Formation in the Chewelah-Loon Lake area. Their correlations are lithologically convincing from the Togo Formation through the McHale Slate (figs. 9A, B). Slate of the Togo Formation is correlated with the upper part of the Wallace Formation; the uppermost quartzite of the Togo Formation, not present in the Stensgar Mountain quadrangle, with member a of the Striped Peak Formation; Edna Dolomite, with member b of the Striped Peak; McHale Slate, with member c of the Striped Peak. If their section of the Striped Peak Formation has not undergone thrusting, the Stensgar Dolomite may be stratigraphically equivalent to Member d, a maroon and red siltite and argillite unit (fig. 10), assuming a facies change. If their section is faulted, the Stensgar Dolomite may correlate with an upper dolomite member in the Striped Peak Formation (fig. 9B).

Correlation of the Deer Trail Group with the Belt units of the Clark Fork quadrangle (Harrison and Jobin, 1963; fig. 1) by Miller and Clark (1975, fig. 7) equate slate of the Togo Formation with the upper part of the Wallace Formation, the uppermost quartzite of the

Togo Formation (quartzite missing in Stensgar Mountain quadrangle) with unit 1 of the Striped Peak Formation, Edna Dolomite with unit 2, and McHale Slate with unit 3 (figs. 9A, B). The Stensgar Dolomite may be stratigraphically equivalent to unit 4, a maroon and red argillite and quartzite, if this unit underwent a facies change.

Thicknesses of the Togo Formation through the Stensgar Dolomite, corrected for stratigraphic thinning owing to compressional deformation (see table 1), are not much greater than the thicknesses of the correlative Belt units in the Chewelah-Loon Lake area. Aggregate thicknesses of these units are comparable: more than 3,450 m for the Deer Trail Group and 3,380 m for the Belt Supergroup.

McMechan (1981, fig. 10) offered a different correlation of the Deer Trail Group with the upper part of the Belt Supergroup in the Clark Fork quadrangle (fig. 1). Her correlation suggests that the Togo Formation, Edna Dolomite, and McHale Slate are stratigraphically equivalent to the Wallace Formation and Unit 1 of the Striped Peak Formation (fig. 9C). The Stensgar Dolomite is correlated with unit 2.

Inclusion of the Buffalo Hump Formation in the Deer Trail Group can be questioned on the basis of stratigraphic relations. The first major erosional unconformity in the section in the study area appears to cut down through the basal slate of the Buffalo Hump Formation to the top of the Stensgar Dolomite. Quartzite occurs above the unconformity in places and marks the change in sedimentation pattern from quiet-water, low-energy environments to widespread high-energy environments mostly through the middle and upper part of the Buffalo Hump Formation. The quartzite and slate above the unconformity may be more closely related to the sedimentation pattern of the basal conglomerate in the overlying Windermere Group. However, the stratigraphic reassignment of the Buffalo Hump Formation to the Windermere Group is not clearly justified at this time.

Miller and Clark (1975) did not map the Buffalo Hump Formation in the Chewelah-Loon Lake area. In one correlation (Miller and Clark, 1975, fig. 7A), they suggested that the Buffalo Hump either was not deposited or was eroded away before deposition of the Addy Quartzite. In another correlation (Miller and Clark, 1975, fig. 7B), they suggested that the Buffalo Hump is stratigraphically equivalent to an uppermost argillite and siltite member of the Striped Peak Formation. McMechan (1981, fig. 10) suggested that the Buffalo Hump correlates with units 3 and 4 of the Striped Peak Formation in the Clark Fork quadrangle.

Windermere Group

Introduction

Strata of the Windermere Group were first named the Windermere Series by Walker (1934, p. 13-20) from the Windermere map area, southeastern British Columbia. The Windermere Series was recognized farther to the southwest in British Columbia in the Nelson map area by Rice (1941, p. 14-24) and in the Salmo map area by Walker (1926, p. 3-10) and Little (1950). These rocks were later referred to as the Windermere System by Reesor (1957, p. 158-162). Becraft and Weis (1963, p. 15-18) suggested correlating the rocks above the Deer Trail Group with the Windermere System. Miller and others (1973, p. 3723), who extended the name into Washington and Idaho, called the rocks the Windermere Group. In the Stensgar Mountain quadrangle, the Windermere Group consists of the Huckleberry Formation and the overlying Monk Formation. The Windermere is considered Middle(?) and Late Proterozoic in age.

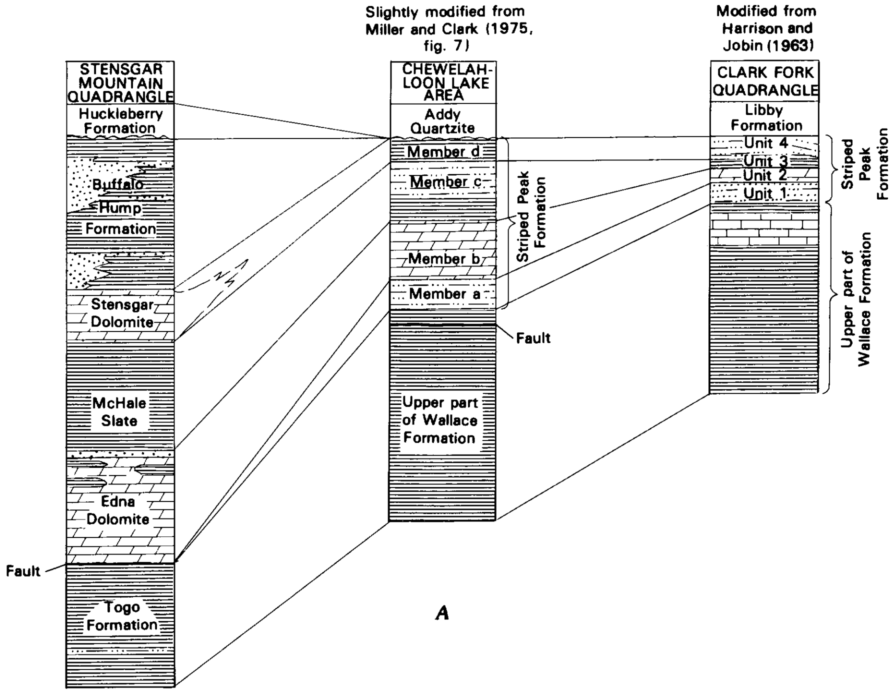
Huckleberry Formation

The Huckleberry Formation consists of a basal conglomerate member and an upper volcanic rocks member. The conglomerate was first recognized by Jones (1928, p. 115) and named by Bennett (1941, p. 8) for exposures on Huckleberry Mountain. It occurs in a band trending north-northeast across the mountain at about 15^{o} to the north-trending segment of the ridge line. The conglomerate member is situated relatively high topographically, but it is not a ridge-forming unit and outcrops of it are usually poor.

The conglomerate is typically a black, gray, or green-gray rock with phyllitic aspects. The clasts, which constitute 20-40 percent of the rock, are flat pebbles predominantly of black, brown, and light-gray argillite and siltstone as much as 4 cm long. Less abundant ovoid cavities probably originated from solution of carbonate clasts. Subrounded to angular pebbles of light-gray to brown and green quartz and quartzite as much as 6 cm long are much less common. In places, cubic cavities mark sites of dissolved pyrite crystals. Most of the rock has an argillic matrix; some of it has a sandy matrix, and in a few places the rock is sandstone. A 1-m-thick bed of dark-gray conglomerate with subrounded dolomite clasts as much as 15 cm long in a matrix of dolomite and argillite occurs in the NW1/4 sec. 24, T. 31 N., R. 38 E. near the base of the unit. In the conglomerate member as a whole, approximate proportions of clast types appear to be argillite: dolomite: siltite: quartz and quartzite=16:2:1:1. In four samples from the member on Huckleberry Mountain, Aalto (1971, p. 763) found the following average modal proportions: 39 percent clasts consisting of: undulose quartz, 16 percent; carbonate, 10 percent; mudstone, 4 percent; chert, 4 percent; quartzite, 3 percent; polycrystalline quartz, 1 percent; opaque minerals, 1 percent; sandstone, trace; plagioclase, trace; untwinned potassium feldspar, trace; and 61 percent fine matrix and detrital material. These clast proportions differ greatly from clast proportions in stratigraphically equivalent beds to the northeast (Aalto, 1971, p. 762). From this information one can conclude that the conglomerate member most likely had local sediment sources.

The conglomerate consists mainly of granular quartz, both as clasts and as the major constituent of quartz siltite and quartzite clasts. The matrix is predominantly silt-sized quartz and muscovite. Other minerals present in minor amounts are biotite, chlorite, hematite, pyrite, pumpellyite, and tourmaline (clasts).

