U.S. DEPARTMENT OF THE INTERIOR TO ACCOMPANY MAP MF-2354
U.S. GEOLOGICAL SURVEY

GEOLOGIC MAP OF THE CHEWELAH 30' X 60' QUADRANGLE, WASHINGTON AND IDAHO

By F.K. Miller
Digital preparation by P.M. Cossette and Pamela D. Derkey

STRUCTURE

INTRODUCTION

[Chewelah quadrangle in this report refers to the Chewelah 1:100,000-scale quadrangle, as distinguished from the Chewelah 7.5’ or Chewelah 15’ quadrangle]

Rocks within the Chewelah quadrangle record a complex and varied structural history from the Late Proterozoic to the Eocene, one that is characterized by several discrete periods of igneous intrusion, folding, normal and reverse faulting, and repeated reactivation of faults during succeeding periods of faulting. The western third of the quadrangle includes a segment of the Kootenay arc, a highly faulted and folded allochthonous sequence of Middle and Late Proterozoic and Paleozoic rocks. Within the quadrangle these rocks have a consistent north-northeast strike. The Jumpoff Joe Fault sharply defines the east boundary of the Kootenay arc, juxtaposing it against coeval rocks of the Belt Supergroup and Paleozoic rocks that have distinctly different structural and lithostratigraphic characteristics. Much of the eastern half of the Chewelah quadrangle is underlain by the western part of the (informal) Priest River complex, which consists of intermixed, heterogeneous two-mica granitic rocks and highly recrystallized Middle Proterozoic rocks. The shallowly dipping Eocene Newport Fault separates the midcrustal crystalline rocks of the Priest River complex from an overlying, relatively thin, spoon-shaped, detached flap of essentially unmetamorphosed Belt rocks. Numerous Cretaceous plutons intrude the hanging-wall Belt rocks, and scattered, syntectonic Eocene nonmarine sedimentary and volcanic rocks are found chiefly in low-lying areas. Eocene normal faults associated with unroofing of the Priest River complex and development of the Newport Fault appear to be the youngest structures affecting rocks in the quadrangle. Granitic rocks that range in age from Triassic or Jurassic to Eocene and represent five discrete intrusive episodes form about 50 percent of the rocks in the area.

KOOTENAY ARC

The western third of the quadrangle is underlain by rocks and structures of the Kootenay arc, which is a broad bend in the regional stratigraphic and structural framework of the Cordillera north and south of the United States-Canada boundary. The arc extends from the north margin of the Columbia River Basalt Group, about 25 km south of the Chewelah quadrangle, through northeastern Washington and northern Idaho to approximately Revelstoke, British Columbia, a distance of about 400 km. Rocks in the Chewelah quadrangle exhibit the characteristics and multiple deformations that Watkinson and Ellis (1987; Ellis, 1986) describe as typical of the Kootenay arc. Within the quadrangle, pre-Mesozoic rocks in the arc form a highly faulted and folded north-northeast-striking, mainly west-dipping sequence that appears to be the deformed upright west half of a faulted, eastward overturned anticline. The north-northeast strike of the rocks and the structural elements that affect them are characteristic of the trend of the southern part of the Kootenay arc, most of which lies in the United States. Beginning at about the United States-Canada boundary, the Kootenay arc makes a broad north and then northwestward bend.

Deformational episodes affected rocks of the Kootenay arc in the Late Proterozoic, the late Paleozoic to middle Mesozoic, and in the Eocene; probably more than one episode occurred in the late Paleozoic to middle Mesozoic interval. Except for Eocene faults, many, if not most, faults in the Kootenay arc show evidence of reactivation in one or more periods of faulting. Because of this, it has not been possible to reliably assign many faults to discrete episodes of faulting nor to accurately date the time(s) that they were active. This is particularly true for faults of the late Paleozoic to middle Mesozoic episode. In addition to being reactivated, some faults were folded or crosscut by younger faults during succeeding episodes of deformation.

Few faults within this part of the Kootenay arc are even moderately well exposed, which compounds the problem of subdividing or classifying faults into genetically related groups, so little is known about the amount or direction of fault dips. Lacking information on fault attitudes precludes distinction between normal and reverse faults, reduces constraints on structural and stratigraphic interpretations, and contributes greatly to the highly interpretive nature of the cross sections. In sections crossing the Kootenay arc, direction of relative motion on faults east of the Stensgar Mountain Fault is indicated mainly by the relative ages of rocks juxtaposed across the fault, except for a few reverse or thrust faults where direct evidence or regional interpretation strongly suggest that younger rocks are faulted over older rocks or where there is relatively good evidence for reactivated faulting. Some faults that were clearly reactivated are labled as such in the cross sections and on the map. Many of the faults between the Stensgar Mountain and Lane Mountain Faults have probably also been reactivated, but evidence for reactivation is lacking.

Late Proterozoic structures are chiefly normal faults associated with continental-scale rifting and deposition of the Late Proterozoic Windermere Group. Devlin and Bond (1988; Devlin and others, 1985) convincingly showed that the Windermere strata were deposited in a rift environment in the Late Proterozoic. However, because most of the older faults in the quadrangle have been reactivated, direct evidence for the extent of Late Proterozoic faulting is lacking. Despite reactivation of faults, the differences in thickness, lithology, and distribution of the Windermere units over short distances, in addition to the fact that Windermere units in adjacent fault blocks were deposited on different units of the Deer Trail Group, strongly suggest that there was considerable fault activity in the Late Proterozoic (Miller, 1994). In the quadrangle and to the northeast in the Metaline area, changes in lithology and thickness of Windermere units over short distances indicate proximity to basin-margin normal faults.

Most of the faults associated with the Kootenay arc are late Paleozoic to middle Mesozoic in age and probably represent several discrete episodes of faulting within that time interval. Both normal and reverse faults developed, but the few crosscutting relations found are conflicting, which suggests that more than one period of reverse faulting and possibly more than one period of normal faulting occurred. Additionally, many of these faults were reactivated both in the late Paleozoic to middle Mesozoic interval and in the Eocene.

Within the quadrangle, the late Paleozoic to middle Mesozoic age range for these faults is constrained only by the Jurassic quartz monzodiorite of Lane Mountain, which postdates all but the Eocene faults in the area, and by the Ordovician Ledbetter Formation, which is the youngest unit affected by the late Paleozoic to middle Mesozoic faults. In the Metaline area, 35 km northeast of the quadrangle, rocks as young as Late Devonian are affected (Joseph, 1990). Roback and Walker (1989) suggest that rocks of the Quesnellia terrane west of the quadrangle were multiply deformed from the Pennsylvanian(?) to the Middle Jurassic and that the latest deformation, which they dated at 170 Ma, was associated with the collision of the Quesnellia terrane. If all faulting in the Kootenay arc during this broad time interval was associated strictly with this collision, the faulting would be limited to a relatively narrow time window. Some of these faults appear to be much older than others, however, these older faults possibly were associated with subduction or strike-slip faulting that predated, but was related to, the collision event.

The Stensgar Mountain and Lane Mountain Faults are the two most prominent faults in the Kootenay arc within the Chewelah quadrangle. Evans (1987, 1988) defined the base of the Windermere Group as the Stensgar Mountain Fault, which he considered to be the roof structure of a duplex thrust fault zone. However, differences in intensity and style of deformation, as well as in relative ages of rocks in the hanging wall and footwall, suggest that the Stensgar Mountain Fault is a normal fault. Highly folded, faulted, and cleaved phyllitic rocks of the Deer Trail Group in the footwall are juxtaposed against much less deformed, predominantly block-faulted rocks of the Windermere Group and lower Paleozoic formations in the hanging wall. This difference in deformation above and below the fault is most striking in the number and types of faults cutting each fault block, but it is evident at all scales down to the microfabric of the rocks. Most of the Deer Trail rocks below the fault, especially the thick argillite sequences, were intensely cleaved and folded on scales ranging from millimeters to tens of meters. Many of the argillitic rocks are cut by two or more distinct sets of cleavages; the strongest cleavage gave rise to a phyllitic foliation.

Despite the deformational contrast between the two fault blocks, the intensity of deformation is relatively uniform within each, which suggests that the two blocks were deformed under different conditions. The relatively undeformed Windermere and lower Paleozoic rocks were subsequently downdropped against the older Deer Trail rocks by the Stensgar Mountain Fault, which by inference has a large displacement.

The Lane Mountain Fault is a west-dipping thrust or reverse fault that cuts the Stensgar Mountain Fault and creates two belts of Windermere Group rocks, each floored by the Stensgar Mountain Fault. Both cross sections show the Lane Mountain Fault cutting and offsetting the Stensgar Mountain Fault. Because accurate information on the surface and subsurface dip of both faults—especially the Stensgar Mountain Fault—is almost totally lacking, the amount of offset shown on the Lane Mountain Fault is highly conjectural. Meager information on the attitude of the Stensgar Mountain Fault suggests that the dip is variable; for at least a part of its length, the attitude of the fault is apparently parallel to the hanging-wall Windermere units along that segment. South of Addy, the fault is clearly folded.

Both the Lane Mountain and Stensgar Mountain Faults may have been active during one or more of several discrete time-separated episodes of faulting during the late Paleozoic to middle Mesozoic interval. In addition to the Lane Mountain Fault, the Stensgar Mountain Fault is cut by a number of other faults. It does, nonetheless, appear to be less deformed than many of the faults that cut rocks of the Deer Trail Group between it and the Lane Mountain Fault, which suggests that it is older than the Lane Mountain and related faults but does not represent the oldest episode of faulting in the late Paleozoic to middle Mesozoic interval.

The highest concentration of faults in the Chewelah quadrangle is found in the footwall of the Stensgar Mountain Fault and appears to include both normal and reverse faults. In that area, probably about the same concentration of faults found in the post-Togo Formation part of the Deer Trail Group are also present in the Togo but the uniform lithology, absence of marker beds, folding, and multiple cleavages in the Togo, make it nearly impossible to identify and trace individual faults.

The Haller Creek Fault is an inferred fault that strikes southward from sec. 29, T. 34 N., R. 39 E. (Miller, 1996a). Its inferred existence and position is based on separation of two significantly different sequences of the Metaline Formation and the fact that Paleozoic rocks west of the fault dip east and are upright. East of the fault and almost everywhere in and around the quadrangle, regardless of dip direction, all sedimentary sequences young westward. West of the Haller Creek Fault, the thick- and thin-bedded limestone member of the Metaline is much thicker than it is east of the fault, and the thick-bedded dolomite member is not recognized. The Haller Creek Fault is offset by the Dunn Mountain Fault, but the inferred location of the segment south of the Dunn Mountain Fault is poorly constrained.

Rocks north of Stensgar Creek and west of U.S. Highway 395 conspicuously depart from the north-northeast trend of the Kootenay arc to strike almost orthogonal to the arc trend. These rocks appear to have been passively pushed into their present orientation by intrusion of the main body of the Starvation Flat Quartz Monzonite, which underlies a large area in the northwestern part of the quadrangle. Correspondingly, the small satellitic pluton centered in the Stensgar Creek drainage may have deflected the Addy Quartzite and the greenstone member of the Huckleberry Formation from a northeast to a northwest strike. The rocks that have this regionally anomalous trend wrap around the main body of the Starvation Flat Quartz Monzonite and then resume parallel to the structural trend of the arc.

JUMPOFF JOE FAULT

The Jumpoff Joe Fault places rocks of the Deer Trail Group against Paleozoic rocks and rocks of the Belt Supergroup and is interpreted to be a west-dipping thrust fault; it appears to define the eastern limit of the Kootenay arc. In the quadrangle, the fault extends from at least Springdale to Johnson Mountain (fig. 1), but it is exposed at only a few places. At those places, the actual fault surface is located to within a few tens of centimeters and must be dug out to be seen. On the hill west of Jumpoff Joe Lake and on the small hill 2 km northeast of Chewelah, it appears to have a moderate west dip. Although not exposed on Eagle Mountain, the straight trace of the fault there suggests that it dips steeply; at other places, it is clearly offset by younger faults, and in a few places it may be folded. North of Johnson Mountain, the Jumpoff Joe Fault is intruded by younger granitic rocks, but projects to the Eocene normal fault that passes through Bayley Lake. This normal fault drops the Jumpoff Joe Fault down so that it is not exposed again west of the Newport Fault.

A number of structural and stratigraphic differences across the Jumpoff Joe Fault suggest it is a complex, multiply reactivated, regional-scale fault with a long movement history. These differences include: (1) intensity and style of rock deformation, (2) regional structural trends, (3) lithostratigraphy of the Deer Trail Group compared to that of the partly coeval Belt Supergroup, (4) lithostratigraphy of Paleozoic units, and (5) the fact that rocks of the Windermere Group occur only west of the fault.

Highly faulted, folded, and cleaved phyllitic rocks of the Deer Trail Group west of the Jumpoff Joe Fault are much more strongly deformed than partly coeval rocks of the Belt Supergroup east of the fault. West of the Jumpoff Joe Fault, the concentration of faults in the Deer Trail is an order of magnitude greater than in Belt rocks east of the fault. Also, regional and small-scale, north-northeast-striking structural elements of the Kootenay arc are extremely consistent west of the Jumpoff Joe Fault but are virtually absent east of the fault.

Based on lithostratigraphy, Miller and Whipple (1989) showed that part of the Deer Trail Group is correlative with part of the upper part of the Belt Supergroup. However, stratigraphic differences between the two units indicate they originated at a considerable distance from one another, implying that they were subsequently juxtaposed by faulting. In addition, Paleozoic sequences above the Addy Quartzite west of the Jumpoff Joe Fault differ from those above the Addy east of the fault. West of the Jumpoff Joe Fault, Paleozoic rocks are represented by thick sequences of the Addy Quartzite, Metaline Formation, and Ledbetter Formation. East of the fault, the Addy is consistently much thinner, possibly due to stratigraphic thinning, but probably also because the upper contact of the Addy there is everywhere faulted. East of the fault, well-documented rocks of the Metaline Formation are found at only one place, 2 km southeast of Springdale, where poor exposures and faulting preclude comparison of the unit's internal stratigraphy with that of the thick and extensive Metaline sections west of the fault. The Ledbetter Formation is not found east of the Jumpoff Joe Fault, nor are the relatively thin, lithologically distinct Devonian and Mississippian formations, so widely exposed east of the fault, found west of it.

Rocks of the Windermere Group unconformably overlie the Deer Trail Group and are unconformably overlain by the Addy Quartzite west of the Jumpoff Joe Fault. East of the fault, the Belt Supergroup is everywhere unconformably overlain by the Addy Quartzite, and the Windermere Group is absent. The thickness of the Windermere west of the Jumpoff Joe Fault is highly variable due to (1) local syndepositional faulting, (2) probable eastward stratigraphic thinning of its lowest two units, and (3) variation in the pre-Addy level of erosion. From the vicinity of Stensgar Mountain, both members of the Huckleberry Formation thin eastward toward the Jumpoff Joe Fault. Over the same distance, the overlying Monk Formation thickens irregularly toward the fault, from zero in the Stensgar Mountain area to as much as 400 m on Empey Mountain. It is not clear from these contradictory relations if the absence of the Windermere Group east of the Jumpoff Joe Fault is due to (1) stratigraphic thinning, (2) Late Proterozoic movement on a proto-Jumpoff Joe Fault that influenced Windermere deposition, (3) juxtaposition of Windermere and non-Windermere-bearing sequences by Phanerozoic thrusting, or (4) some combination of the above. Because the Windermere Group is consistently absent east of the Jumpoff Joe Fault not only in the Chewelah quadrangle but also for a considerable distance northeastward, the fault is thought to have originated as a major Late Proterozoic normal fault that bounded the Windermere sedimentary wedge (Miller, 1994). Large-scale reactivation in late Paleozoic to middle Mesozoic time is thought to have resulted in net contractional movement across the fault.

Concurrent mapping (Miller and others, 1999) extends the Jumpoff Joe Fault northeastward to the United States-Canadian boundary, where it separates rock assemblages that have essentially the same characteristics as those separated by the Jumpoff Joe Fault in the Chewelah quadrangle. Near the international boundary west of the fault, about 8,000 m of the Windermere Group lie between rocks of the Deer Trail Group and the Late Proterozoic and Cambrian quartzite. East of the fault, only the lower part of a relatively undeformed sequence of the Belt Supergroup is preserved, therefore the absence of Windermere rocks between it and the Cambrian section can only be presumed. As in the Chewelah quadrangle, the Deer Trail rocks are extremely deformed and cut by numerous pre-Jurassic faults, structural features that are absent in Belt rocks east of the fault.

At the international boundary (fig. 3), the Jumpoff Joe Fault is roughly aligned with the St. Mary Fault 55 km to the northeast; the two faults may represent disconnected parts of a single structure (Miller, 1994). Northwest of the St. Mary Fault, 9,000 m of the Windermere Supergroup (lateral equivalent of the Windermere Group in the United States) unconformably overlie the Purcell Supergroup (lateral equivalent of the Belt Supergroup in the United States) and are unconformably overlain by Late Proterozoic and Cambrian rocks (Rice, 1941). Southeast of the fault, Late Proterozoic and Cambrian rocks unconformably lie directly on the Purcell Supergroup. Lis and Price (1976) interpreted this relation to indicate that the St. Mary Fault is a syndepositional Late Proterozoic normal fault that created an elevated sediment source area on its southeast side and a downdropped basin of accumulation on its northwest side. The fault was reactivated with an opposite sense of throw in post-Windermere time.

The intervening area between the Jumpoff Joe and St. Mary Faults is largely underlain by Cretaceous granitic rocks that postdate latest movements on both faults (Archibald and others, 1984) and by a complexly deformed and metamorphosed assemblage of Proterozoic to Mesozoic sedimentary rocks. The Purcell trench detachment (Rehrig and others, 1987; Armstrong and others, 1987) and older faults (Archibald and others, 1984) that may be concealed in the trench (fig. 3) also intervene between the Jumpoff Joe Fault and the St. Mary Fault. Because of the similarity of structural and stratigraphic relations across the two faults and their approximate alignment, the Jumpoff Joe and St. Mary Faults are considered to have been either a single continuous structure or parts of a single fault system. Furthermore, that fault or fault system represented a major Late Proterozoic bounding structure for the Windermere sedimentary wedge and persisted as a major locus of crustal weakness at least through the late Paleozoic to middle Mesozoic time interval.

Despite possible normal movements in the Late Proterozoic, net strain on the Jumpoff Joe Fault is contractional judging from the great dissimilarity of Middle Proterozoic and Paleozoic sequences juxtaposed by the fault. If extension related to development of the Stensgar Mountain Fault in the late Paleozoic to middle Mesozoic and metamorphic core complex unroofing in the Eocene caused reactivation of a proto-Jumpoff Joe Fault, then the total amount of contractional strain is even greater than the net strain indicated by differences in lithostratigraphy across the fault. Present understanding of lithofacies distribution in the western Belt basin is not sufficient to be used as a tool to estimate the amount of net contractional strain required to juxtapose the Deer Trail Group and Chewelah Belt Supergroup sequences. The multiple-deformation history of the region also compounds the uncertainty of any estimate.

BELT SUPERGROUP

The 9,000-m-thick sequence that comprises the Belt Supergroup in the Chewelah area east of the Jumpoff Joe Fault represents the westernmost occurrence of the Belt Supergroup. This sequence and the one north of Newport form a normal, or slightly telescoped, progression in lithofacies with Belt sequences east of the Chewelah quadrangle. However, the Deer Trail Group west of the Jumpoff Joe Fault, which is correlative with part of the upper part of the Belt Supergroup at Chewelah and Newport (Miller and Whipple, 1989), requires a change in lithofacies far greater than could be expected in the distance separating it from the Chewelah and Newport Belt sequences. This seemingly abnormal difference in lithofacies is attributed to net foreshortening along the Jumpoff Joe Fault.

Structures developed in rocks of the Belt Supergroup east of the Jumpoff Joe Fault contrast sharply in density, type, and trend with those developed in rocks west of the fault. In the sequences of the Belt Supergroup at both Newport and Chewelah, the general strike of fold axes, bedding, and most faults is approximately north. The Belt sequence north of Newport forms the west limb of a broad antiformal structure with a roughly north-south axis named the Snow Valley Anticline by Schroeder (1952). Harms and Price (1992) considered this fold to be related to the Newport Fault, but the fold is almost perfectly concentric about the two-mica granodiorite of Dubius Creek and, therefore, is either related to emplacement of that body or predates it. A number of faults are developed in the western part of the sequence north of Newport, but many of these are Eocene normal faults related to the Newport Fault. The rest are steep reverse faults that predate the Cretaceous granitic rocks. None of these faults appear to be folded as are many in the Kootenay arc, nor is there evidence for pre-Cretaceous reactivation of any of these faults. The reverse faults near and west of the Pend Oreille River that are queried on the geologic map and on cross sections A-A' and B-B' are highly inferential.

The eastward swing of bedding at the north and south ends of the Belt sequence east of Chewelah suggests that these rocks are also the western part of an antiformal structure that has a north-striking axis, the eastern half of which is cut off by intrusion of granitic rocks of the Priest River complex. However, north of the Grouse Creek and Cottonwood Creek drainages, the west-dipping section is progressively more overturned to the west and appears to be nearly recumbent north of Eagle Mountain. The axial plane of the overturned fold is nearly horizontal and apparently lies at a relatively shallow depth, because the Prichard Formation with a moderate westward dip is not overturned in the bottom of the canyon on the north side of Chewelah Mountain but is strongly overturned on the mountain tops bounding the canyon on the north and south. This overturning, which necessitates a shear direction of top-to-the-west, is inconsistent with any known pre-Eocene faulting in the Chewelah area. The overturning may be caused by lateral movement associated with ductile attenuation of rocks to the east during Eocene unroofing of the Priest River complex (see cross sections A-A' and B-B').

PRIEST RIVER COMPLEX, NEWPORT FAULT, AND EOCENE NORMAL FAULTS

The (informal) Priest River complex, which is a large body of predominently two-mica granitoid rocks that includes lesser amounts of metasedimentary and meta-igneous rocks, is one of several metamorphic core complexes in northern Washington, northern Idaho, and southern British Columbia. From a point near the middle of the Chewelah quadrangle, it extends eastward beneath the Newport Fault, emerges 13 km east of the quadrangle on the east side of the Priest River, and continues for another 20 km to the Purcell trench (fig. 3). In a north-south direction, the complex extends from about latitude 48° N to at least the United States-Canada boundary. The west side of the complex, which is located in the Chewelah quadrangle, includes rocks that intrude rocks of the Belt Supergroup, but the east side may be bounded by a detachment fault concealed beneath thick Quaternary deposits in the Purcell trench (Rehrig and others, 1987). McCarthy and others (1993) present evidence to suggest that unroofing of the part of the complex located between latitudes 48° N and 49° N was accomplished more by accomodation along the Newport Fault than by movement along a detachment fault in the Purcell trench.

The Priest River complex is composed of more than 20 distinct plutonic units that are interspersed with irregularly shaped masses of mixed metamorphic and granitic rocks. In the Chewelah quadrangle, two-mica granitic rocks of the Phillips Lake Granodiorite and highly recrystallized metasedimentary rocks derived from the Belt Supergroup form the southwestern part of the complex. Lithologically, the Phillips Lake is an extremely heterogeneous unit. It consists chiefly of two-mica granodiorite in its western part but contains irregularly increasing amounts of leucocratic dike rocks and metamorphic rocks eastward toward the Newport Fault. Roughly paralleling this eastward trend of increasing heterogeneity is a progressive loss of potassium feldspar, a decrease in potassium-argon cooling ages and, beginning about 4 km west of the Newport Fault, progressive development of a mylonitic fabric.

The Newport Fault has a 200-km-long, U-shaped trace, which continues east and north of the quadrangle almost to the United States-Canada boundary (fig. 3) (Miller and Engels, 1975; Miller, 1994). It is a spoon-shaped, shallowly dipping normal fault (Harms and Price, 1992) with a U-shaped trace; dip of the fault steepens northward along both limbs of the U. The fault juxtaposes low-metamorphic-grade rocks of the Belt Supergroup and intermediate- to shallow-depth plutonic rocks in the hanging wall against an infrastructure of two-mica granitic rocks, coarsely recrystallized Belt rocks, and pre-Belt(?) basement gneiss (Miller, 1971; Harms and Price, 1992).

A zone of chlorite breccia and chloritized, brittlely deformed rocks that ranges from about 150 to 250 m thick mark the upper part of the fault zone in the footwall. This brittle deformation overprinted ductilely deformed mylonite that extends variable distances away from the fault. Thinner, discontinuous zones of chlorite breccia up to 120 m wide are found in the footwall up to 6 km from the main trace. Harms and Price (1992) mention that the mylonite ranges from 15 to 500 m thick but is best developed beneath the southern and eastern parts of the fault. In the Winchester Creek area, beneath the western part of the fault, footwall rocks are mylonitized up to 4 km west of the main fault trace. The foliation in rocks southeast of the Silver Point Quartz Monzonite is a mylonitic foliation and is accompanied by a well-developed stretching lineation. This zone of mylonite is probably the extension of the mylonitic rocks mapped by Rhodes (Rhodes and

Hyndman, 1984) to the south, and also may be the extension of the footwall mylonitic rocks along the east side of the Newport Fault (McCarthy and others, 1993). This mylonitic fabric may terminate against the Cretaceous Fan Lake Granodiorite at depth, because no surface exposures of Fan Lake Granodiorite are mylonitized. It is possible, however, that, at the present level of erosion, the zone of mylonite bends to the northeast missing the Fan Lake Granodiorite, but cuts the pluton at depth.

Harms and Price (1992) calculate that between 23 and 30 km of lateral extension occurred on the western limb of the Newport Fault. Restoration of the sequence comprising the Belt Supergroup at Newport with the one at Chewelah using the larger of these figures would make the Prichard Formation-Burke Formation contact in the two sequences almost exactly coincident. However, these two Belt sequences are lithostratigraphically dissimilar in a number of respects. The argillite of Half Moon Lake, Shepard Formation, Snowslip Formation, St. Regis Formation, Revett Formation, and Burke Formation are all thicker in the Chewelah sequence; some units are thicker by nearly 100 percent. Several of these units, particularly the upper ones, exhibit differences in lithofacies between the two sequences. In addition to these differences, the abundant thick mafic sills enclosed within the Prichard Formation in the southern part of the Newport sequence are conspicuously absent in the Prichard in the southern part of the Chewelah sequence.

Using sections of the Belt Supergroup in western Montana that are separated by about the same distance as a basis for comparison, the differences in thickness, lithofacies, and concentration of sills between the Newport and Chewelah Belt sequences suggest that they were not originally contiguous and that they actually may have originated at a greater distance apart than presently separates them. However, Harms and Price (1992) documented unequivocally that the Newport Fault and related bounding structures were produced by extension. This apparent conflict suggests that the two Belt sequences may have been telescoped by late Paleozoic to middle Mesozoic thrusting prior to Eocene extension. Because all thrust faulting in the region predates the voluminous 100-Ma granitic rocks, presumably any older contractional structure(s) in the area has been obliterated by granitic plutonism.

Eocene normal faults, some of which cut Cretaceous plutons, are associated with the development of the Priest River complex and other metamorphic core complexes in the region. These faults are well developed in the Dunn Mountain area, the Colville Valley, and the Pend Oreille River valley. The low-dipping, relatively coherent mass of Windermere and lower Paleozoic rocks that underlies Dunn Mountain, herein referred to as the Dunn Mountain block, is bounded on the east by the Dunn Mountain Fault, a large Eocene normal fault. Although the fault is concealed and its dip direction is unknown, two lines of evidence suggest that it is an east-dipping normal fault associated with Eocene extension rather than an older west-dipping reverse fault. First, at the north end of Dunn Mountain, the Starvation Flat Quartz Monzonite is highly brecciated and chloritized adjacent to the projected trace of the fault. Throughout the region, only Eocene faults are known to cut Cretaceous plutonic rocks. Second, Eocene volcanic rocks and breccia occur discontinuously along the east side of the fault. Outside of the large Eocene grabens to the west, Eocene volcanic and sedimentary rocks occur almost exclusively in small isolated basins bounded on at least one side by Eocene faults. Presumably the faults are responsible for creating topographic lows into which the Eocene volcanic and sedimentary materials were deposited. Alternatively, late in the extensional process these faults downdropped and preserved small areas that now represent previously more extensive Eocene rocks.