A slaty cleavage is well developed in the argillite matrix of the conglomerate. In places where clasts



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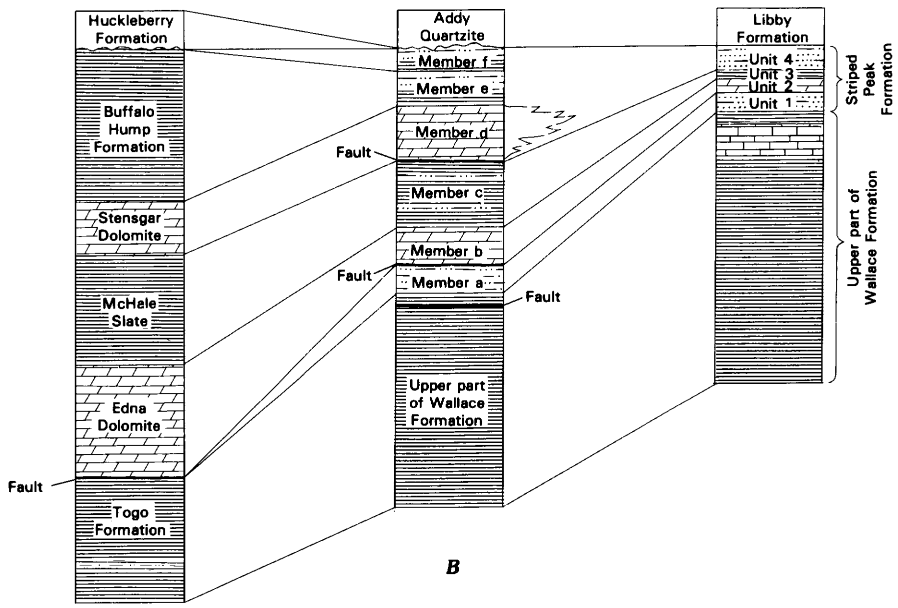

B

Figure 9. Correlation of formations of the Deer Trail Group with units in the Belt Supergroup. A and B, Alternative correlations from Miller and Clark (1975) because of uncertainties in the stratigraphy of the Striped Peak Formation in the Chewelah-Loon Lake area. C, Correlations in the Clark Fork quadrangle from McMechan (1981).

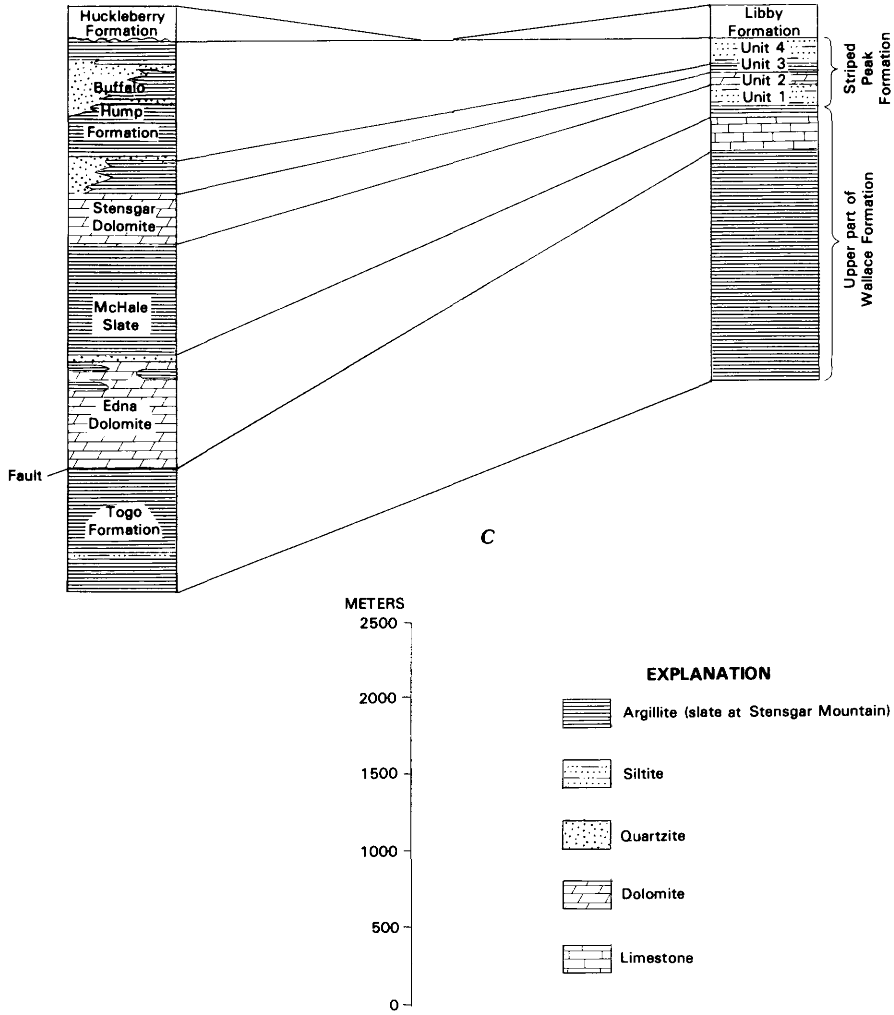

Figure 9. Continued.

are few, as in the southwest corner of the study area, the rock appears similar to slate of the Buffalo Hump Formation. One of the main differences between the two formations in that area is the character of the cleavage: anastomosing in the Huckleberry Formation and planar in the Buffalo Hump Formation.

Elsewhere in the Huckleberry Formation, the cleavage is emphasized by the many flat pebbles of argillite and the augen-shaped pebbles of quartzite and ovoid voids that are relicts of dissolved pebbles of dolomite; the pebbles are oriented with their long dimensions parallel to the cleavage. Very thin pebbles of slate, some about 0.1 mm thick, seem unlikely to have had the competence to remain as coherent clasts during deposition; they were probably thinned after deposition. Some of this thinning may have occurred during compaction. In sec. 29, T. 32 N., R. 39 E., flat pebbles are oriented parallel to apparent bedding; this bedding may be a sedimentary fabric enhanced during compaction. Aalto (1971, p. 760), however, noted that undeformed conglomerate lacked primary fabric, and, therefore, a tectonic origin of much of the apparent flattening of clasts is likely (see section on "Structure"). Tectonic foliation in stratigraphically equivalent rocks was also found to the northeast of the Stensgar Mountain quadrangle (Aalto, 1971, p. 758).

Apparent thickness of the conglomerate member ranges from 200 m in E1/2 sec. 29, T. 32 N., R. 39 E. to 500 m in NE1/4 sec. 26, T. 31 N., R. 38 E., this change suggests either a depositional thickening of the unit to the southwest, or greater erosion of the unit to the northeast prior to deposition of the overlying volcanic rocks member, or truncation of the base of the unit by the Stensgar Mountain thrust.

Stewart (1972) suggested that the conglomerate at the base of the Windermere Group lies along a belt of Late Proterozoic rocks extending from Alaska to Mexico. In his view, the basal conglomerate of the assemblage and the basaltic volcanic rocks in the Late Proterozoic sequence, including the volcanic rocks member of the Huckleberry Formation, may mark the initial phases of splitting of a continental mass with the two pieces drifting apart.