Movement on the Dunn Mountain Fault has tilted and uplifted the Dunn Mountain block about 3,600 m relative to the hanging wall. Southeast of Dunn Mountain, the Dunn Mountain Fault terminates against an older fault which was reactivated in the Eocene. Displacement across the Dunn Mountain Fault is approximately twice as great at its north end as it is at its south end. The difference appears to be caused by five inferred down-to-the-west Eocene normal faults east of Dunn Mountain. These faults curve into and terminate against the Dunn Mountain Fault and appear to have been active in concert with it.

The concealed Eocene normal fault west of Wrights Mountain downdrops Late Proterozoic and Cambrian Addy Quartzite against rocks of the lower part of the Deer Trail Group and continues southward at least to the vicinity of Camas Valley. Tilting by down-to-the-east movement on the fault probably caused the west dips in the Sanpoil Volcanics just south of Waitts Lake during or shortly after deposition of the unit (see cross section B-B').

Sixteen km north-northeast of Chewelah, renewed movement on a fault that is probably related to the Jumpoff Joe Fault produced a similar localized Eocene basin (Gager, 1982), as well as a 6-km-long chlorite breccia zone in the Phillips Lake Granodiorite. This breccia zone is probably part of a lesser detachment fault that continues southward, and in conjunction with the large east-dipping normal fault west of Wright Mountain, downdrops the block in which the Eocene volcanic and sedimentary rocks of the Colville Valley are preserved.

A number of Eocene faults are inferred in the Colville Valley and several more in the Pend Oreille River valley, where some of them offset the Sanpoil Volcanics. The concealed normal faults west of the Pend Oreille River are highly inferential. Additional Eocene normal faults are probably concealed beneath Quaternary deposits in major river valleys.

FOLDS

Folds in the Chewelah quadrangle include the major antiformal structures in the Kootenay arc and in the rocks of the Belt Supergroup sequences at Newport and Chewelah as discussed above, as well as a number of smaller folds. With few exceptions, almost all small-scale folds are found in the Kootenay arc and have axes approximately parallel to those of the larger north- to north-northeast-trending folds. Most small-scale folds are symmetrical to asymmetrical in shape and have amplitudes ranging from 1 cm to several meters.

Northwest of Wrights Mountain, an anticline-syncline pair in rocks of the Deer Trail Group is cut by multiple reverse faults which may have formed in the later stages of folding. North of there, on Gold Hill and Deer Mountain, an anticline and syncline in Deer Trail, Windermere, and Cambrian rocks, may be related to this fold pair. West and northwest of Lane Mountain, a north-trending, variably north-plunging anticline-syncline pair folds rocks of the Deer Trail Group and most of the faults that cut them. Both folds are slightly overturned to the east at their north ends and appear to be cut off by the Stensgar Mountain Fault. The folds do not appear to be present in rocks of the hanging wall, implying that they are older than the fault. Several smaller, similar folds developed in Proterozoic rocks a few km to the northeast may form the core of a complexly faulted flexure, which is asymmetric and plunges moderately north. A section across this structure indicates a west-over-east shear couple and, on the geologic map, a possible right-lateral shear component.

INTRUSIVE EPISODES

Granitic rocks within the quadrangle were intruded in the Late Triassic or Early Jurassic, Middle Jurassic, Late Cretaceous, and Eocene. The alkalic Flowery Trail Granodiorite (quartz monzodiorite) is the only Late Triassic or Early Jurassic pluton, and the quartz monzodiorite of Lane Mountain is the only Middle Jurassic pluton. Both are mineralogically distinct from the Cretaceous and Tertiary plutons, having much lower quartz contents and higher hornblende:biotite ratios than the younger intrusive bodies.

By far the largest volume of granitic rocks was emplaced in the Cretaceous. These rocks include two petrogenetically distinct types, hornblende-biotite granitoids and two-mica granitoids, which, within the resolution of conventional potassium-argon dating, were intruded in the same time interval. Even though exposures of crosscutting relationships between plutons are generally not adequate to establish the sequential order of emplacement for all granitoids, they are sufficient to establish that emplacement of the two petrogenetically distinct types overlapped in time.

Co-existing mineral pairs from most plutons in the hanging wall of the Newport Fault and from plutons in the footwall that are sufficiently distant from the Priest River complex yield concordant or near-concordant potassium-argon ages that are considered to approximate emplacement ages. Within the Priest River complex, potassium-argon dating yields cooling ages that range from near-emplacement ages on the west side of the complex to the age of unroofing in parts of the complex closer to the Newport Fault (Miller and Engels, 1975; Harms and Price, 1992).

Eocene intrusions are restricted to the two plutons composed of the Silver Point Quartz Monzonite, the quartz monzodiorite of Ahern Meadows, the quartz monzonite of Loon Lake, and numerous hypabyssal dikes. If the Sanpoil Volcanics in the hanging wall of the Newport Fault are the extrusive equivalent of the coeval Silver Point Quartz Monzonite, as suggested by Russell (1989), then determination of the emplacement depth of the Silver Point would provide a minimum uplift figure for the footwall rocks of the Newport Fault.

DESCRIPTION OF MAP UNITS

Qm **Mine tailings (Quaternary)—**Tailings and waste rock from magnesite and silica quarrying operations. Shown only at Lane Mountain, Finch, and Moss-Allen Quarries at Huckleberry Mountain

Qls **Landslide deposits (Quaternary)—**Unconsolidated basalt rubble resulting from slides along steep erosional scarps formed in basalt flows of Columbia River Basalt Group. Also, two small slides: one in McKale Canyon, and one on southeast flank of Deer Mountain. Identified chiefly by geomorphic form on aerial photographs

Qag **Glacial and alluvial deposits (Quaternary)—**Till from continental glaciation and all alluvial material in modern drainages. Glacial and alluvial deposits generally pale tan or pale gray; some alluvial deposits locally reflect colors of bedrock source. Unconsolidated boulders, gravel, sand, and silt. Semi-consolidated to consolidated clay, silty clay, and sandy clay. Clay-bearing parts well bedded to indistinctly bedded. Coarser parts are well bedded, indistinctly bedded, and massive; some sand and gravel deposits are well-bedded, displaying large- and small-scale cross bedding. Clasts in alluvial deposits reflect local bedrock sources. Clasts in glacial deposits are mostly from bedrock units recognized in quadrangle, but also include exotic metamorphic, granitic, and volcanic clasts. Clasts range from subrounded to well rounded; nearly spherical cobble-sized clasts are abundant locally. Flattened clasts commonly are imbricated. In general, degree of rounding does not appear to be correlative with clast size. Thickness of glacial and alluvial deposits is highly variable, ranging from thin, discontinuous deposits near bedrock or in steep canyons, to possibly more than 100 m in major river valleys

QTs **Consolidated alluvial and (or) glacial deposits (Quaternary and (or) Tertiary)—**Conglomerate, sedimentary breccia, minor arkosic and lithic sandstone, iron-oxide-cemented sandstone and conglomerate, and mudstone. Restricted to southwestern 15-minute block of quadrangle. Most of unit is poorly indurated conglomerate, both clast and matrix supported. Bedding features are rare except where sandstone beds or lenses are present. At many places, unit is monolithologic or nearly monolithologic, especially where it flanks areas of extensive Late Proterozoic and Cambrian Addy Quartzite. Part of unit characterized by conspicuous red-orange soil; outer surface of most clasts stained red-orange. Red-orange sediments are generally much less indurated than less pigmented parts of unit. Between south and middle forks of Chamokane Creek (fig.1), imbricated flat pebbles in conglomerate bed indicate flow direction of S. 20 W., which does not reflect modern slope direction. Maximum thickness unknown, but probably at least 40 m on south side of Empey Mountain. Red-orange part of unit may be pre-Wisconsin glacial deposit, but could represent late Tertiary sedimentation, possibly during Pliocene warm period; may actually be different unit from more highly indurated, less pigmented parts

Tcl **Clay deposits (Miocene?)—**Clay, silty clay, sandstone, and siltstone. Also contains layers of bog iron that are rust-colored cohesive material consisting of sand, silt, and clay cemented by hydrous iron oxides and commonly having botryoidal form. Shown only in three small areas 6 km east-northeast of Loon Lake. Considered by Hosterman (1969) to have formed by alteration of basalt of Columbia River Basalt Group

Tcr **Columbia River Basalt Group (Miocene)—**Fine-grained tholeiitic basalt comprising several flows. All considered to be N_2 flows of Grande Ronde Basalt by Swanson and others (1979) based on composition, texture, and magnetic polarity (Waggoner, 1990). Thickness of flows ranges from a few meters to several tens of meters. Flow breccia layers between some flows. Maximum thickness estimated to be about 180 m; thins northwestward

Tt **Tiger Formation (Eocene)**—Conglomerate, lithic arkose, and siltstone. Highly variable laterally and vertically. Poorly to moderately well indurated. East of Pend Oreille River, consists of poorly stratified, moderately well indurated clast- and matrix-supported cobble and boulder conglomerate. Clasts are quartzite, siltite, argillite, and dolomite derived from Belt Supergroup and, locally, Eocene Sanpoil Volcanics. West of Pend Oreille River, poorly to moderately well indurated siltstone, lithic arkose, and conglomeratic arkose predominate. In places near Newport Fault, Tiger Formation is poorly indurated, nearly monolithologic conglomerate that consists of locally derived boulders of two-mica granitic rock in a matrix of granitic rock debris of about same grain size as parent granitic rocks; conglomerate strongly resembles deeply weathered, in-place granitic rocks. Tiger Formation contains middle Eocene palynomorphs (Harms and Price, 1992). Formation considered to be syntectonic basin-fill deposits related to movement on Newport Fault (Gager, 1984; Harms and Price, 1992)

Tcs **Conglomerate and sedimentary breccia (Eocene)**—Moderately well-indurated pebble to boulder conglomerate exposed in a limited area between Addy Mountain and Bayley Lake (fig. 1). All clasts are locally derived, mostly from Starvation Flat Quartz Monzonite and greenstone member of Huckleberry Formation. Thickness unknown. Considered by Gager (1982) to be local basin-fill deposits similar to Tiger Formation but related to movement on normal fault bounding conglomerate on east

Tcb **Chlorite breccia and cataclastic rocks associated with the Newport Fault Zone (Eocene)**—Green or gray, finely comminuted cataclastic rocks and highly fractured and chloritized rocks in footwall of Newport Fault Zone. Near fault plane, includes nearly aphanitic chlorite breccia containing millimeter- to centimeter-long, angular, internally fractured, matrix-supported feldspar fragments. Away from fault plane, grades progressively into less fractured rock in which all mafic minerals are chloritized and fracture planes are chlorite coated. Best developed along southern part of fault where zone is between 150 and 250 m thick

Tcc **Tectonic breccia of Cusick Creek (Eocene)**—Breccia and gouge ranging in size from powder to house-size blocks; many larger blocks consist of recemented breccia. Restricted to hanging wall of Newport Fault; derived from local hanging-wall units. Western contact with Newport Fault fairly sharp; eastern contact gradational. Outside Cusick Creek area, hanging-wall rocks immediately adjacent to Newport Fault are rarely exposed, but are presumed to be similar to tectonic breccia at Cusick Creek

Tbl **Cataclastic rocks of Bayley Lake (Eocene)**—Highly comminuted cataclastic rocks and breccia associated with Eocene extensional tectonism (Gager, 1982). Much of unit is chlorite breccia. Most rocks are pale gray or pale green; differs from darker green cataclastic rocks associated with Newport Fault Zone, possibly because protolith in Bayley Lake area was relatively leucocratic. Rocks coherent, but cut by numerous chlorite-coated fractures. Intensity of brecciation and width of cataclastic zone are variable along strike. Appears to die out northeastward

Ts **Sanpoil Volcanics (Eocene)**—Volcanic flow rocks, sedimentary and volcanic breccia, conglomerate, and lithic arkose. Includes rocks comprising Pend Oreille Andesite of Schroeder (1952), previously unnamed volcanic flows and sedimentary rocks south of Waitts Lake, flow or intrusive andesite northeast of Waitts Lake, and isolated outcrops of flow and sedimentary rocks east of Dunn Mountain (fig. 1). Rocks comprising Pend Oreille Andesite of Schroeder (1952) assigned by Pearson and Obradovich (1977) to Sanpoil Volcanics on basis of age and compositional similarities to Sanpoil Volcanics in Turtle Lake quadrangle (Becraft and Weis, 1963) 30 km southwest of map area. Flow rocks consist of olivine, orthopyroxene, clinopyroxene, hornblende, biotite, and feldspar phenocrysts set in gray, green, or lavender microcrystalline matrix of plagioclase and altered glass; not all mafic minerals listed above are present in all flow rocks. Layering of volcanic rocks reliably defined only by interlayered sedimentary rocks. East of Pend Oreille River, unit contains no sedimentary rocks. South of Waitts Lake, unit consists of conglomerate, sedimentary breccia, lithic arkose, and minor siltstone interlayered with at least three andesitic lava flows; sedimentary rocks contain abundant carbonaceous material and, at one locality, a coal seam. Conglomerate both matrix- and clast-supported; most clasts poorly rounded and poorly sorted. Stratification is apparent only where rare, finer-grained beds or lenses are present. In noncontiguous outcrops on east side of Dunn Mountain, unit consists of (1) flow rocks similar to those south of Waitts Lake, (2) unlayered, highly fractured, pale-gray, pink, and white altered and silicified volcanic breccia and flow(?) rocks, and (3) possible flow rocks or vent agglomerate (from R.G. Yates, Waggoner, 1990)

Thd **Hypabyssal dikes (Eocene)**—Fine-grained mafic dikes of widely variable mineralogic and chemical composition. Light to dark gray. Contain phenocrysts of hornblende, biotite, plagioclase, potassium feldspar, quartz, and rare pyroxene; most dikes contain only one to three of these minerals as phenocrysts. Groundmass ranges from very fine-grained to aphanitic; larger dikes have chilled margins. Apatite, zircon, sphene, and opaque minerals are most common accessory minerals. Dikes large enough to show on map are present north of Eloika Lake, west of Nelson Peak, and east of Chewelah. Dikes ranging from 0.5 m to 10 m in width are found throughout quadrangle but are most abundant in middle and lower parts of Belt Supergroup between Loon Lake and Flowery Trail Road (fig. 1). Russell (1993) showed that major and trace element chemistry of hypabyssal dikes indicate genetic relation to Eocene Silver Point Quartz Monzonite (Tsp). Both hornblende and biotite from a dike 400 m north of Elk in southeastern part of quadrangle yield potassium-argon age of 48 Ma (Miller, 1974c; recalculated using current IUGS constants, Steiger and Jaeger, 1977)

Tsp **Silver Point Quartz Monzonite (Eocene)**—Hornblende-biotite monzogranite and granodiorite (fig. 2). Unit was formally named quartz monzonite before adoption of currently used rock classification (Streckeisen, 1973). Forms two large noncontiguous plutons in southern part of quadrangle; southern pluton extends at least 7 km south of quadrangle. Rock is porphyritic with phenocrysts averaging about 3 cm in length. Groundmass has granitic texture with no preferred orientation of mineral grains and distinctive bimodal grain size; about 30 percent of hornblende, biotite, plagioclase, and potassium feldspar crystals are about 4 mm long, and about 70 percent of these four minerals, as well as quartz, are about 1 mm long. Plagioclase composition is about an_{20}; potassium feldspar is perthitic and shows no microcline twinning; hornblende and slightly less abundant biotite give rock color index of about 15. Sphene, apatite, zircon, allanite, and opaque minerals found in all rocks. Both plutons extremely homogeneous with respect to composition and texture except in northern part of northern body. There, rock has primary igneous foliation, is noticeably more mafic, and is granodioritic in composition. From rock near Loon Lake, both hornblende and biotite give potassium-argon ages of 48 Ma; rock from near Sacheen Lake gives 52 Ma and 49 Ma, respectively (Miller, 1974c; recalculated using current IUGS constants, Steiger and Jaeger, 1977). Zircon from Sacheen Lake rock yielded discordant uraniun-lead age of 52.1 ±1.2 Ma (Whitehouse and others, 1992)

Tam **Quartz monzodiorite of Ahern Meadows (Eocene)**—Hornblende-biotite quartz monzodiorite and quartz monzonite (fig. 2). Forms small pluton 2.5 km southeast of Springdale. Rocks are nonporphyritic, medium- to coarse-grained, equigranular, and have no directional fabric. Plagioclase composition ranges from an_{25} to an_{30}; potassium feldspar is white, microperthitic, and has no microcline twinning; hornblende and biotite about equal in amounts give rock average color index of 28. Some hornblende has pyroxene cores. Sphene extremely abundant, up to 0.5 percent in some rocks; other accessories include apatite, zircon, and opaque minerals. Biotite and hornblende give potassium-argon ages of 50 Ma and 65 Ma, respectively, but hornblende is contaminated by pyroxene that contains excess argon (Miller and Clark, 1975; ages recalculated using current IUGS constants, Steiger and Jaeger, 1977). Emplacement age is considered to be 50 Ma. Quartz monzodiorite of Ahern Meadows probably petrogenetically related to Silver Point Quartz Monzonite

Tll **Quartz monzonite of Loon Lake (Eocene)**—Hornblende-biotite quartz monzonite (fig. 2). Forms small pluton 3 km east of Springdale. Fine grained, hypidiomorphic granular, nonporphyritic, no directional fabric; grain size slightly variable. Composition and texture, except for grain size, very uniform. Plagioclase averages an_{22}; potassium feldspar is pale pink, microperthitic, and shows no microcline twinning; hornblende and biotite about equal in amounts giving rock average color index of 18. Almost all hornblende has pyroxene or partially altered pyroxene in core. Sphene very abundant; other accessory minerals include apatite, zircon, allanite, and opaque minerals. Fresh rock extremely tough and difficult to break. Biotite gave potassium-argon age of 51 Ma. Coexisting hornblende containing altered pyroxene could not be totally purified and gave ages of 320 Ma (32 percent pure hornblende), 294 Ma (40 percent pure hornblende), and 195 Ma (58 percent pure hornblende) (Miller and Clark, 1975; ages recalculated using current IUGS constants, Steiger and Jaeger, 1977). Age of unit is considered to be 51 Ma; quartz monzonite of Loon Lake probably petrogenetically related to Silver Point Quartz Monzonite

To **O'Brien Creek Formation (Eocene)**—Tuff, arkose, arkosic sandstone, and conglomeratic arkose. White- to pale-gray, thin-bedded tuff containing scattered grains of euhedral to subhedral biotite overlies arkosic rocks. Pale-gray to tan, lithic arkosic rocks in 2- to 20-cm-thick beds. Found only in small area in eastern part of sec. 9, T. 33 N., R. 44 E.; underlies Tiger Formation and overlies either Sanpoil Volcanics or Late Proterozoic and Cambrian Addy Quartzite. Potassium-argon age of biotite from tuff is 53 Ma (Pearson and Obradovich, 1977)

Kmm **Monzogranite of Midnight Mine (Cretaceous)**—Leucocratic, quartz-rich, muscovite-biotite monzogranite (fig. 2). Northeastern part of pluton located in southwestern corner of quadrangle. Medium to coarse grained; sparsely and irregularly porphyritic; contains pale-pink, slightly perthitic orthoclase phenocrysts as much as 3 cm in length. Aggregates of dark-gray quartz as much as 15 mm (7 mm average) long characterize monzogranite. Calcic oligoclase and potassium-feldspar (perthitic orthoclase) grains in groundmass average 3 to 5 mm in length. Color index averages about 5; biotite:muscovite averages about 2:1. Rock has no directional fabric, and grain size is slightly bimodal. Ludwig and others (1981) reported very slightly discordant uranium-lead age of 75 Ma on zircon from lithologically similar pluton at Midnight Mine located 10 km to southwest

Klr **Monzogranite of Little Roundtop (Cretaceous)**—Very coarse-grained biotite monzogranite (fig. 2). Forms two noncontiguous plutons, one east of Deer Lake and the other south of Deer Lake. Most rock is very deeply weathered and poorly exposed. Average grain size over 1 cm. Average plagioclase composition an_{20}; some potassium feldspar shows microcline twinning, but most does not; quartz forms large dark-gray anhedral crystals; most biotite is slightly finer in grain size than other minerals, has chloritized rims, and gives rock average color index of 7. Much of monzogranite has irregularly distributed concentrations of 2- to 4-cm-long potassium feldspar phenocrysts; not obvious everywhere owing to very coarse grain size of groundmass. At some places, plutons have leucocratic, muscovite-bearing border phase whose groundmass has bimodal grain size. Biotite in limited exposures east of Eloika Lake contains closely spaced, interlaced needles of sillimanite. Monzogranite of Little Roundtop is lithologically indistinguishable from coarse-grained phase of porphyritic quartz monzonite (monzogranite of Streckeisen, 1973) of Becraft and Weis (1963) in Midnight Mine area of Turtle Lake quadrangle, 8 km southwest of Chewelah quadrangle. Zircon from that body gives slightly discordant $^{206}Pb/^{238}U$ age of 91 Ma (Ludwig and others, 1981)

Kbgm **Muscovite monzogranite of Blue Grouse Mountain (Cretaceous)**—Medium- to coarse-grained muscovite monzogranite, locally garnet-bearing. Modal composition of rock is monzogranite (fig. 2); normative composition is granite, indicating rock is highly evolved. Forms four aligned, noncontiguous bodies east of Deer Lake. Average plagioclase composition is an_3, potassium feldspar is white or pale-pink microcline. Quartz forms medium-gray anhedral masses; muscovite has same grain size as other minerals and averages 6 percent of rock. Contains no mafic minerals except for sparse specks of magnetite, limonite, or pyrite. Largest pluton east of Deer Lake has wide greisen zone around its southern part and is probably related to nearby huebnerite-bearing quartz veins. Muscovite gives potassium-argon age of 80 Ma (Miller and Clark, 1975; ages recalculated using current IUGS constants, Steiger and Jaeger, 1977), but age could reflect complex cooling history of region. Age could be as old as 100 Ma, based on similarity to nearby granitic rocks of this age, or about 90 Ma, based on close spatial association of muscovite monzogranite with monzogranite of Little Roundtop

Starvation Flat Quartz Monzonite (Cretaceous)—Named by Clark and Miller (1968) for exposures at Starvation Flat (fig. 1) in northwestern part of quadrangle. Unit underlies about 175 km^2. Forms three discrete plutons: a large body located northwest of Addy Mountain and the Arden pluton which forms its core, as well as a smaller body centered in Stensgar Creek drainage (fig. 1). Most rock is medium to coarse grained and homogeneous in appearance. Large pluton has slight compositional zonation: more mafic than average along its southern and western borders, average composition in most of its interior, and leucocratic in its core (Arden pluton). Consists of:

Ksh **Hornblende-biotite monzogranite and granodiorite**—Average plagioclase composition about an_{35}; potassium feldspar is pale-pink anhedral perthite; hornblende:biotite ratio about 7:10; color index ranges from 13 to 18 in rock of average composition (fig. 2) and 17 to 27 along south and west borders of pluton. Euhedral to subhedral biotite grains distinctive because many are thicker than they are wide. Sphene most abundant accessory mineral; others include apatite, magnetite, zircon, allanite, and tourmaline. Hornblende and biotite from larger pluton gave potassium-argon ages of 99 Ma and 100 Ma respectively (Miller and Clark, 1975; ages recalculated using current IUGS constants, Steiger and Jaeger, 1977). Biotite from smaller pluton gave K-Ar age of 100 Ma (Miller and Engels, 1975; age recalculated using current IUGS constants, Steiger and Jaeger, 1977).

Ksha **Arden pluton**—Leucocratic biotite monzogranite and muscovite-biotite monzogranite (fig. 2). Appearance superficially resembles average Starvation Flat Quartz Monzonite; chief differences are slightly coarser grain size, lower color index, absence of hornblende, and presence of muscovite in Arden pluton. Underlies about 35 km^2 near center of the larger Starvation Flat pluton. Contacts not exposed, but lithologic gradation abrupt over a few meters. Arden pluton is uniform with respect to composition and texture. Rock is hypidiomorphic granular, medium to coarse grained. Average color index 9; contains less than 1 percent muscovite; latter appears to be primary. Accessory minerals include sphene, apatite, garnet, zircon, tourmaline, and allanite

Kfl **Fan Lake Granodiorite (Cretaceous)**—Hornblende-biotite granodiorite and monzogranite (fig. 2); characterized by large stubby hornblende crystals and high color index. Medium- to coarse-grained; hypidiomorphic granular; porphyritic where large hornblende crystals are present. Plagioclase composition averages an_{25}; white potassium feldspar shows no microcline twinning; hornblende and biotite in about equal amounts give rock average color index of 19. Sphene very abundant; other accessory minerals include allanite, zircon, apatite, and opaque minerals. Composition and texture appear to be fairly uniform throughout body. Exhibits no directional fabric even where adjacent two-mica granitic units do, so probably post-dates those units. Hornblende and biotite yield potassium-argon ages of 97 Ma and 95 Ma, respectively (Miller and Engels, 1975; ages recalculated using current IUGS constants, Steiger and Jaeger, 1977)

Kc **Biotite monzogranite of Camden (Cretaceous)**—Medium-grained biotite monzogranite. Partly surrounded by, and may be compositional variant of, Fan Lake Granodiorite. Forms widely separated, poorly exposed, deeply weathered outcrops. Average plagioclase composition an_{25}; some potassium feldspar shows microcline twinning, most does not; biotite is only mafic mineral and gives rock average color index of 12. Accessory minerals include magnetite, apatite, zircon, and minor sphene. Rock heavily epidotized or chloritized at some places. Considered to be Cretaceous in age on basis of textural similarity to, and spatial association with, Fan Lake Granodiorite

Kdc **Two-mica granodiorite of Dubius Creek (Cretaceous)**—Medium- to coarse-grained muscovite-biotite granodiorite and monzogranite. Most of pluton lies east of quadrangle. Contains 2- to 3-cm-long white microperthitic potassium feldspar phenocrysts whose concentration and distribution are very irregular; major part of unit is nonporphyritic. Plagioclase composition averages an_{20}; potassium feldspar is microcline; muscovite forms small crystals averaging about 2 percent of rock; biotite is medium to coarse grained, giving rock average color index of 10. Much of unit contains sparse, pale-yellow-green epidote, probably secondary. Accessory minerals include apatite, zircon, allanite, and minor opaque minerals. Muscovite and biotite from sample collected 2 km east of quadrangle yield potassium-argon ages of 102 Ma and 95 Ma, respectively (Miller and Engels, 1975; ages recalculated using current IUGS constants, Steiger and Jaeger, 1977)

Kgp **Galena Point Granodiorite (Cretaceous)**—Porphyritic, medium- to coarse-grained biotite granodiorite (fig. 2). Phenocrysts of pale-pink microperthitic potassium feldspar average about 4 cm in length; groundmass potassium feldspar shows no microcline twinning. Plagioclase composition averages an_{25}. Biotite is only mafic mineral, forming about 12 percent of rock. Accessory minerals include apatite, zircon, opaque minerals, and minor allanite. Numerous leucocratic dikes in outer part of pluton and in country rocks. Contact-metamorphic aureole extends 1 to 1.5 km from granodiorite; cordierite and andalusite common in metapelitic rocks close to pluton. Body extends 18 km north of quadrangle. Biotite gives potassium-argon age of 101 Ma (Miller and Engels, 1975; age recalculated using current IUGS constants, Steiger and Jaeger, 1977)

Kb **Blickensderfer Quartz Monzonite (Cretaceous)**—Very coarse grained muscovite-biotite monzogranite (fig. 2). Average grain size is about 1 cm. Locally porphyritic, but most rock is hypidiomorphic granular. Average plagioclase composition is an_{20}; white potassium feldspar is microcline; quartz forms large smoky-gray crystals and crystal aggregates; biotite, only slightly more abundant than muscovite, gives rock average color index of 5. Zircon, apatite, and allanite are accessory minerals. Texture and grain size uniform throughout pluton except in localized chilled margins. Muscovite and biotite give potassium-argon ages of 102 Ma and 100 Ma, respectively (Miller and Engels, 1975; ages recalculated using current IUGS constants, Steiger and Jaeger, 1977)

Granodiorite of Hall Mountain (Cretaceous)—Forms six noncontiguous plutons; all except Loop Creek pluton of granodiorite of Hall Mountain are north and east of Chewelah quadrangle. Within quadrangle consists of:

Khlc **Loop Creek pluton**—Muscovite-biotite granodiorite (see fig. 1 and modal diagram, fig. 2). Medium to coarse grained in central part; fine grained near margins. Plagioclase averages an_{30}. Potassium feldspar is microcline; characteristically encloses relatively large plagioclase and biotite, imparting poikilitic look to microcline on stained slab. Biotite only mafic mineral; gives rock average color index of 10. Biotite:muscovite ratio about 4:1. Accessory minerals include epidote, clinozoisite, allanite, zircon, apatite, rutile, and minor opaque minerals. Southern part of pluton cut by numerous, closely spaced quartz veins. Biotite from Hall Mountain pluton located 34 km north of Loop Creek pluton yields potassium-argon age of 99 Ma (Miller and Engels, 1975; age recalculated using current IUGS constants, Steiger and Jaeger, 1977)

Kpl **Phillips Lake Granodiorite (Cretaceous)**—Muscovite-biotite granodiorite; ranges from tonalite to monzogranite (fig. 2). Unit consists of volumetrically large proportion of petrogenetically related dikes, sills, and small stocks of leucocratic monzogranite. Granodiorite, tonalite, and monzogranite are similar in appearance because plagioclase and potassium feldspar are same color. Rocks are medium to coarse grained, irregularly porphyritic, and have poorly formed 2- to 3-cm-long phenocrysts in one-third to one-half of body. Texture of rock is distinctive due to micas interstitial to and, in much of rock, wrapped around felsic minerals; where this texture best developed, rock has crude foliation. Average plagioclase composition an_{22}; potassium feldspar is microcline and microperthitic orthoclase; quartz is smokey to lavender gray; muscovite:biotite ratio averages 1:3; and biotite gives rock average color index of 11. In western part of body, average composition is granodiorite to monzogranite; potassium feldspar is microcline; rock is relatively nonfoliate; and leucocratic dikes, sills, and small stocks form less than 10 percent of volume of rock. Progressively eastward toward Newport Fault, (1) potassium feldspar content decreases and rock composition grades to tonalite, (2) microperthitic orthoclase replaces microcline as potassium feldspar, (3) development of foliation and mylonitization increases, and (4) proportion of leucocratic dikes, sills, and stocks increases to as much as 50 percent of unit. Inclusions and screens of metamorphic rocks ranging from centimeters to several hundred meters in length are common, especially in Phillips Lake Granodiorite south of Winchester Creek (fig. 1). Phillips Lake Granodiorite forms western part of (informal) Priest River complex (Rehrig and others, 1987), which extends 35 km east of quadrangle to Purcell trench (fig. 3). Biotite from Phillips Lake Granodiorite gives potassium-argon age of 94 Ma, and muscovite from related pegmatite gives age of 101 Ma (Yates and Engels, 1968). Potassium-argon ages of both muscovite and biotite decrease progressively eastward toward Newport Fault, where both minerals yield cooling ages of 50 Ma (Miller and Engels, 1975; Harms and Price, 1992)

Kli **Leucocratic intrusive rocks (Cretaceous)**—Medium- to fine-grained monzogranite. Forms small bodies and isolated discontinuous dikes and sills. As mapped, probably includes leucocratic rocks associated with several large Cretaceous plutons. Texture and grain size variable on outcrop scale. Some bodies contain millimeter- to centimeter-long bipyramidal quartz crystals, white altered feldspar, and sparse muscovite or sericite in very fine grained matrix of quartz and feldspar. Isolated exposures in sec. 31, T. 35 N., R. 39 E. are alaskitic, garnet-bearing muscovite monzogranite that intrudes, and is presumed to be related to, two-mica monzogranite body that underlies Quaternary deposits throughout much of area called North Basin (fig. 1). Small body in sec. 23, T. 32 N., R. 39 E. is fine-grained biotite monzogranite considered to be related to nearby Starvation Flat Quartz Monzonite on basis of proximity. Leucocratic dikes range in width from about 1 m to at least 30 m. In sec. 26, T. 32 N., R. 39 E., one dike large enough to show on map intrudes fault between Huckleberry and Buffalo Hump Formations. Assigned Cretaceous age on basis of lithologic similarity to leucocratic dike rocks associated with known Cretaceous intrusive rocks nearby

Knb **Two-mica monzogranite of North Basin (Cretaceous)**—Medium- to coarse-grained, nonporphyritic muscovite-biotite monzogranite and granodiorite. Exposed in only a few deeply weathered outcrops in northwest part of quadrangle, but probably underlies much of area called North Basin (fig. 1). Exposed rock is slightly foliated; has hypidiomorphic-granular texture. Biotite:muscovite ratio estimated to be about 5:1. Color index averages about 12. Biotite has slightly disaggregated appearance possibly caused by shearing in latter stages of emplacement or by uplift in Eocene time. Assigned Cretaceous age on basis of lithologic similarity to nearby Cretaceous two-mica plutons

Knc **Monzogranite of Narcisse Creek (Cretaceous)**—Medium- to coarse-grained biotite monzogranite. Underlies about 10 km^2 in northwestern part of quadrangle; extent beyond quadrangle unknown. Monzogranite of Narcisse Creek (fig. 1) as shown and described by Joseph (1990) on the Colville 1:100,000-scale quadrangle includes at least three separate plutons. Usage in this report is restricted to relatively leucocratic biotite monzogranite found east of Hatch Lake and extending at least to Spruce Canyon Youth Camp located 20 km northeast of Hatch Lake (fig. 1). Monzogranite has average color index of about 10, but is more variable in mafic content than most Cretaceous plutons in region. Plagioclase is calcic oligoclase; potassium feldspar is pinkish-gray microcline; biotite is present as relatively thin 3-mm-wide crystals. Texture is hypidiomorphic granular. Most of rock has no directional fabric, but some has incipient foliation. Biotite from near Spruce Canyon Youth Camp yields K-Ar age of 100 Ma (Miller and Engels, 1975; recalculated using current IUGS constants, Steiger and Jaeger, 1977)

Kbm **Monzogranite of Big Meadows (Cretaceous)**—Muscovite-biotite monzogranite and granodiorite. Fine to coarse grained, irregularly porphyritic; phenocrysts as much as 2 cm. Very heterogeneous texture and composition. Average plagioclase composition an_{25}, but varies widely; potassium feldspar is microcline; muscovite averages 1 to 2 percent of rock; biotite gives rock average color index of 7. Very poor foliation irregularly developed in much of rock. Body extends about 3 km east of quadrangle boundary. Assigned Cretaceous age based on lithologic similarities to Cretaceous plutonic rocks in region

Koc **Monzogranite of Otter Creek (Cretaceous)**—Sillimanite-bearing muscovite-biotite monzogranite. Forms a single body located northeast of Eloika Lake. Medium grained, nonporphyritic. Most of rock has poorly developed subtle foliation that is apparent in thin section; much feldspar is milled and quartz is milled and flattened. Mica is bent and wrapped around quartz and feldspar. Sillimanite spatially associated with mica or replaces mica. Plagioclase averages an_{25}; some potassium feldspar is microcline, much shows no microcline twinning. Muscovite and biotite contents extremely variable; color index ranges from about 5 to 15. Assigned Cretaceous age on basis of lithologic similarity to Cretaceous rocks in region; probably predates 97 Ma Fan Lake Granodiorite

Klgs **Leucocratic granitic rocks of Scotia (Cretaceous)**—Heterogeneous mixture of alaskite, pegmatite, aplite, and two-mica monzogranite; also includes pods and screens of metamorphic rocks derived from Middle Proterozoic Prichard Formation and mafic sills. All granitic rocks are leucocratic and contain muscovite and small amounts of biotite. Plagioclase composition ranges from albite to oligoclase; potassium feldspar is mostly microcline. Grain size ranges from fine to very coarse. Foliation, lineation, ductile deformation, and cataclasis common in eastern part of unit. All contacts highly gradational. Age of granitic rocks considered to be Cretaceous, based on lithologic similarity to Cretaceous leucocratic two-mica plutonic rocks associated with Phillips Lake Granodiorite

Ksv **Granodiorite of Spring Valley (Cretaceous)**—Biotite granodiorite; contains very sparse muscovite that may or may not be primary. Medium to coarse grained, nonfoliate to slightly foliate. Shows slight to moderate ductile deformation and cataclasis. Plagioclase is oligoclase; potassium feldspar shows no microcline twinning; biotite content averages 12 percent. Contains very abundant allanite in addition to zircon, apatite, and opaque minerals. Erroneously reported to cut Fan Lake Granodiorite (Miller, 1974c); contact with Fan Lake could be fault. Considered to be Cretaceous in age on basis of lithologic similarity to nearby Cretaceous granitic rocks

Kbr **Two-mica monzogranite of Blanchard Road (Cretaceous)**—Medium- to coarse-grained porphyritic muscovite-biotite monzogranite. Distinguished by muscovite phenocrysts 2 to 3 cm long; also contains irregularly distributed potassium feldspar phenocrysts of about same size. Plagioclase is oligoclase; only part of potassium feldspar shows microcline twinning; biotite is only mafic mineral, giving rock average color index of 8. Muscovite:biotite ratio ranges from 1:5 to 1:1. Much of rock, especially southeastern part of body, exhibits well-developed foliation and lineation caused by ductile deformation related to development of Eocene metamorphic core complex(es) in region. Assigned Cretaceous age on basis of lithologic similarity to nearby Cretaceous two-mica plutonic rocks

KJcc **Hornblende-biotite quartz diorite of Cusick Creek (Cretaceous or Jurassic)**—Medium- to coarse-grained, highly mafic quartz diorite and tonalite. Hypidiomorphic granular, but some cataclasis near Newport Fault. Plagioclase is andesine. Potassium feldspar shows no microcline twinning; content highly variable. Most biotite and hornblende is altered; hornblende has cores of pyroxene. Color index ranges from 15 to 35. Owing to wide range in potassium feldspar content and color index, pluton may be a hybrid. Extent of body north of quadrangle unknown. Considered Jurassic or Cretaceous in age because rock has lithologic similarities to plutons of both ages in region

Jlm **Quartz monzodiorite of Lane Mountain (Jurassic)**—Medium- to coarse-grained, highly porphyritic biotite-hornblende quartz monzodiorite; composition ranges to granodiorite and quartz monzonite (fig. 2). Subhedral to euhedral sodic andesine grains average about 8 mm in length. Quartz forms rounded crystals and crystal aggregates about 1 cm across and smaller interstitial grains between other minerals. Almost all potassium feldspar occurs as 1.5- to 9-cm-long, smooth-sided, euhedral, prismatic phenocrysts of orthoclase that contain patches of microcline; potassium feldspar is conspicuously lacking in groundmass. Ratio of hornblende to biotite averages about 3:2, but ranges from rocks with about 1:3 to rocks without biotite. Sphene is very abundant and obvious even in outcrop. Other accessory minerals include zircon, apatite, allanite, and opaque minerals. Texture differs from that of other plutonic rocks in region, except for Eocene Silver Point Quartz Monzonite, by its trimodal grain size: large potassium feldspar phenocrysts, intermediate-sized felsic minerals, and small mafic minerals mixed with quartz. Small mafic minerals and quartz are interstitial to relatively large felsic minerals. Concentration of phenocrysts ranges from only a few to over 100 per square meter. Other than variation in phenocryst concentration and ratio of hornblende to biotite, both texture and mineralogy uniform throughout pluton. Hornblende and biotite from rock near center of sec. 23, T. 31 N., R. 39 E. yielded potassium-argon ages of 161 Ma and 162 Ma, respectively (R. J. Fleck, U.S. Geological Survey, written commun., 1988)

}|f **Fault-zone rocks (middle Mesozoic to late Paleozoic)**—Highly sheared and deformed argillite, phyllite, carbonate-bearing phyllite, and dolomite. Rocks cut by close-spaced fractures; nearly all rock between fractures is sheared. Almost no sedimentary features preserved. Deformed sedimentary rocks probably derived mostly from lower and middle parts of Deer Trail Group. Fractured quartz veins that contain numerous fragments of phyllite and dolomite are abundant; some veins are gold and copper bearing (Patty, 1921; Clark and Miller,1968)

J^ft **Flowery Trail Granodiorite (Jurassic or Triassic)**—Hornblende-biotite quartz monzodiorite and quartz monzonite; ranges to granodiorite and monzonite (fig. 2). Fine to coarse grained, but is even grained on outcrop scale. Preferred alignment of minerals widely variable on outcrop scale; may be primary. Average plagioclase composition an_{30}; potassium feldspar is microcline and microperthitic orthoclase; average quartz content 11 percent, much lower than Cretaceous plutons; hornblende:biotite ratio greater than 2:1; average color index 31. Composition and texture highly variable throughout body; composition more alkalic than Jurassic or Cretaceous plutons. East end of pluton metamorphosed by Cretaceous Phillips Lake Granodiorite. Hornblende and biotite give potassium-argon ages of 198 Ma and 100 Ma respectively (Miller and Clark, 1975; recalculated using current IUGS constants, Steiger and Jaeger, 1977). Amount of discordance suggests emplacement age older than 205 Ma

M_u **Carbonate and clastic sedimentary rocks, undivided (Mississippian to Cambrian)**—Limestone, dolomite, and carbonaceous shale. Probably includes parts of

Metaline Formation, Ledbetter Formation, and unnamed Devonian and Mississippian units. Rocks mapped as unit M_u are either fault bounded or surrounded by alluvium; undetected faults are probably present between isolated outcrops of unit. Exposures in northwestern part of quadrangle, 7 km west of Arden, include carbonaceous shale that contains interbedded centimeter- to meter-thick, dark-gray, carbonaceous limestone beds. There, limestone and shale are laminated to thinly parallel planar bedded and contain sparse chert beds as much as 1 m thick; some chert occurs as fault-bounded pods surrounded by intensely fractured and cleaved carbonaceous shale. East of Addy Mountain (fig. 1), rocks assigned to this unit consist of pale-gray, brecciated dolomite that could be part of Devonian and (or) Mississippian units MD_3, MD_2, MD_1, and Ddl but more likely belong to Metaline Formation. In southern part of Colville Valley (fig. 1), most outcrops of this unit are nondistinctive, medium-gray limestone and dolomite; Devonian and Mississippian fossils are reported from a few of these outcrops east of Jumpoff Joe Fault (Miller and Clark, 1975). Much of unit is extremely brecciated and bedding features are obliterated