The volcanic rocks member forms a high terrane of generally bold outcrops in a band crossing Huckleberry Mountain parallel to the conglomerate member. The volcanic rocks are massive, black, and dark gray with green tones that have suggested the name "green-

Table 1.—Stratigraphic thicknesses of formations of the Deer Trail and Windermere Groups, and Early Cambrian strata in the Stensgar Mountain quadrangle and part of the Belt Supergroup near Stensgar Mountain area

[Thickness shown in meters]

	Stensgar Mountain quadrangle, this report	Stevens County (Bennett, 1941)	Magnesite belt (Campbell and Loofbourow, 1962)	Turtle Lake quadrangle (Becraft and Weis, 1963)	Hunters quadrangle Calculated thicknesses, based on map by Campbell and Raup (1964)	Chewelah-Loon Lake area (Miller and Clark, 1975)	Apparent thicknesses, this report	Estimated original thicknesses, this report		Maximum thicknesses of correlative strata Chewelah-Loon Lake area (Miller and Clark, 1975)
	Old Dominion Limestone	Thickness not given	Thickness not given	1,340	2,300	1,500	>1,500	No estimate		>750
	Addy Quartzite	900-1,500	Thickness not given	1,200	1,700	1,200	1,100	1,100(?)		1,200
Windermere Group	Monk Formation	Not recog-nized	Not recog-nized	Not present	Not present	≥250	500	950	Windermere Group	≥250
Windermere Group	Huckleberry Formation-volcanic rock member	915(max.)	900	Not present	900(max.)	≥350	750-1,100	750-1,100	Windermere Group	≥350
Windermere Group	Huckleberry Formation conglomerate member	915(max.)	450	Not present	900	Thickness not given	200-500	400-1,000	Windermere Group	Minor
	Windermere Group (total)	Not recognized	1,350	Not present	1,800	>600	1,750-1,800	2,450-2,550		≥600
Deer Trail Group	Buffalo Hump Formation	Thickness not given	Thickness not given	Not present	800	≥900	>750	>1,400		Not present
Deer Trail Group	Stensgar Dolomite	90-215	90-120(avg.) 235(max.)	120	350	150-300	>500	>500	Striped Peak Formation	250
Deer Trail Group	McHale Slate	Thickness not given	300-450	230-730	750	≤600	>1,500	750-1,000	Striped Peak Formation	600
Deer Trail Group	Edna Dolomite	Thickness not given	450-750	400-750	930	≥250	>700	>1,050	Striped Peak Formation	530
Deer Trail Group	Togo Formation	Thickness not given	>1,200	6,100	Not determined	900	>600	>1,150	Wallace Formation	1,800-2,000
	Deer Trail Group (total)	>9,150	>2,050	>6,850	>2,830	>2,800	>4,050	>4,850	Wallace Formation	3,180-3,380

stone" applied to these rocks. Most of the rock unit is little deformed; microscopic textures of the original basalt and tuff are preserved. Typically, the rocks are cut by numerous joints, which appear to be randomly oriented. In the places where a well-developed cleavage is present, the rock resembles a phyllite. Low-grade metamorphism is pervasive and no layering or pillows are preserved.

In the quadrangle, the volcanic rocks member consists of metabasalt flows and basaltic tuff. The flows are aphyric to porphyritic with as much as 10 percent phenocrysts; the phenocrysts are chiefly medium-grained plagioclase (An_{45} minimum) as much as 2 mm long in a very fine grained groundmass with grains as much as 0.2 mm long. The plagioclase, in both the phenocrysts and the groundmass, is much altered to epidote and clay minerals and is replaced by chlorite, pumpellyite, and sphene. Mafic minerals, which once composed 50 percent of the rock, are completely altered to chlorite, to sphene that is partly altered to leucoxene, and to tremolite. A minor amount of magnetite, partly altered to hematite, is present. Some of the rock has amygdules of chlorite and epidote.

One sample of the metatuff is composed of 50 percent volcanic rock fragments containing abundant felty plagioclase laths as much as 0.1 mm long, actinolite, chlorite, sphene that is partly altered to leucoxene, and minor biotite and pumpellyite. The lithic fragments are in a matrix of fragmental plagioclase laths, partly to completely replaced by epidote, and of cuspate shards of basaltic glass altered to masses of green chlorite that are outlined by zones rich in sphene and leucoxene.

Chemical analyses of samples from the volcanic rocks member, given in Miller and Clark (1975, table 1) and in Devlin and others (1985, table 3), are similar to "normal" tholeiitic basalt (Nockolds, 1954, table 7).

One sample of phyllitic rock consists mostly of quartz and chlorite, and minor amounts of hematite, leucoxene, magnetite, muscovite, and sphene. The rock may once have had a basaltic composition like the rest of the volcanic rocks member including the tuffaceous beds. If so, the present high proportion of quartz suggests that the rock has been silicified.

Apparent thickness of the volcanic rocks member ranges from 765 m in sec. 23, T. 31 N., R. 38 E., to 1,120 m in sec. 29, T. 32 N., R. 39 E. The narrower outcrop width of the member to the southwest is probably due to erosion and subsequent burial of these rocks beneath the Addy Quartzite, because the entire Huckleberry Formation is truncated and wedges out beneath the quartzite in the adjacent Hunters quadrangle (Campbell and Raup, 1964).

The Huckleberry Formation has been correlated with the Shedroof Conglomerate and Leola Volcanics, that overlie the Priest River Group in the Metaline area (Campbell and Loofbourow, 1962, p. F27-F28). Park and Cannon (1943, p. 7-11) had previously correlated the Shedroof Conglomerate and Leola Volcanics with similar rocks, called the Irene Conglomerate and Irene Volcanics, in southern British Columbia. These rocks are subdivisions of Daly's (1912, p. 141-147) Summit Series. Becraft and Weis (1963, p. 15-18) also suggested correlation of the Huckleberry Formation with the Shedroof Conglomerate and Leola Volcanics, and with the Toby Conglomerate (formerly the Irene Conglomerate) and Irene Volcanics. Aalto (1971) considered the conglomerate member of the Huckleberry Formation to be stratigraphically equivalent to the Toby Conglomerate and suggested that these rocks are the products of glacial marine sedimentation.

Six samples of fresh-looking basalt from the volcanic rocks member of the Huckleberry Formation were dated by Miller and others (1973) using the K-Ar method. Acceptable ages range from 929 to 734 Ma (recalculated using current IUGS constants in Dalrymple, 1979), or Middle and Late Proterozoic. Inasmuch as the rocks yielding these ages underwent low-grade metamorphism, the ages may reflect a metamorphic event. Assuming that the metamorphic event occurred soon after emplacement or without change of the K/Ar ratios, Miller and others (1973) interpret these ages to indicate the approximate time of extrusion. After study of the same exposures of the volcanic rocks member that Miller and others (1973) dated, Devlin and others (1985) expressed doubt that the K/Ar ages are indicative of the time of extrusion, but they did not demonstrate that the K/Ar ages are in error. U-Pb ages of zircons from gneissic granite in the Sifton and Deserters Range, north-central British Columbia, indicate a maximum age of 728 Ma for the Windermere Group in that region (Evenchick and others, 1984). This age is younger than the ages of Miller and others (1973), but it is possible that Windermere sedimentation began later in north-central British Columbia than in northeastern Washington. The recalculated ages of Miller and others (1973) are adopted in this report. The conglomerate member is clearly older, but probably not much older, than the volcanic rocks member; both members are considered here to be Middle(?) and(or) Late Proterozoic in age. Thus the age of the Huckleberry Formation is here regarded as Middle(?) and(or) Late Proterozoic.

Gabbro and Metabasalt Dikes

Mafic dikes 5-550 m thick intruding the Deer Trail Group and the conglomerate member of the Huckleberry Formation commonly form bold outcrops. The dikes are fine to medium grained, black and dark gray with green tones. Most rocks consist of plagioclase and mafic minerals in approximately equal proportions.

The metabasalt dikes are generally aphyric. Twinned plagioclase laths as much as 0.5 mm long are mostly replaced by epidote, sericite, albite, carbonate, biotite, hornblende, pumpellyite, sphene, and quartz. Most of the mafic minerals are now altered to biotite, chlorite, pumpellyite, and sphene. Minor amounts of ilmenite partly altered to sphene and leucoxene, magnetite partly altered to hematite, rutile in quartz, and talc are present in some dikes. Porphyroblasts of euhedral pyrite as much as 0.5 mm long cut across cleavage in phyllitic parts of some dikes.

The gabbro dikes, with grains 1-3 mm long, appear to have been originally composed of plagioclase (An_{46} minimum, 50-70 percent), clinopyroxene (30-50 percent), and ilmenite (2 percent). One sample has 5 percent olivine. The plagioclase, which was altered to white mica, epidote, and clays, is replaced by biotite, pumpellyite, and sphene. The clinopyroxene was deuterically altered to brown, green, or colorless hornblende, and later was altered to epidote, chlorite, and tremolite. Hornblende was altered to biotite, chlorite, and sphene. Ilmenite was altered to leucoxene and sphene. The minor quartz in the rock is a primary constituent because it contains trace amounts of rutile. Hematite and stilpnomelane are also present.

The dikes resemble the volcanic rocks member of the Huckleberry Formation in composition and in having numerous randomly oriented fractures and local zones of phyllite. Following the suggestion of Campbell and Loofbourow (1962, p. F25-F26), these dikes are interpreted as feeders of the volcanic rocks member and are, therefore, also Middle(?) and(or) Late Proterozoic in age.

Monk Formation

The Monk Formation was named by Daly (1912, p. 147-150) for a section of slate, phyllite, schist, and conglomerate in his Summit Series in southeastern British Columbia. The formation is recognized in the Metaline area (Park and Cannon, 1943, p. 11) and in the Chewelah-Loon Lake area (Miller and Clark, 1975, p. 26-27) where dolomite occurs in the formation. The Monk Formation is present in the Stensgar Mountain quadrangle in the N1/2 sec. 15, T. 31 N., R. 39 E., just east of Carrs Corner. The section of which these strata are a part extends into the adjacent Waitts Lake quadrangle.

The Monk Formation in the study area from west to east consists of 30 m of dolomite that is faulted on the west against the Togo Formation, 30 m of conglomerate, and almost 60 m of slate. It is contiguous on the east in the Waitts Lake quadrangle with the Huckleberry Formation (F.K. Miller, unpublished map, 1978); this relation indicates that the steeply dipping Monk Formation is right side up with its top to the west.

The fine-grained dolomite, mottled black and gray, has beds as much as 60 cm thick. The mineral dolomite constitutes 95 percent of the rock, and it has grains as long as 0.05 mm. The rock appears intensely brecciated and contains irregular veinlets of micro- to cryptocrystalline pyrite (4 percent) which give the rock its gray color. Talc and quartz are also present.

The brown conglomerate contains 25 percent pebbles and cobbles as large as 15 cm. Most of the clasts are flat fragments of shale; less abundant are rounded to subrounded clasts of dolomite, quartzite, and arkosic sandstone containing carbonate cement (fig. 10). Some of the dolomite pebbles were dissolved, leaving ovoid cavities in the matrix. Most of the clasts are flattened parallel to a well-developed cleavage, which may be subparallel to bedding. Some shale pebbles are less than 1 mm thick. Most pebbles, the shale pebbles more than those of the dolomite and quartzite, are elongated as much as 5:1 in the plane of cleavage. A few quartzite cobbles are well-rounded and do not appear to be deformed. The matrix is siltstone and fine grained sandstone. It consists mostly of quartz, muscovite, and carbonate, and minor biotite, hematite, leucoxene, and plagioclase (clasts, An_{46}).

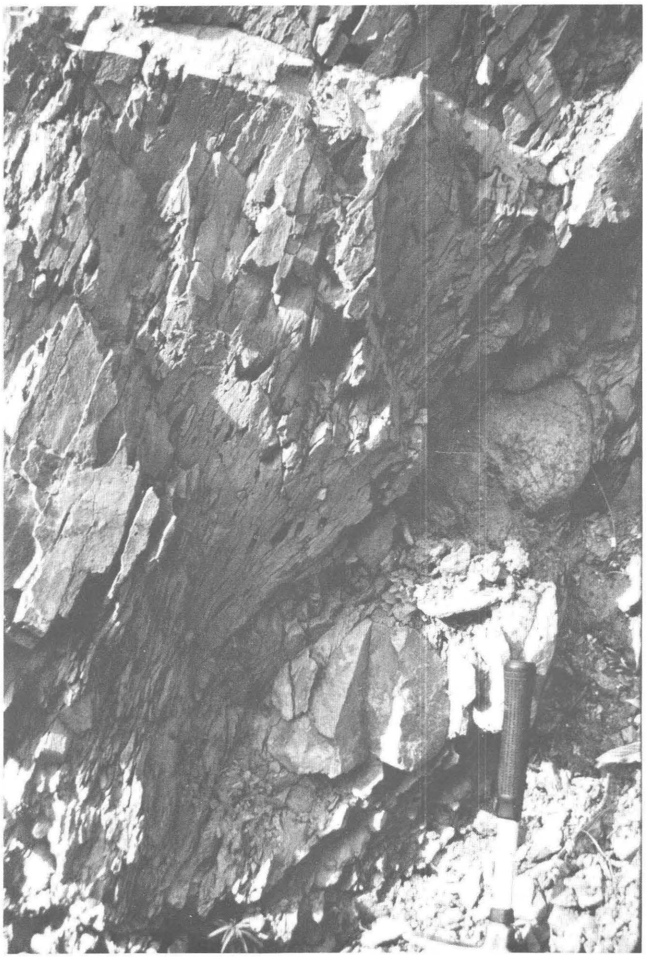

Figure 10. Conglomerate in the Monk Formation, NE1/4 sec. 15, T. 31 N., R. 39 E. Handle of hammer is 30 cm.

The conglomerate grades rapidly downwards (eastward) into dark gray slate with well developed cleavage.

The combined sections of the Monk Formation in the Stensgar Mountain and Waitts Lake quadrangles may be as much as 490 m thick; this thickness is based on attitudes observed in the field. The contact between the dolomite and conglomerate appears to be a shear zone, but no stratigraphic duplication is evident in the section.

The absence or lack of exposure of the Monk Formation above the Huckleberry Formation on Huckleberry Mountain can be explained by an angular unconformity between the Addy Quartzite and the Huckleberry Formation. The attitudes of bedding in the Addy Quartzite differ by 8°, on the average, from bedding attitudes in the underlying rocks, principally in the Deer Trail Group, and may be a measure of the angular discordance. In addition, the fact that the Addy Quartzite truncates the Huckleberry Formation in the Hunters quadrangle (Campbell and Raup, 1964) indicates that an angular unconformity at the base of the Addy is likely.

The Monk Formation is younger than the volcanic rocks member of the Huckleberry Formation, but by this evidence it could be either Middle or Late Proterozoic. Park and Cannon (1943, p. 11) report a gradational contact between the Monk Formation and Cambrian quartzite in the Metaline quadrangle, but Burmester and Miller (1983) mapped the Three Sisters Formation of Walker (1934) as having a gradational contact with the underlying Monk Formation; the Three Sisters Formation may also lie unconformably beneath the Addy Quartzite on Huckleberry Mountain. In this report, the Monk Formation is considered Late Proterozoic in accordance with the age designation of Burmester and Miller (1983).

Paleozoic Rocks

Addy Quartzite

The Addy Quartzite was named by Weaver (1920, p. 61-63). The formation is exposed along most of the crest of Huckleberry Mountain and Lane Mountain, where quartzite beds form bold outcrops. West-facing dip slopes are covered with deep quartzite regolith.

The unit is typically fine- to medium-grained white quartzite, stained brown by iron oxides (fig. 11). Beds range in thickness from 1 mm to 2 m, even in a single outcrop. Other lithologies are present in certain parts of the formation. In the northeast corner of sec. 2, T. 31 N., R. 38 E., the unit consists of quartzite with abundant hematitic interstitial material in irregular spots and gray quartzite interlayered with purple quartzite beds and minor limestone. The rock in this area also contains coarse sand-sized quartz grains and rounded quartz pebbles as much as 1 cm in diameter. This conglomeratic facies may be continuous with a zone of relatively coarse-grained quartzite located in the center of sec. 11, T. 31 N., R. 38 E., as these coarse-grained rocks are approximately along strike from one another. A lens of quartz pebble conglomerate with well-rounded clasts as much as 8 cm in diameter occurs near the base of the formation in SE1/4 sec. 22, T. 31 N., R. 38 E. near the boundary with section 23. Black argillite is present along the contact with the Huckleberry Formation in this area. Black thin-bedded to laminated siltstone with argillaceous beds is present near the base of the formation in SE1/4 sec. 11, same township, near the boundary with section 12. Also near the base of the formation in SW1/4 sec. 1, same township, the quartzite is characterized by complex white, gray, pink, and black crossbeds and isoclinal folds of bedding, probably owing to soft-sediment deformation. On Lane Mountain, the Addy contains a thick zone of tightly cemented, thick-bedded quartzite. Near the north end of the mountain in the study area, some of the rock is mottled bright yellow and is very friable. This part of the formation contains at least one zone, 5 m thick, of silvery-gray and brown, crumbly micaceous quartzite with muscovite grains of probable sedimentary origin as much as 1 mm across. A gritty conglomeratic basal zone within the Addy Quartzite, noted by Bennett (1941, p. 9), was not seen in the Stensgar Mountain quadrangle.

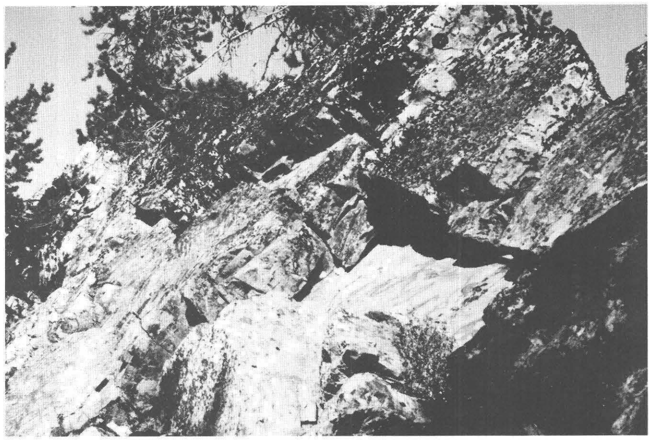

Figure 11. The Addy Quartzite on Huckleberry Mountain, W1/2 sec. 1, T. 31 N., R. 38 E.; near base of formation. Thick bed in center of photograph is 1 m thick. Bedding dips to left.