Ml **Limestone (Mississippian)**—Limestone and lesser amounts of dolomitic limestone and dolomite. Rock is medium to fine grained, medium to pale gray on fresh surface; weathers medium gray. Bedding thickness ranges from 0.5 cm to more than 4 m; defined by faint, parallel-planar lamination on weathered surfaces and by 0.5- to 3-cm-thick irregularly discontinuous chert bands. Some chert forms irregularly shaped masses up to 0.5 m thick. Chert is white, gray, and pink. Medium-grained marble that contains scattered tremolite crystals is found near Cretaceous granitic rocks. All rock extremely fetid. Thickness east of Springdale is about 200 m, but upper and lower contacts of unit not exposed and section could be internally faulted. Mississippian age assignment based on conodonts *Bactrognathus* sp., *Hindeodus* aff. *H. cristulus* (Youngquist and Miller), *Hindeodus* aff. *H. crassidentatus* (Branson and Mehl), *Kladognathus* sp. indet., and *Ozarkodina* sp. indet. (Waggoner, 1990)

MD_3 **Dolomite and slate (Mississippian and (or) Devonian)**—Light-gray and cream-colored dolomite interbedded with maroon and pale-green argillite. Lower 15 m of unit is thin- to thick-bedded, saccharoidal to aphanitic, pale-orange-weathering dolomite. Pale-orange-weathering dolomite overlain by about 75 m of apparently lithologically similar dolomite that contains interbeds of pale-gray-green argillaceous dolomite and maroon slaty argillite. Bed thickness of argillaceous rocks ranges from 1 to 50 cm; maroon argillite beds become thicker and more numerous in upper part of 15- to 75-m interval. Upper part of this interval consists of about 30 m of pale-green and maroon argillite. All argillite is well laminated and contains thin seams of dolomite. Upper 50 m of unit MD_3 is dolomite lithologically similar to that in lower 15 m of unit. Thickness of unit appears to be about 170 m but is uncertain owing to possible internal faulting. Conformably overlies unit MD_2. Upper contact of unit either eroded or faulted. Age assignment is based on presumed stratigraphic position below fossiliferous Mississippian limestone and above fossiliferous Devonian limestone

MD_2 **White and pale-gray dolomite (Mississippian and (or) Devonian)**—White and pale-gray dolomite. Beds range from a few centimeters to 2 m in thickness; average about 1 m and show faint internal laminations. Lower 5 m of unit contain abundant oolites. Most of unit is coarse grained, possibly due to recrystallization. Contains no fossils except for sparse nondefinitive algae. Unit appears to be 150 to 200 m thick, but could be faulted internally. Conformably overlies unit MD_1. Age assignment is based on presumed stratigraphic position below fossiliferous Mississippian limestone and above fossiliferous Devonian limestone

MD_1 **Dark-gray dolomite (Mississippian and (or) Devonian)**—Medium- to fine-grained, dark-gray dolomite; locally mottled light gray. Contains sparse quartz sand grains throughout. Some dolomite appears argillaceous, but none is even slightly limy, nor does unit contain any interbedded limestone. Well-defined bedding ranges from a few centimeters to almost 2 m in thickness. Oolites common, confined to distinct beds. Some interbeds of matrix-supported conglomerate contain angular to slightly rounded clasts of dark-gray dolomite. Thickness of unit estimated to be between 180 and 210 m, but base not exposed. Age assignment based on presumed stratigraphic position below fossiliferous Mississippian limestone and above fossiliferous Devonian limestone

Ddl **Dolomite and limestone (Devonian)**—Light-gray and cream-colored dolomite interbedded with medium-gray limestone. Some limestone is punky owing to intricate solution cavities; may be argillaceous or carbonaceous. Very poorly exposed; stratigraphic relations and thickness uncertain. Contains Late Devonian brachiopods *Cytrospirifer* sp. and *Tenticospirifer* (Miller and Clark, 1975) and Late Devonian conodonts *Palmatolepsis quadrantinodosa inflexa* Muller, *Pelekysgnathus*? sp. indet., *Polygnathus semicostatus* Branson and Mehl, and *Polygnathus* sp. indet.

Ol **Ledbetter Formation (Ordovician)**—Dark-gray carbonaceous shale and slate and minor carbonaceous limestone and chert interbeds. Carbonaceous shale and slate range from massive to very faintly laminated. Nearly all is highly cleaved; cleavage at various angles to bedding, but rarely parallel to bedding. In probable upper part of Ledbetter Formation, intervals of shale tens of meters thick are highly siliceous; some or all may be bedded chert. Siliceous rock in these zones is medium to dark gray and thin bedded to laminated. Limestone beds range from 1 cm to 1 m thick; all are medium to dark gray and carbonaceous. Chert is massive to laminated, most is brecciated. Large chert masses lacking internal stratification occur as fault-bounded pods. Transition from Ledbetter Formation to underlying limestone of Metaline Formation appears to be abrupt. Upper contact of unit everywhere faulted. Internal stratigraphy poorly understood due to discontinuous poor exposures, extreme internal deformation, lack of marker units, and lack of visible stratification in much of formation. Internal deformation includes single and multiple cleavage(s), small-scale folds, and faults that have unknown sense and amount of offset. Ledbetter at its type locality differs from formation in Chewelah quadrangle. Thin limestone beds are concentrated in lower part of formation at type locality 55 km northeast of quadrangle (Park and Cannon, 1943) but are sparsely scattered throughout formation in Chewelah quadrangle. Apparent thickest section of Ledbetter is 550 m as calculated from outcrop width; however, upper contact of unit is fault, and degree of thickness distortion by internal deformation is unknown. Ordovician age based on abundant graptolites (Park and Cannon, 1943; Carter, 1989a, 1989b), conodonts (Hogge, 1982), and trilobites, (Schuster, 1976)

0_gc **Phyllite and quartzite of Gardiner Creek (Ordovician or Cambrian)**—Medium- to dark-gray phyllite, white to brown vitreous quartzite, and minor interbeds of dark-brown, sandy dolomite. Unit is about 60 percent phyllite and 40 percent quartzite. Occurs as alternating zones of indistinctly bedded phyllite up to 60 m thick and thick-bedded quartzite up to 30 m thick. Sandy dolomite is restricted to quartzite zones. Correlation with other units in quadrangle questionable. Except for dolomitic beds, unit lithologically most resembles upper member of Late Proterozoic and Cambrian Addy Quartzite (_au); however, unit may overlie Metaline Formation and be Ordovician in age

Metaline Formation (Ordovician and Cambrian)—Limestone, dolomite, shaly limestone, and carbonate-bearing quartzite. Metaline Formation forms a pure and impure carbonate rock sequence lying between Ledbetter Formation and Late Proterozoic and Cambrian Addy Quartzite. Metaline Formation in Chewelah quadrangle appears to differ markedly from unit at type locality. Even within the quadrangle this formation differs from place to place, possibly indicating juxtaposition of unlike lithofacies by thrust faults or abrupt lateral changes in lithofacies. In the quadrangle, Metaline Formation is subdivided into four informal members, one of which contains a locally subdivided interval of dolomite beds. However, due to faulting and discontinuous exposure, internal stratigraphy of informal members and of formation as a whole is poorly understood. At type locality, 60 km northeast of Chewelah quadrangle, an approximately 2,650-m-thick section of Maitlen Phyllite lies between Gypsy Quartzite (lateral equivalent of Addy Quartzite) and Metaline Formation (Park and Cannon, 1943; Burmester and Miller, 1983). Placement of lower contact of formation in Chewelah quadrangle places most of rocks comprising Maitlen Phyllite as defined by Lucas (1980) in lowest member (O_mq) of Metaline Formation. Unit designated by Lucas (1980) as Maitlen in Dunn Mountain area is an order of magnitude thinner than and lithologically different from type Maitlen (Park and Cannon, 1943). Maitlen Phyllite as defined by Lucas (1980) is not adopted here because its rocks could not be distinguished from rocks of thick argillite zones in upper member of Addy Quartzite. Metaline consists of:

O_mu **Undivided part**—Limestone and shaly limestone. Large area south of Dunn Mountain (fig. 1) is mapped as undivided limestone and shaly limestone partly because internal folding, incomplete exposure, and probable undetected faults do not allow confident subdivision and partly because rocks in that area appear to have some lithologic and bedding characteristics that differ from rocks of informal members (Miller, 1996b). In this area, shaly limestone and subordinate pure limestone have parallel-planar and irregular bedding, respectively. Color ranges from pale gray to dark gray, weathers medium to light gray to gray tan. Shaly limestone contains high concentration of argillaceous material and resembles shaly limestone member (O_ms). Pure limestone beds that have highly irregular bed thickness ranging from a few centimeters to about 1 m are scattered throughout sequence and are unlike limestone beds of informal members. Southeast of Springdale, undivided part of Metaline Formation appears to conformably overlie Addy Quartzite and consists of thick- to thin-bedded gray to blue-gray limestone that has irregularly shaped, yellow-brown-weathering argillaceous seams irregularly interwoven through the rock. North of Jared (fig. 1), in northern part of quadrangle, light- and medium-gray, thick-bedded, brecciated, and recemented dolomite is probably part of Metaline Formation. Rocks there, shown as lower dolomite and upper dolomite by Miller (1974b), may correspond to dolomite member of Metaline Formation at its type locality and in western part of Chewelah quadrangle

O_ms **Shaly limestone member**—Thin-bedded shaly limestone and calcareous shale. Both upper and lower contacts of unit are sharp or gradational over interval of no more than 10 to 15 m. Limestone content of rocks ranges from nearly pure to nearly absent; rocks at pure end of range are rare. Many beds, especially in upper part of unit, are distinctly carbonaceous, and many bedding surfaces have carbonaceous and argillaceous films. Regardless of carbonate content, all rocks are extremely fine grained. Most bedding is parallel planar. Bed thickness ranges from about 5 cm to submillimeter laminations; averages about 1 cm. Thin-bedded rocks are commonly fissile; argillaceous rocks are slightly phyllitic. Color ranges from pale to dark gray; most rocks of unit weather pale gray or pale grayish-tan. Member in most places highly cleaved and folded; fold amplitudes range from millimeters to tens of meters

O_md **Thick-bedded dolomite member**—Massively to thinly bedded, coarse-grained dolomite. Most of member is relatively pure dolomite in beds up to 5 m thick; average bed thickness about 2 m. Bedding difficult to recognize in much of member. Color is white to pale gray and commonly contains diffuse, discontinuous, dark-gray streaks, most of which appear to be roughly parallel to bedding. Grain size of almost all rock is coarse to very coarse. Near Starvation Flat Quartz Monzonite, rock is very coarse grained, white, dolomitic marble and contains sparsely scattered acicular tremolite. Apparent thickness, calculated from outcrop width, is about 800 m, but upper contact of member is a fault

O_ml **Thick- and thin-bedded limestone member**—Interlayered, thin-bedded, fine-grained limestone; thick-bedded, coarse-grained limestone; and limestone conglomerate. Typically 1- to 3-m-thick intervals of centimeter-thick, parallel-planar limestone beds separated by 1- to 10-m-thick intervals of limestone conglomerate. Limestone is medium gray, blue gray, and dark gray and weathers pale blue gray. In thin-bedded intervals, bed thickness ranges from submillimeter laminations to about 8 cm; average is about 2 cm. In thick-bedded intervals, bed thickness is irregular, although most contacts with thin-bedded zones are nearly planar. Conglomerate consists of rounded and angular clasts of limestone that range in size from pebbles to boulders in a matrix of fine- to coarse-grained, gray limestone. Conglomerate is both clast and matrix supported; some beds of edgewise conglomerate. In many thick beds, clasts are bent, apparently at time sediments were soft. Many beds, especially in lower 100 m of member, contain films of tan to orange argillaceous limestone anastomosing through blue-gray limestone. Member may be as thick as 680 m (including dolomite beds of unit O_mld). Locally, includes:

O_mld **Dark-gray dolomite beds**—Coarse-grained dolomite and dolomite breccia. One- to 3-cm-thick, parallel-planar-bedded dolomite intervals are as much as 2 m thick. Parallel-planar-bedded intervals separated by dolomite sedimentary breccia. Clasts in breccia are largely sharp-edged tabular fragments as long as 6 cm; most are matrix supported. Some clasts have irregular shape or are rounded. Thin, sinuous, discontinuous seams of white, coarse-grained dolomite present in many dark-gray beds; probably late or postdiagenetic. Maximum thickness of dolomite beds calculated from outcrop width is about 210 m, but may be thickened by faulting

O_mq **Limestone and carbonate-bearing quartzite member**—Limestone, sandy limestone, carbonate-bearing quartzite, pebble conglomerate, and argillite. Lower contact of member is base of lowest carbonate-bearing quartzite bed above interbedded quartzite and argillite of upper member of Addy Quartzite. Upper contact of member is top of highest quartz-grain-bearing limestone bed. Lower part of limestone and carbonate-bearing quartzite member includes at least two intervals of white, carbonate-cemented quartzite up to 10 m thick separated by interval of brownish-gray argillite. At many places, carbonate cement in quartzite is leached, producing friable or porous, yellow-stained quartzite. Quartzite intervals overlain by thick argillite zone, part of which has thin, wavy, discontinuous lenses of quartzite. In Dunn Mountain area, argillite zone is overlain by distinctive, 2-m-thick bed of carbonate-cemented quartz-pebble conglomerate; extent of bed beyond Dunn Mountain area unknown. Pale-gray limestone that contains abundant lensoidal trains of matrix-supported, coarse-grained quartz sand forms upper part of member. This limestone interbedded with lesser thin argillite beds and possibly another 5-m-thick interval of white, carbonate-cemented quartzite. Thickness of member as calculated from outcrop width about 150 m

Addy Quartzite (Cambrian and Late Proterozoic)—Vitreous quartzite that has lesser amounts of interbedded siltite and argillite. In some parts of quadrangle, subdivided into four informal members following usage of Miller (1983) and Lindsey and others (1990): (1) upper member, (2) coarse-grained member, (3) purple member, and (4) lower member. Unconformably overlies Late Proterozoic Windermere Group with very slight angularity detectable only on regional scale. Unconformity cuts through Windermere Group into Middle Proterozoic Deer Trail Group about 35 km southwest of quadrangle. Thickness of Addy as calculated from outcrop width averages about 1,400 m west of Jumpoff Joe Fault, but may be much less east of fault. Consists of:

_zau **Undivided part (Cambrian and Late Proterozoic)**—White, purple, pink, gray, and tan vitreous quartzite interbedded with lesser amounts of gray, greenish-gray, purplish-gray, and brownish-gray argillite and siltite. Lower 10 percent of this undivided unit is dominantly white, thick-bedded quartzite; grades upward into several-hundred-meter-thick interval of purple, pink, and white crossbedded quartzite that contains dark-gray to black bands. Purple-hued quartzite interval overlain by slightly thinner interval of white, coarse-grained, crossbedded quartzite, which in turn is overlain by interval of interbedded, white and gray, medium-grained quartzite that is interlayered with argillite and siltite zones up to 10 m thick. On hill south of Springdale, in Fan Lake area, in Leslie Creek area east of Addy Mountain, and on mountain north of Jared (fig. 1), Addy Quartzite is moderately to highly recrystallized by Cretaceous plutons

_au **Upper member (Cambrian)**—Interbedded vitreous quartzite and argillite and minor siltite. Argillite zones average about 5 m thick and separate quartzite zones of equal or greater thickness. Some argillite and siltite contains discontinuous lenses of fine-grained, wavy-bedded quartzite. Much argillite has only faint bedding or totally lacks internal bedding features. Colors of argillite are, in order of decreasing abundance, purplish gray, gray, greenish gray, and brownish gray. Quartzite beds are white and gray, planar and wavy bedded; some are cross bedded. Most quartzite is medium grained; grain size and thickness of quartzite beds decrease upsection in this member, and thicknesses of argillite zones increase. Thickness of member on Dunn Mountain as calculated from outcrop width is 425 m. At north end of town of Addy near center of sec. 13, T. 33 N., R. 39 E., member contains Early Cambrian trilobite *Nevadella addyensis* and brachiopod *Kutorgina* sp. (Okulitch, 1951). Fossils appear to be in lower part of member in thinly interbedded argillite and quartzite. Thin discontinuous zone of carbonate-bearing siltite present just above fossil zone. Lindsey and others (1990) note body and trace fossils are abundant in lower part of upper member

_ac **Coarse-grained member (Cambrian)**—White, gray, and pink, medium- and coarse-grained quartzite. Informal member defined by Lindsey and others (1990). Contains matrix-supported pebbles throughout and lenses as much as 5 cm thick of clast-supported pebbles locally. Pebbles, as large as 3 cm in diameter, average about 1 cm in diameter and are well rounded. Abundance of pebbles and general coarseness of quartzite decreases in upper part of member. Most quartzite in this member is poorly to moderately well sorted. Sparse, pale-greenish-gray, brownish-gray, and purplish-gray argillite or siltite interbeds generally less than 10 cm thick; argillite is common as bedding-plane partings and films at top of beds. Beds of quartzite 10 cm to 2 m thick form blocky-weathering outcrops. Crossbedding on medium and large scales is fairly abundant. Thickness on Dunn Mountain as calculated from outcrop width averages 370 m, but thickness of member appears to vary laterally more than other members of Addy Quartzite

_Zap **Purple member (Cambrian and Late Proterozoic)**—Black- and dark-gray-striped purple, pink, gray, and white medium-grained quartzite and minor thin siltite and argillite beds. Dark stripes and reddish hues of quartzite are distinctive. Stripes both parallel and crosscut bedding; dark stripes are trains of disseminated hematite and hematite-coated quartz grains. In Iron Mountains (fig. 1 and sec. 9, T. 33 N., R. 40 E. on map), in lower part of member, several beds 30 to 60 cm thick contain about 30 percent hematite. Overall, bed thickness of member ranges from 2 cm to more than 1 m. Abundant crossbedding is typically highlighted by dark stripes. Thickness of member on Dunn Mountain as calculated from outcrop width is 460 m. Unit is one of most distinctive stratigraphic markers in region

Zal **Lower member (Late Proterozoic)**—White vitreous quartzite and minor interbedded argillite. Almost all quartzite is white and medium to fine grained. Beds are thick to massive and show little internal stratification. Crossbedding observed at only a few localities. Member thickens to the northeast (Miller, 1983; Groffman, 1986; Lindsey and others, 1990) and thins to the southeast (Miller and Clark, 1975). On Dunn Mountain, average thickness of member as calculated from outcrop width is about 150 m. Although not subdivided east of Jumpoff Joe Fault, member is represented by only 3 m of white quartzite on Quartzite Mountain

Windermere Group—In map area, consists of Monk Formation and Huckleberry Formation; latter formation divided into upper greenstone member and lower conglomerate member (Campbell and Loofbourow, 1962). Unconformities bound Windermere Group at top and bottom. Conglomerate and greenstone members of Huckleberry Formation are laterally equivalent to Shedroof Conglomerate and

Leola Volcanics, respectively, in Metaline area 70 km to northeast (Park and Cannon, 1943) and to Toby Conglomerate and Irene Volcanics, respectively, in southern British Columbia (Reesor, 1957; Miller and others, 1973). Windermere is not found east of Jumpoff Joe Fault. Consists of:

Monk Formation (Late Proterozoic)—Conglomerate, megabreccia, diamictite, feldspathic and lithic quartzite, siltite, and argillite. Extremely variable lithostratigraphy from place to place. Formation on Empey Mountain (fig. 1) informally subdivided into three members; other places mapped as undivided part (Zmu) or as conglomerate member (Zmc). Consists of:

Zmu **Undivided part**—Argillite, siltite, dolomite, and conglomerate. In Iron Mountains, lower 15 m are dark-gray to greenish-gray conglomerate, both matrix- and clast-supported. Clasts are well rounded, some nearly spherical; average size is about 6 cm. Most clasts are dark-gray, andesitic-looking volcanic rock. Matrix is siltite and argillite that contains sparse, matrix-supported, round, millimeter-size quartz grains. Locally, rip-ups of argillite enclosed within conglomerate. Conglomerate rests directly on greenstone member of Huckleberry Formation in secs. 5 and 8, T. 33 N., R. 40 E.; may pinch out northeastward and southwestward. Conglomerate is overlain successively by discontinuously exposed (1) white and tan dolomite, (2) thin-bedded, carbonate-bearing siltite interlayered with thin beds of argillite, and (3) minor fine-grained quartzite. Carbonate content of siltite appears to diminish upsection. In Iron Mountains, upper part of Monk Formation is finely laminated grayish-maroon and greenish-gray siltite and argillite. Same rock types are found in partial sections of undivided part of Monk Formation east of Iron Mountains near Bayley Creek and southwest of Chewelah. In those sections, conglomerate clasts are quartzite, dolomite, and argillite derived from Middle Proterozoic Deer Trail Group. Monk is unconformably overlain by Addy Quartzite. Thickness of Monk variable owing to depth of pre-Addy erosion and possibly to syndepositional faults. Apparent thickest section on east flank of Iron Mountains is about 300 m as calculated from outcrop width

Zma **Argillite member**—On Empey Mountain (fig. 1), occurs as two distinct argillite zones separated by greenstone (Zmg) and conglomerate (Zmc). Upper argillite zone is medium- and pale-gray argillite and siltite. In places, argillite is pale green or khaki; latter may be either normal pigmentation or alteration color. Parallel-planar laminated, but laminations are indistinct. Some argillite beds in upper zone are characterized by sparse, matrix-supported quartz grains 1 mm in diameter and are interbedded with minor amount of fine-grained quartzite in 1- to 5-cm-thick beds. Lower argillite zone is argillite and siltite and contains minor thin quartzite interbeds in multiple fining-upward sequences. Parallel planar bedded; dark to light gray. Lithology of lower argillite zone anomalous compared to other sections of Monk Formation in map area. Assigned to Monk Formation, but could also represent as yet unrecognized strata of Deer Trail Group stratigraphically above Buffalo Hump Formation. Upper contact of argillite member is unconformity; up to 250 m preserved on Empey Mountain as calculated from outcrop width

Zmc **Conglomerate member**—Conglomerate, megabreccia, diamictite, and feldspathic and lithic quartzite and siltite. Lithology of unit ranges greatly over short distances; probably reflects close proximity of highly contrasting depositional environments created by syndepositional faulting. Conglomerate member 4.5 km west-northwest of Waitts Lake is mostly megabreccia that consists of angular dolomite blocks ranging in size from a few centimeters to over 200 m. Blocks are in matrix of gray and tan siltite and conglomeratic siltite with some argillite and lithic quartzite. Blocks are randomly oriented, many wrapped in matrix sediments as if blocks slid into unconsolidated, water-saturated sediment. Cobble-sized clasts in matrix are mixed lithologies and well rounded. Ratio of breccia blocks to matrix varies greatly over short distances. Clasts in conglomerate member on west side of Lane Mountain are sparse and matrix is tan and khaki siltite and argillite. On east side of Empey Mountain, conglomerate member below greenstone member (Zmg) of Monk Formation consists of interbedded conglomerate, conglomeratic lithic arkose, siltite, and argillite in parallel-planar, multiple, fining-upward groups of beds. Conglomerate member above greenstone member is green, matrix-supported diamictite. On Empey Mountain, contact with argillite member (Zma) appears gradational; contact with Deer Trail Group is unconformity. Thickness of conglomerate member variable due to syndepositional faulting, Mesozoic faulting, pronounced facies changes over short distances, and pre-Paleozoic erosion. About 300 m of member preserved west-northwest of Waitts Lake

Zmg **Greenstone member**—Altered basalt flow rocks and minor volcanic breccia. Rocks lithologically indistinguishable from greenstone (Zhg) in underlying Huckleberry Formation. Collapsed vesicles abundant at some localities. Appears to grade into conglomerate member (Zmc) of Monk Formation, which lies above and below. Thins southward; not recognized in Monk sections north of Empey Mountain

Huckleberry Formation (Late Proterozoic)—Extrusive and intrusive rocks of basaltic composition, conglomerate, and diamictite; all basaltic rocks metamorphosed to greenstone. Informally subdivided into greenstone member (upper) and conglomerate member (lower) (Campbell and Loofbourow, 1962). Intrusive greenstone related to flows of greenstone member; thin discontinuous zones of conglomerate are present in greenstone member. Consists of:

Zhg **Greenstone member**—Greenstone derived from basaltic lava flows, tuff, volcanic breccia, and volcaniclastic rocks. Thicknesses of flows range from a few meters to several tens of meters. Individual flows poorly defined due to lack of lithologic contrast across flow boundaries and by nearly total lichen cover. Some flows separated by flow breccia consisting of light-green, angular clasts, averaging 5 mm to 5 cm across, in dark-green matrix. Pillow structures fairly abundant, especially in lower part of member, but pillows subtle and difficult to recognize. Composition is tholeiitic basalt (Miller and Clark, 1975). Flow rocks contain 1-mm-long pyroxene and plagioclase crystals set in microcrystalline and (or) glassy groundmass altered to chlorite, quartz, albite, calcite, epidote, and opaque minerals. All plagioclase in rock altered to albite and calcite; pyroxene conspicuously unaltered. Much of member is phyllitic; phyllitic parts are probably tuffaceous, volcaniclastic, or are finely brecciated flow rocks. On Gold Hill (fig. 1) and south flank of Deer Mountain, member is thin and, in addition to greenstone, includes conglomerate that has matrix of chlorite-green volcanic material and diamictite. A few meters of argillite and dolomite included as part of member on Gold Hill may actually be Monk Formation instead. Thickness of greenstone member in Huckleberry and Iron Mountains as calculated from outcrop width averages 975 m, but upper contact of member is an unconformity or lower contact is a fault. Thins eastward; not present east of Jumpoff Joe Fault. Locally includes:

Zhgc **Volcanic conglomerate**—Poorly rounded to well-rounded clasts of volcanic rocks in matrix of medium- to fine-grained volcaniclastic material. Unit differentiated only west-northwest of Waitts Lake and in limited area northeast of Stensgar Mountain. All is matrix supported and bounded by tuffaceous or flow rocks of greenstone member of Huckleberry Formation

Zhi **Intrusive greenstone**—Lithologically similar to massive flow rocks of greenstone member, but most rocks are coarser grained; some are gabbroic. Forms several large masses in Huckleberry Creek and Service Creek drainages (fig. 1). Also forms numerous dikes and sills too small to show on map

Zhc **Conglomerate member**—Diamictite, conglomerate, sandy siltite and argillite, and lithic quartzite. Pale-green and pale-gray, matrix-supported diamictite and conglomerate are most common lithologies. Except in lower 50 m, clasts are sparse and, in some areas, nearly absent. Clast types include dolomite, argillite, siltite, quartzite, and milky quartz; all but latter derived from units of Deer Trail

Group. Most matrix material is sand- or silt-size quartz, feldspar, sericite, chlorite, and abundant carbonate minerals and lithic material. Matrix-supported, well-rounded, millimeter-size quartz grains sparsely, but ubiquitously, scattered throughout rock. Generally, rock is phyllitic and has reflective silvery surfaces. Phyllitic foliation defined by parallel mica and flattened grains and clasts; plane of flattening commonly oriented no more than 10° or 20° from bedding. Bedding is poorly developed in most of member or was destroyed by development of schistosity; bedding identified with confidence only where rare lithic quartzite beds are found. Maximum thickness of conglomerate member as calculated from outcrop width at Huckleberry Mountain is 480 m. Thins and pinches out eastward and northeastward

Deer Trail Group—Deer Trail Group crops out exclusively in north-northeast-striking, highly deformed belt known as magnesite belt (Campbell and Loofbourow, 1962). Northern two-thirds of magnesite belt lies within Chewelah quadrangle. From youngest to oldest, Deer Trail Group consists of Buffalo Hump Formation, Stensgar Dolomite, McHale Slate, Wabash Detroit Formation, Chamokane Creek Formation, and Togo Formation (Miller, 1996a). Base of Togo Formation not exposed. Upper contact of Deer Trail Group is faulted everywhere except 3 km south of Chewelah, where rocks of Windermere Group lie unconformably on Stensgar Dolomite, and 6 km north-northwest of Chewelah where unconformity cuts downsection so that Windermere rocks lie on McHale Slate. Deer Trail Group is lithostratigraphic equivalent of part of upper part of Belt Supergroup (Miller and Whipple, 1989). Consists of:

Ydtu **Undivided part (Middle Proterozoic)**—Argillite, phyllite, and quartzite. Includes five small isolated areas of outcrop north and south of Empey Mountain in southwestern part of quadrangle where very sparse exposures are highly sheared and deformed; also includes highly recrystallized schist west of Bayley Lake (fig. 1). Rocks are nondistinctive and could not be assigned to specific formations of Deer Trail Group