Most of the quartzite is 95-99 percent quartz with grains that have rounded clastic outlines or that form a polygonal mosaic of grains. The matrix consists of silt-sized quartz clasts with minor chlorite, clay, epidote, hematite, leucoxene, magnetite, stilpnomelane, and tourmaline (clasts). Some of the quartz clasts contain inclusions of zircon and apatite. The orange color of some of the quartzite on Lane Mountain is due to stilpnomelane in the matrix. The yellow color of the mottled friable beds on the same mountain is due to a mixture of chlorite, epidote, sphene, and stilpnomelane.

The Addy Quartzite in secs. 10, 11, and 14, T. 31 N., R. 38 E. is about 1,100 m thick. This thickness is only slightly thinner than that for the formation in the Turtle Lake quadrangle (Becraft and Weis, 1963, p. 12) and near Addy (Miller and Clark, 1975, p. 28), but it is substantially thinner than the formation in the Hunters quadrangle (Campbell and Raup, 1964; see table 1).

The Addy Quartzite is considered Early Cambrian in age based on trilobites found at two localities in the vicinity of Addy (Okulitch, 1951; Miller and Clark, 1975, p. 28).

Campbell and Loofbourow (1962, p. F28) correlated the Addy Quartzite with the Gypsy Quartzite in the Metaline area (Park and Cannon, 1943, p. 13-15). Later Becraft and Weis (1963, p. 15-18) extended this correlation to the Quartzite Range Formation in the Salmo area (Walker, 1934; Little, 1950) and the Hamill Series in the Nelson area (Reesor, 1957) in southern British Columbia.

Old Dominion Limestone of Weaver (1920)

The Old Dominion Limestone was named by Weaver (1920, p. 66-68), presumably for exposures on Old Dominion Mountain east of Colville (fig. 1), and the name was used by subsequent workers, Campbell and Loofbourow (1962, p. F25); Becraft and Weis (1963, p. 12-13); and Campbell and Raup (1964). The unit is present in the northwest corner of the Stensgar Mountain quadrangle where it is very poorly exposed. The Old Dominion was previously mapped as conformably overlying the Addy Quartzite (Weaver, 1920; Campbell and Raup, 1964). In this study, however, the contact between the two units is interpreted as a fault because the contact is steeper than the bedding, appears to truncate the upper part of the Addy Quartzite, and is continuous with a fault drawn along the eastern contact of the Old Dominion in the adjacent Dunn Mountain quadrangle (fig. 1) by Miller and Yates (1976).

The limestone is very fine grained, light to medium blue gray with beds less than 2 cm thick. Some of the rock is silty and breaks into thin plates along shaly beds. Some of the limestone float contains ellipsoidal limestone bodies as much as 2 cm long. In thin section, the calcite grains are generally less than 0.01 mm long with lenses of nearly cryptocrystalline calcite and lenses of calcite with grains as much as 0.05 mm long. Siliceous clasts include silt-sized quartz and feldspar, sand-sized grains of chert, or possibly silicified fossil fragments, and nodules of silty limestone. Minor amounts of hematite and magnetite are present.

A section of these strata about 1,500 m thick is present in the northwestern part of the quadrangle where the section is incomplete. Campbell and Raup (1964) showed approximately 2,300 m of carbonate rocks above the Addy Quartzite in the Hunters quadrangle. On the basis of regional relations, the Old Dominion presumably conformably overlies the Addy Quartzite.

Bennett (1941, p. 9), who noted the occurrence of the Early Cambrian brachiopod Kutorgina cingulata (Billings) in carbonate rocks correlated with the Old Dominion Limestone above the Addy Quartzite, designated the entire formation as Early Cambrian in age. Miller and Clark (1975, p. 29-30) suggested a correlation of at least part of the Old Dominion Limestone with the Metaline Formation of Middle and Late(?) Cambrian and Early Orodovician(?) age (Repetski, 1978) in the Chewelah-Loon Lake area (same as the Metaline Limestone of Park and Cannon, 1943, p. 17-19). In this report, although the age of the Old Dominion Limestone is considered Early Cambrian, it is regarded as younger than the Addy Quartzite and may include Middle Cambrian and younger strata.

The Cambrian carbonate rocks in the Kettle Falls and Hunters quadrangles and on southern Old Dominion Mountain east of Colville (fig. 1), were studied by Mills (1977, p. 19-20), who assigned each of these sections to the Metaline. Snook and others (1981, p. 4) mentioned a facies change between the Metaline and the Old Dominion Limestone. The Maitlen Phyllite, which underlies the Metaline elsewhere, was not found in the study area and could have been faulted out between the Addy Quartzite and Old Dominion Limestone.

Mesozoic Rocks

Cataclasite

A poorly-exposed zone of cataclasite as much as 90 m thick is present along the boundary between secs. 22 and 27, T. 31 N., R. 39 E. The rock is a dark gray to black quartzite breccia with lenticular fragments of quartzite as much as 10 cm long. The breccia locally has a pale gray to yellow, fine- to medium-grained matrix of epidote, tremolite, muscovite, chlorite, and sphene. The origin of the quartzite is not known; although it is presently adjacent to the Togo Formation, it is lithologically similar to the Addy Quartzite. The cataclasite, which is next to the Lane Mountain thrust, probably records movement on the thrust. The age of the cataclasite, therefore, is probably Mesozoic and the same age as the Lane Mountain thrust (see section on "Structure").

Fault Breccia

A steep north-northeast-striking zone of fault breccia 3.5 km long extends from NE1/4 sec. 36, T. 31 N., R. 38 E. to SE1/4 sec. 19, T. 31 N., R. 39 E. The zone, which has a maximum width of 90 m and contains slivers of dolomite and slate 15 m or more wide that are elongate subparallel to the zone. Identifying the formations from which each dolomite sliver originated is difficult. One sliver of dolomite is clearly from the Edna Dolomite because it consists of rusty-brown, thick-bedded to massive dolomite. The slate may have come from the McHale Slate, Buffalo Hump Formation, or Togo Formation. The fault breccia contains disseminated pyrite in the dolomite, copper mineralized quartz veins, and barite veins. The breccia formed during Mesozoic faulting and is Late Triassic to Cretaceous in age.

Mesozoic or Tertiary dikes

Felsite

White to pale-brown, fine-grained felsic dikes as much as 8 m thick are present at four localities: intruding the Togo Formation in NE1/4 sec. 30, T. 31 N., R. 39 E.; intruding the Stensgar Dolomite in SE1/4 sec. 35, T. 31 N., R. 38 E. and in NW1/4 sec. 16, T. 31 N., R. 39 E.; and intruding slate of the Buffalo Hump Formation in NW1/4 sec. 36, T. 31 N., R. 38 E., near the Wells Fargo mine. The rocks contain 10-25 percent phenocrysts, chiefly plagioclase as much as 2 mm long and, in most dikes, subordinate quartz with hexagonal or rhombic outlines. The quartz phenocrysts are embayed by a quartzo-feldspathic groundmass with grains less than 0.3 mm long. The plagioclase is mostly altered to white mica and clays. Epidote, magnetite partly altered to hematite, and stilpnomelane are also present in minor amounts. Potassium feldspar was not identified, but it may be present in the groundmass. The composition of these dikes may be quartz latite.

The felsic dikes intrude the Deer Trail Group. Although they have not been found in younger rocks, they could be from one or more Mesozoic or Cenozoic intrusive episodes, such as those described in the Chewelah-Loon Lake area (Miller and Clark, 1975, p. 33-60) and Waitts Lake quadrangle (Miller and Engels, 1975). In this report, therefore, the dikes are considered to be Mesozoic or Tertiary in age.

Andesite

Two dikes of andesite at least 10 m thick intrude the Togo Formation in W1/2 sec. 22, T. 31 N., R. 39 E. The rocks have conspicuous elongate mafic phenocrysts as much as 6 mm long comprising 25 and 40 percent, respectively, of the dikes. In one dike, the phenocrysts are mostly hornblende; in the other dike, they are both hornblende and biotite. In both dikes, mafic minerals make up almost 50 percent of the rock, and the hornblende is altered to biotite and opaque oxides (ilmenite and(or) magnetite). The rest of the rock is calcic plagioclase slightly altered to epidote and white mica. Apatite is also present.

The low intensity of alteration of the andesite dikes suggests that they are younger than the metabasalt dikes related to the Huckleberry Formation and may be of Phanerozoic age. Miller and Clark (1975, p. 58-60) described mafic dikes of possible Eocene age intruding Proterozoic metasedimentary rocks and possibly Oligocene andesites in the Chewelah-Loon Lake area. The andesite dikes in the Stensgar Mountain quadrangle have similarities to and differences from the mafic dikes in the Chewelah-Loon Lake area. The dikes in the study area may also be of Tertiary age, even though they may not correlate with the mafic dikes to the east.

Quaternary Deposits

Glacial Deposits and Alluvium

Glacial deposits and alluvium are combined in one unit because they are difficult to separate in the field. Glacial deposits, which consist mostly of ground moraine, are especially abundant in the northeast corner of the quadrangle. At least one terminal moraine, now breached by postglacial erosion, crosses the north end of Klein Meadows (sec. 5, T. 31 N., R. 38 E.). The glacial deposits are characterized by subrounded boulders as much as 2 m in diameter of quartzite and granitic rocks. Because the granitic lithologies do not occur as bedrock in the Stensgar Mountain quadrangle, they must have been brought in by regional ice sheets. The glacial deposits that are widespread in the northeast corner of the study area are 8 km from the Colville River Valley in which Clark and Miller (1968, p. 3) postulated ice more than 600 m thick. A thickness of ice of that amount would be sufficient to account for most of the glacial till in the northeast corner of the Stensgar Mountain quadrangle. The age of the glacial deposits is presumably Pleistocene.

The alluvium consists of probable lake and swamp deposits, such as underlie Klein Meadows, and is present in beaver ponds along North Fork Deer Creek and Cedar Creek. Colluvium is present in the northwest corner of the quadrangle, and stream gravels and alluvial fan deposits, such as in sec. 4, T. 30 N., R. 39 E. The age of these deposits is presumably Pleistocene and Holocene.

STRUCTURE

Bedding in the Deer Trail Group in the study area has an average strike of N. 30° E. and dip of 45° NW. This attitude, which differs by 8° from the average dip of the Addy Quartzite (strike N. 19° E., dip 43° NW.), may be a measure of the small angular discordance owing to tilting of Proterozoic rocks before deposition of the Addy Quartzite.

The following description of the structure of the rocks in the Stensgar Mountain quadrangle depends upon the interpretation of the contact between the Windermere and Deer Trail Groups as a fault. On plate 1, the contact is shown as the Stensgar Mountain thrust. Definitive evidence bearing on the nature of this contact, however, is not present in the study area. Mapping by Miller and Yates (1976) and Miller (unpub., data, 1979) in the Addy quadrangle (fig. 1) strongly suggests that this contact at the base of the Windermere is a thrust. Part of this same contact in the Hunters quadrangle (fig. 1) was interpreted as a fault by Campbell and Raup (1964). Therefore, the basal contact of the Windermere Group in the Stensgar Mountain quadrangle is interpreted as a fault in this report.

The Deer Trail Group in the Stensgar Mountain quadrangle lies within a duplex thrust (Boyer and Elliott, 1982, fig. 12). The floor thrust is the Lane Mountain thrust and the roof thrust is the Stensgar Mountain thrust. Within the duplex, the Deer Trail Group comprises three thrust plates, each of which has a characteristic stratigraphy (fig. 12). Plate a consists of the Togo Formation. Plate b, which lies above plate a, consists of the Edna Dolomite through the Buffalo Hump Formation. Plate c, which lies above plate b, consists of the McHale Slate through the Buffalo Hump Formation. Steep northwest-striking faults that cut the thrust plates may be connecting splays between the roof and floor thrusts (Boyer and Elliott, 1982, fig. 7).

The Deer Trail Group has a pervasive slaty cleavage that strikes N. 8° E. and dips 66° W. on the average. The mean angle between the bedding and cleavage is 26°. In some rocks a well-developed cleavage also parallels the bedding. Near parallelism of some formational contacts and bedding with cleavage suggests either that these primary surfaces have been rotated toward the cleavage, or that bedding surfaces were fortuitously oriented so that the cleavage formed parallel to the bedding. Once spherical reduction spots of diagenetic origin in the Buffalo Hump Formation are now ellipsoidal and flattened in a plane subparallel to cleavage. The axes of the ellipsoids suggest that cleavage development accompanied shortening on the order of 70 percent at large angles to slaty cleavage and elongation on the order of 200 percent parallel to cleavage and to some microcrenulations. This conclusion is in accord with studies on the significance of slaty cleavage (Cloos, 1947; Ramsay, 1967, p. 177-182; Oertel, 1971; Siddans, 1972; Wood, 1974).

Using the graphic method outlined by Ramsay (1967, p. 129-132), the average attitude of bedding in the Deer Trail Group before cleavage formed can be determined. Assuming strains like the ones suggested

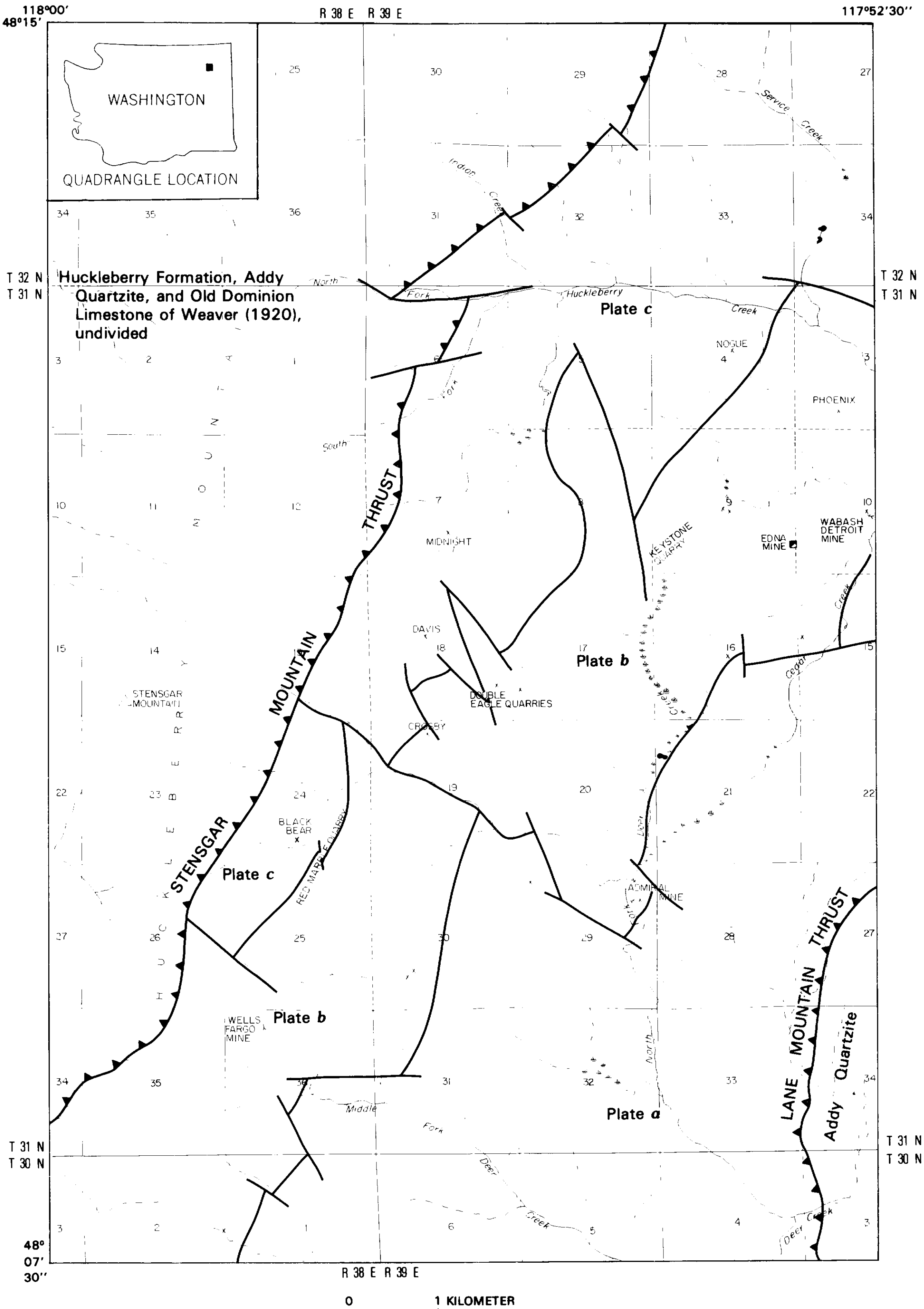

Figure 12. Structural plates a, b, and c, of the Deer Trail Group in Stensgar Mountain quadrangle, Wash. Plate a consists of the Togo Formation; plate b, the Edna Dolomite through the Buffalo Hump Formation; and plate c, the McHale Slate through the Buffalo Hump Formation. Northwest-striking faults cutting thrust plates may be connecting splays between roof and floor thrusts (sawteeth on upper plate). See plate 1 for details of faulting.

by the deformed reduction spots, the average bedding had a strike of N. 46° E. and a dip of 42° NW. before the deformation. The inference from this analysis is that the beds underwent significant tilting between Early Cambrian time and the time of formation of the cleavage.

Well-developed cleavages in the conglomerate member of the Huckleberry Formation and in the Monk Formation (Windermere Group units) are parallel to the slaty cleavage in the Deer Trail Group; this relation strongly suggests that these cleavages in the two groups are structurally identical fabric elements. The slaty cleavage is, therefore, younger than the Windermere Group. Both conglomerates in the Windermere Group contain clasts flattened in the plane of cleavage; this evidence substantiates the hypothesis of large shortening of these rocks at high angles to cleavage. Although much of the volcanic rocks member of the Huckleberry Formation appears undeformed, the member as a whole has undergone some deformation, as shown by the thinly cleaved phyllitic zones scattered throughout the unit.

Folds in bedding and cleavage generally parallel intersections of bedding and cleavage. The folds, which exhibit a wide variety of styles (concentric, similar, isoclinal, disharmonic, and chevron), are grouped into a dominant set plunging 38°, N. 12° W. on the average, and a subordinate set plunging steeply westward. Though the two sets of folds intersect locally, they do not show clear evidence of which set is the older one. The two sets may be conjugate.

The macroscopic anticline and syncline formed by the quartzite bed in the Togo Formation (pl. 1) plunges 30°, N. 20° W., subparallel to most of the minor folds in the Deer Trail Group. In addition, cleavage in the hinges of the large folds parallels slaty cleavage in the Togo slates. For these reasons the large folds are probably related to the same deformation in which the minor folds and slaty cleavage formed.

Minor folds of bedding found in the Addy Quartzite plunge at low angles north-northeast to northwest, like folds in the Deer Trail Group; this similarity suggests that these folds may be related to the same deformation in which slaty cleavage was formed in the older rocks. This evidence would date the deformation as Early Cambrian or later. Little evidence of deformation was found in the Old Dominion Limestone.

Orientation of the principal direction of shortening approximately west-northwest is consistent with a tentative correlation of this deformation with the episode of Late Triassic to Cretaceous tectonism deduced for the region (Yates and others, 1966; Rinehart and Fox, 1972, p. 72; Fox and others, 1977, p. 24). Mills and Nordstrom (1979, p. 201) studied rocks in northern Stevens County that underwent deformation similar to that of the rocks in the Stensgar Mountain quadrangle, and they concluded that some episodes of the deformation may be younger than Early Triassic and older than mid-Cretaceous.

The relative timing of thrusting and the penetrative deformation is suggested by the large anticline and syncline in the Togo Formation. Because the faults near the fold hinges are not folded, the penetrative deformation most likely occurred before some of the faulting, although possibly not before the earliest movement along the duplex thrust.

Steep fracture cleavages consist of narrow breccia zones a few millimeters wide and spaced 1-10 cm apart. These cleavages offset bedding and slaty cleavage as much as several centimeters. Bedding and slaty cleavage adjacent to the fracture cleavages are locally folded, generally in an open, concentric style. These structural elements postdate slaty cleavage and may have developed before thrusting ceased.

THICKNESSES OF THE FORMATIONS

Estimates of the original sedimentary thicknesses of the formations in the study area are subject to uncertainties arising from tectonically caused changes of the thicknesses. The presence of slaty cleavage implies distortions of formation thicknesses and must be considered in estimating stratigraphic thicknesses (Cloos, 1947, p. 910-911).

The significance of slaty cleavage has been the subject of geological research since 1815 (Siddans, 1972). Current views of slaty cleavage are summarized by Ramsay (1967, p. 177-182), Wood (1974), and Hobbs and others (1976, p. 231-252). The general conclusion of these reviews is that slaty cleavage is a flattening phenomenon; cleavage forms perpendicular, or approximately perpendicular, to the axis of maximum finite shortening (Ramsay, 1967, p. 180; Siddans, 1972, p. 219; Wood, 1974, p. 398-399; Hobbs and others, 1976, p. 252). This interpretation is consistent with the orientation of planes of flattening of reduction spots and deformed clasts in the Buffalo Hump Formation and deformed clasts in conglomerates of the Windermere Group. Strain data from slates suggest that the minimum amount of shortening required before cleavage appears is more than 30 percent (Cloos, 1947, p. 878; Oertel, 1971, p. 536; Wood, 1974, fig. 4). Well-developed slaty cleavage, such as is present in the study area, may reflect shortening perpendicular to the cleavage of more than 50 percent (Wood, 1974, p. 399).

The angle between bedding and cleavage must be considered in order to determine whether shortening perpendicular to cleavage has resulted in thinning or thickening of strata. The average angle between bedding and cleavage in the Stensgar Mountain quadrangle is 26°, as determined from contoured point diagrams of cleavages and bedding. Angles between bedding and cleavage, at 137 outcrops, mostly in the lower part of the Deer Trail Group, were found to range from 0° to 90° with a geometric mean in the 30° to 40° range. Therefore, although some strata are oriented at low angles to cleavage and may have been thinned during the deformation, other strata oriented at large angles to cleavage may have undergone thickening. To determine what stratigraphic thicknesses the formations may have had before cleavage developed, the individual sections must be evaluated with respect to orientations of cleavage and bedding as well as intensity of cleavage development.

For purposes of estimating the stratigraphic thicknesses, a minimum of 50 percent decrease in thickness perpendicular to cleavage is considered reasonable for the argillaceous units. Changes in thickness of the dolomitic units is more difficult to assess because blocks of relatively undeformed dolomite occur in zones of intensely cleaved dolomite. A minimum of 30 percent thinning perpendicular to cleavage may be reasonable for some of the dolomite sections. Quartzite may have undergone the least amount of thickness change, as indicated by general absence of cleavages in quartzite. However, significant local thinning is indicated by stretched and flattened clasts and possibly also by large, relatively abrupt changes in thickness of the quartzite, such as in sec. 24, T. 31 N., R. 38 E., which may be pinch and swell structure on the macroscopic scale. For purposes of estimating stratigraphic thicknesses, the quartzites are assumed to have undergone no thinning. Estimated stratigraphic thicknesses are listed in table 1.

The Togo Formation consists of largely slate with minor siltstone and quartzite. The apparent minimum thickness of the section in the southwest part of the quadrangle is 600 m; its cleavage is oriented at low angles to the bedding. Originally, these strata may have been at least 1,150 m thick (table 1).

The Edna Dolomite consists of mostly dolomite, with minor slate and quartzite. Large blocks of relatively undeformed dolomite occur locally, but cleavage is commonly parallel to bedding. The maximum apparent thickness of the section near the east edge of the quadrangle is 700 m. The formation may have been at least 1,050 m thick.

The McHale Slate is nearly all slate with minor interbedded siltstone and limestone. The formation is characterized by well developed slaty cleavage; locally slaty cleavage grades to a fracture cleavage. For the section in the central part of the quadrangle with an apparent thickness of 1,500 m, cleavage is at a large angle to bedding. Therefore, the formation in this area may have been much thinner by an unknown amount. Given a thinning perpendicular to cleavage of 50 percent, the original stratigraphic thickness could have been as little as 755 m to as much as 1,000 m depending on the orientations of the minimum and intermediate strain axes (see Wood, 1974, fig. 4). If the direction of maximum elongation was subhorizontal, trending north-northeast, 750 m is probably closer to the original thickness.

The sections of the Stensgar Dolomite vary widely in thickness in part owing to faulting. The thickest section of the Stensgar is in the Keystone Quarry area where the dolomite is generally not cleaved, although well developed cleavage is present locally near the base of the section. The apparent thickness of 500 m is probably essentially unchanged from the original thickness.

The Buffalo Hump Formation contains quartzite beds which may have been thinned much less than the pelitic facies of the formation. The thickest section of the formation has an apparent thickness of 750 m, and slate makes up 670 m of the section. These strata, slate and quartzite, are estimated to have been at least 1,400 m thick. Sections with more quartzite would have undergone less thinning.

The conglomerate member of the Huckleberry Formation has a well developed cleavage in the argillic matrix; the cleavage is most likely subparallel to bedding. The clasts are flattened parallel to cleavage, especially the slate clasts; they are extremely thin and must have been thinned drastically. The apparent thickness of 200-500 m may represent a stratigraphic thickness of 400-1,000 m. It is possible that the apparent stratigraphic thinning of the conglomerate member to the northeast is a result of greater local tectonic thinning of the unit there (macroscopic pinch and swell structure) or of faulting along the Stensgar Mountain thrust.

The volcanic rocks member of the Huckleberry Formation typically has an apparently random pattern of jointing. Although cleavage is developed in places, the general lack of cleavage in the unit and preservation of microscopic primary textures in metabasalt flows and basaltic tuffs suggests that thinning of the unit may have been insignificant. The range of apparent thickness from 750 to 1,100 m may be due to variations in the original thickness or may be due in part to erosion before burial by the Addy Quartzite.

The Monk Formation consists predominantly of slate with minor amounts of conglomerate and dolomite. Shale and dolomite pebbles in the conglomerate are greatly flattened in the plane of cleavage. The thickness of the formation, apparently 500 m in the section along the boundary of the Stensgar Mountain and Waitts Lake quadrangles, may have been 950 m before thinning.

The Addy Quartzite, although apparently present when slaty cleavage was developed, does not appear to have undergone more than minor crenulation of bedding, microscopic strain of quartz grains (undulatory extinction, polygonization, recrystallization, flattening of grains, and development of pressure shadows), and slight flattening of cobbles and pebbles. No cleavages were observed in the unit. The quartzite, however, is substantially thinner in the Stensgar Mountain quadrangle than it is to the southwest in the Hunters quadrangle, so it may have undergone some tectonic thinning in the study area. The formation in the study area may have been originally closer to the 1,800 m thickness reported in the Hunters quadrangle.

Lack of cleavage in the Old Dominion Limestone suggests that tectonic thinning was less than 30 percent. It is not possible to determine whether the stratigraphic thickness of the formation was increased or decreased.

METAMORPHISM

The rocks in the Stensgar Mountain quadrangle may have undergone at least three episodes of low-grade metamorphism. In the first episode during the Middle(?) and Late Proterozoic, low-grade contact metamorphism probably occurred in the vicinity of the metabasalt and gabbro dikes. Campbell and Loofbourow (1962, p. F28-F29) record slight recrystallization of siliceous rocks, local recrystallization of the Edna Dolomite to coarse marble containing tremolite and talc, and extensive chloritization of the Stensgar Dolomite adjacent to the dikes.

In the second episode, the Proterozoic rocks underwent regional metamorphism in the biotite subfacies of the greenschist facies (Turner and

Verhoogen, 1960, p. 537-538). The presence of epidote, muscovite, and chlorite in the matrix of the Addy Quartzite suggests that these rocks were metamorphosed in the albite-muscovite-chlorite subfacies (Turner and Verhoogen, 1960, p. 534-537), although the modal composition of the quartzite may be controlled by the bulk composition and may not fully reflect the metamorphic conditions. No evidence of metamorphism was found in the Old Dominion Limestone. Superficially, then, the section from the Togo Formation through the Old Dominion Limestone suggests a decrease in grade of metamorphism upward in the section. These relations further suggest that the second metamorphism occurred after Early Cambrian time and that the metamorphic zonation was controlled by depth of burial. The burial metamorphism would have had to precede the major tilting of the section that preceded the later dynamothermal metamorphism.

The dynamothermal metamorphism accompanied duplex thrusting in the Mesozoic. The fact that biotite and chlorite recrystallized along slaty cleavages suggests that the metamorphic grade was near the middle of the greenschist facies.

GEOLOGIC HISTORY

The deposition of most of the Middle Proterozoic Deer Trail Group occurred in generally quiet water (Togo Formation, Edna Dolomite, McHale Slate). The Stensgar Dolomite appears to have been deposited under different conditions. It contains several erosional unconformities suggesting shallow water deposition and may have formed in an evaporative basin, as suggested by solution breccia and magnesite deposits. The general lack of pelitic and silicic detritus indicates isolation from terrigenous sources, a very different environment from earlier sedimentation. The changing character of sedimentation and the erosional unconformity between the overlying Buffalo Hump Formation and the Stensgar Dolomite probably reflect increase in tempo of regional tectonism that could conceivably have begun during McHale sedimentation; this tectonism could have resulted in the slump structures observed by Miller and Clark (1975, fig. 6).

The isolation of the Stensgar basin broke down, possibly abruptly, resulting in a flood of pelitic material making up most of the basal part of the Buffalo Hump Formation. The fine sediment was followed by thick and thin beds of quartz-rich detritus that was deposited in a high-energy environment.

The rapid change from fine-grained quiet-water sedimentation to periodic coarse, high-energy sedimentation could be related to continental rifting proposed during the Late Proterozoic (Stewart, 1972, 1976; Burchfiel and Davis, 1975; Stewart and Suczek, 1977), and would have a bearing on the relation of the Buffalo Hump Formation to the rest of the Deer Trail Group. Whatever the explanation for the change in character of sedimentation near the top of the Deer Trail Group, the change must have involved the western margin of the Belt basin.

The lithologies of the Middle(?) and Late Proterozoic Windermere Group (diamictite, conglomerate, mafic volcanic rocks and minor argillite and dolomite) greatly contrast with those of the older Deer Trail Group. The diamictite has been interpreted as possibly of glacial marine origin (Alto, 1971). The Windermere Group has been interpreted as the basal strata of the Cordilleran miogeosyncline (Stewart, 1976, p. 13), the mafic volcanic rocks of the Huckleberry Formation are considered especially indicative of crustal thinning, presumably in a rifting environment (Stewart, 1972, 1976; Burchfiel and Davis, 1975; Stewart and Suczek, 1977; Devlin and others, 1985). During Windermere time very localized contact metamorphism resulted from intrusion of the mafic volcanic rocks of the Windermere Group. Minor(?) deformation continued during the Late Proterozoic, as the Monk Formation was either not deposited or was eroded before deposition of the Lower Cambrian Addy Quartzite.

Subsidence analyses have suggested that major Cordilleran rifting during which the Cordilleran miogeocline and possibly also the Pacific Ocean began to form during latest Proterozoic to earliest Cambrian (Armin and Mayer, 1983; Bond and Kominz, 1984). However, the date of initial rifting, and the timing of the Windermere sedimentation with respect to continental break-up continues to be debated (Devlin and others, 1985).

Transgression led to deposition of the Addy Quartzite over the Proterozoic rocks, but subsequent deformation may have resulted in nondeposition or erosion of the Maitlen Phyllite before deposition of the Old Dominion Limestone in a possible shelf environment. According to the subsidence analyses (Armin and Meyer, 1983; Bond and Kominz, 1984), these rocks are the basal units of the Cordilleran miogeocline.

The thick section of Proterozoic and Lower Cambrian rocks was buried deeply enough for middle to low grade greenschist facies metamorphism to occur in the formations up through the Addy Quartzite. Subsequently, the section was uplifted and bedding tilted from low angles to about 45° sometime in the Paleozoic or Mesozoic, but before major thrusting in the Mesozoic.

Duplex thrusting in the Mesozoic occurred sometime before emplacement of Cretaceous granitoid intrusions in the area. It was accompanied by dynamothermal metamorphism in the greenschist facies. The units affected, especially in the Deer Trail Group, were complexly faulted and folded, and developed slaty cleavage. During this episode most of the formations were drastically thinned tectonically.

In the Cretaceous large granitoid batholiths were emplaced northeast and southwest of the Stensgar Mountain quadrangle (Miller and Clark, 1975; Becraft and Weis, 1963). Mesozoic or Tertiary felsic dikes and Tertiary andesite dikes intrude the Deer Trail Group.

Uplift and erosion of the study area occurred in the Cenozoic. Glaciation affected the area in the Pleistocene and left extensive deposits of ground moraine in the northeastern corner of the quadrangle, possibly from a lobe of an ice sheet that occupied part of the Colville River Valley to the northeast. A possible end moraine deposit at Klein Meadows suggests that some of the glacial moraine could have been derived from the Stensgar Mountain quadrangle.

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