Buffalo Hump Formation (Middle Proterozoic)—Formation is informally subdivided into quartzite and argillite units, but there appears to be two distinct zones of each lithology: lower white quartzite zone overlain by two argillite zones that are separated by thinner quartzite zone. Both rock types show considerable variation in appearance from one area to another and possibly vary in relative abundance. Quartzite appears to form most of formation at each end of magnesite belt (fig. 1); argillite appears to predominate in central part (Miller and Whipple, 1989). Variability in relative amounts of each lithology probably more apparent than real and results more from faulting, folding, and poor exposure than from changes in lithofacies. However, Campbell and Loofbourow (1962), Campbell and Raup (1964), and Evans (1987) consider lithologic variability to result from changes in lithofacies. Maximum thickness of Buffalo Hump Formation about 550 m as calculated from outcrop width, but this figure is only an approximation due to internal deformation, especially in argillite zones. Separated from overlying Windermere Group by erosional unconformity; amount of formation removed by pre-Windermere erosion is unknown. Contact with underlying Stensgar Dolomite not exposed but located within 10 m at many places. Consists of:

Ybu **Undivided part**—Mapped as undivided rocks only on The Island, south of town of Valley. Chiefly laminated argillite and siltite. Even-parallel to nonparallel bedding and lamination; soft-sediment deformation abundant. Formational assignment based on stratigraphic position above Stensgar Dolomite

Yba **Argillite**—Pale, greenish-gray, massive argillite; minor amount laminated. Almost all argillite highly deformed and phyllitic. Appears to be two distinct argillite zones separated by quartzite zone about 50 m thick (included in unit Yba), but could be single argillite zone repeated by faulting. Upper part of upper argillite zone is dark-gray laminated argillite; strong deformation prevents determination of stratigraphic relation to pale-greenish-gray argillite in lower part of zone. Thickness of upper argillite zone uncertain, but could be as much as 180 m; thickness of lower argillite zone at least 50 m

Ybq **Quartzite**—Vitreous quartzite, subordinate argillite, and minor siltite and conglomerate. White, medium- to thick-bedded quartzite is most characteristic lithology of this quartzite unit. Lowest 10 m is sheared, pale-greenish-gray, phyllitic argillite that contains a single 1-m-thick quartzite bed. Argillite overlain by about 10 m of distinctive parallel-planar, micro-laminated, dark-gray and green argillite interbedded with parallel-planar, thin-bedded siltitic argillite. Even though predominently argillite, both 10-m-thick intervals included as part of lower quartzite zone. Thin-bedded rocks grade upward over a few meters into 300-m-thick sequence composed of (1) white, gray, and pink medium- to thick-bedded quartzite interbedded with thin-bedded siltite; (2) thin interbeds of pale-greenish-gray and pale-lavender-gray argillite with poorly defined lamination; and (3) interbedded maroon, green, and gray even-parallel and cross-bedded, partly feldspathic quartzite and siltite very similar to lithologies found in Bonner Quartzite of Belt Supergroup. Lithologies in 300-m-thick sequence alternate in no identifiable order, except that thick-bedded quartzite is most abundant in 50-m-thick zone in uppermost part and in 100-m-thick zone in lowermost part. Upper 50-m-thick quartzite zone separated from thicker lower zone by at least 50 m of argillite. Much thick-bedded quartzite is very coarse grained, some pebble bearing. In sec. 28, T. 32 N., R. 39 E., about 3 m of argillaceous, matrix-supported conglomerate forms basal part of 50-m-thick quartzite zone, and monolithologic dolomite conglomerate or breccia forms uppermost part. Neither conglomerate identified elsewhere in quadrangle

Ys **Stensgar Dolomite (Middle Proterozoic)**—Dolomite; contains minor interbedded argillite. Most of formation is white, tan, pink, or maroon dolomite. Maroon argillite beds in sequences up to 3 m thick restricted chiefly to sharply defined intervals in upper and middle parts of formation; lower part of formation contains thin beds and bedding-plane partings of argillite. Maroon coloration in formation diminishes southwestward. Best exposed and least deformed section is around Jim McGraff Quarry (fig. 1) in sec. 12, T. 31 N., R. 39 E.; but not known if it is representative of deformed Stensgar in rest of magnesite belt, because section occurs east of several large reverse and thrust faults. Lowest exposed beds at Jim McGraff Quarry are maroon argillite and siltite and one 10- to 20-cm-thick quartzite interbed. These maroon beds are overlain by about 50 m of alternating (1) white, gray, and pink, thin-bedded to thinly laminated dolomite that contains maroon argillite partings and thin even-parallel interbeds of chert and (2) massive-appearing, pink, tan, and white dolomite that weathers tan. Alternating sequence overlain by about 50 m of distinctive gray-weathering, medium- to thin-bedded dolomite and chert; no other carbonate rocks of Deer Trail Group are known to weather gray. Uppermost 100 to 150 m of Stensgar Dolomite in Jim McGraff Quarry section is thin- to thick-bedded, tan-weathering, white, gray, pink, and maroon dolomite; contains sparse algal structures, oolites, and distinct beds that contain nodular chert. About 50 to 75 m below inferred top of formation is 2 to 3 m of argillite containing abundant casts of one or more evaporite minerals. Mud-matrix-supported crystals that have octahedral forms suggest one mineral may have been northupite ($MgCO_3•Na_2CO_3•NaCl$; Wasson and others, 1984). May indicate primary origin for large magnesite deposits that are confined to Stensgar Dolomite in this region. Stensgar appears to grade into underlying McHale Slate over about 20 m, but contact zone is poorly exposed. Fault-bounded Stensgar sequences west of reverse and thrust faults have same lithologies as Jim McGraff Quarry sequence, but faulting, folding, and incomplete exposure make comparison of internal stratigraphy impossible

Ym **McHale Slate (Middle Proterozoic)**—Formation is almost entirely argillite. Lower third of formation is medium- to dark-gray argillite that has tan, pale-gray, and white laminae. Bedding thickness ranges from laminations on submillimeter scale to fining-upward couplets as much as 3 cm thick. Parallel-planar lamination common in this part of formation, although extensively disrupted by soft-sediment deformation. Wavy-parallel and wavy-nonparallel beds also abundant; some have prominent erosional bases. Gray argillite grades upward over a few meters into pale-greenish-gray and pale-lavender-gray argillite, which characterize upper two-thirds of formation. Green- and lavender-gray units alternate in intervals from 0.5 m to about 30 m in thickness. Pigmentation may reflect differences in oxidation state due to alteration, primary differences in mineralogy, or both; contacts sharp between differently colored zones. Bedding and lamination well developed throughout McHale, but indistinct in upper two-thirds of formation. Throughout extent, McHale is phyllitic and folded. Thickness as calculated from outcrop width is about 370 m

Ywd **Wabash Detroit Formation (Middle Proterozoic)—**Dolomite and subordinate argillite, quartzite, and carbonate-bearing quartzite and siltite; may contain greenstone in upper part. As defined by Miller (1996a), formation consists of relatively dolomite-rich rocks (approximately upper two-thirds) of now abandoned Edna Dolomite of Campbell and Loofbourow (1962). Formation is chiefly thin- to thick-bedded, tan-weathering, gray and white impure dolomite that has abundant thin interbeds of pale-green and gray argillite and carbonate-bearing siltite. Bedding is even parallel to wavy nonparallel; bed thickness averages between 5 and 15 cm. Chert-bearing dolomite beds present, but not abundant. Stromatolitic beds common; may have been more numerous, but most rocks are noticeably sheared. Argillitic and siltitic strata cleaved and sheared, and at many places all sedimentary structures, including bedding, destroyed. Zones of thick- to thin-bedded vitreous quartzite in uppermost part and in lower part of formation. Unit forms poor outcrops; no continuous exposure of entire formation exists. Grades upward over only a few meters from dolomite into argillite of McHale Slate; at some places, uppermost vitreous quartzite interval appears to directly underlie McHale, and at other places greenstone forms top of Wabash Detroit. Thickness as calculated from outcrop width about 240 m. Includes:

Ywdg **Greenstone—**Altered flows or sills of basaltic composition; matrix chloritized, and plagioclase altered to albite. Contains abundant calcite. Occurs persistently between Wabash Detroit Formation and McHale Slate, more so than indicated on geologic map, because greenstone was not recorded in early stages of mapping and because greenstone is thin and easily removed by even small faults. Persistence of greenstone at this interval suggests it is extrusive, but due to shearing no extrusive features are preserved. Greenstone could be southwestward extension of volcanism of this age represented by Nicol Creek Formation of McMechan and others (1980) in southern British Columbia. However, since no extrusive features were found, greenstone could actually represent sills related to greenstone member (Zhg) of Huckleberry Formation of Windermere Group. Thickness ranges from 0 to 15 m

Ywcu **Wabash Detroit Formation and Chamokane Creek Formation, undivided (Middle Proterozoic)—**Highly sheared and faulted dolomite, dolomitic quartzite, argillite, and quartzite. Mapped as undivided unit in Eagle Mountain area because bedding characteristics and other sedimentary features used to distinguish specific formations are destroyed, and because lithologies are homogenized by faulting

Chamokane Creek Formation (Middle Proterozoic)—Carbonate-bearing and noncarbonate-bearing quartzite and siltite interbedded with dolomite and argillite. As defined by Miller (1996a), consists of (1) predominantly carbonate-bearing clastic rocks of part (approximately lower third) of now-abandoned Edna Dolomite of Campbell and Loofbourow (1962), (2) vitreous quartzite and interbedded argillite that previously formed uppermost part of Togo Formation of Campbell and Loofbourow (1962), and (3) carbonate-bearing rocks that previously formed part of Togo Formation of Campbell and Loofbourow (1962), which lies directly below vitreous quartzite zone. Chamokane Creek Formation is much more poorly exposed than other units of Deer Trail Group. Location of formation commonly identified by red-orange soil developed on unit and by aligned swamps and bogs; vitreous quartzite beds are only resistant part of formation. Thickness of formation about 600 m as calculated from outcrop width of composite section in Empey Mountain map area. Consists of:

Ycc **Carbonate-bearing rocks**—Pale-tan, carbonate-bearing quartzite and siltite interbedded with lesser amounts of light-gray and pale-green argillite; dolomite. Unit essentially lithologically identical above and below approximately 150-m-thick zone of carbonate-free vitreous quartzite and argillite (Yccq) in upper middle part of Chamokane Creek Formation. Carbonate-bearing unit also contains zones of dark-gray argillite similar to argillite in underlying Togo Formation, especially just above and below Yccq. Unit also contains a few zones up to 3 m thick of tan impure dolomite, especially in lower part. Except for dolomite beds, bedding ranges in thickness from 1 to 10 cm and is chiefly wavy nonparallel. Pyrite ubiquitous but sparse, especially in dolomite. Carbonate minerals leached from most of carbonate-bearing quartzite and siltite; creates porous-looking rock, especially east of Empey Mountain and in North Fork of Chamokane Creek drainage (fig. 1). Ready erodibility of this rock accounts for paucity and poor quality of outcrops that characterize formation. Thickness of formation, including Yccq, about 600 m as calculated from outcrop width of composite section, but this figure poorly constrained. Includes:

Yccq **Vitreous quartzite and argillite**—Medium- to fine-grained, vitreous quartzite and interbedded dark-gray argillite; averages about 150 m thick. Upper and lower contacts are bed-by-bed gradation from carbonate-bearing quartzite and siltite (Ycc) to vitreous, noncarbonate-bearing quartzite over interval of 5 to 10 m; transition zone of lower contact contains numerous beds of dark-gray argillite lithologically similar to those typifying Togo Formation. Quartzite beds are quartz cemented and contain almost no other minerals. In northeast part of magnesite belt, quartzite beds separated by zones of dark-gray argillite ranging in thickness from less than 1 mm to 2 m. Argillite progressively less abundant southwestward. Bedding and lamination range from even parallel to wavy nonparallel. Some quartzite crossbedded; much of unit shows extensive soft-sediment deformation and compaction features. Syneresis cracks and fluid-escape structures abundant. Except for lack of carbonate minerals, this unit is nearly identical to rocks of Wallace Formation (Yw)

Yt **Togo Formation (Middle Proterozoic)**—Medium- and dark-gray argillite that contains subordinate beds and intervals of green argillite and green and gray siltite. Also contains rare, thin, sharply defined beds of quartzite and dolomite in lower part of formation. Bedding in dark-gray argillite ranges from wavy nonparallel to even parallel. Bedding thickness ranges from submillimeter laminae to about 10 cm; beds thicker than about 2 cm generally show even-parallel bedding. Laminations defined by white or light-gray siltite that weathers orange, possibly reflecting minor carbonate content. Most nonparallel beds have low to moderate angle of convergence between top and bottom; where angle is high, caused by soft-sediment deformation, which is common throughout unit. Almost all argillite in Togo is highly deformed and slightly to highly phyllitic. Within 0.5 to 1 km of Lane Mountain Fault or large reverse fault west of Lane Mountain Fault, nearly all Togo is phyllite that has little bedding or other sedimentary features preserved. Minimum thickness about 800 m, but internal deformation and lack of marker units preclude accurate estimate. Campbell and Loofbourow (1962) did not map lower part of Togo Formation, but estimated partial thickness of more than 1,220 m; Becraft and Weis (1963) estimated thickness of about 6,100 m; and Evans (1987) estimated a thickness of about 1,150 m. Current mapping suggests unusually broad outcrop is largely due to extreme internal deformation, but unit could be over 2,000 m thick. Base of formation everywhere faulted

Belt Supergroup—Forms two thick sequences of quartzite, siltite, argillite, dolomite, and mixtures of these four rock types in variable proportions. One sequence is north of Newport in hanging wall of Newport Fault, and other is east of Chewelah in footwall. Sequences also separated by part of (informal) Priest River complex. All Belt units in Newport sequence, except for Bonner Formation and part of Mount Shields Formation, are also found in Chewelah sequence. There are some differences in lithofacies of individual units between two sequences, and units are generally thicker in Chewelah sequence. Thickness of formations are calculated from outcrop width and are approximate, because all units are incompletely exposed and could contain hidden faults. Consists of:

Ybmh **Bonner Formation, Mount Shields Formation, and argillite of Half Moon Lake, undivided (Middle Proterozoic)**—Argillite, siltite, and lesser amounts of quartzite. Mapped as undivided unit only on ridge between Skookum and Browns Creeks (fig. 1), east of Pend Oreille River, where much of this rock unit is altered and bleached, but some sedimentary features are preserved. Unit poorly exposed and probably contains hidden faults. Thickness unknown

Ybo **Bonner Formation (Middle Proterozoic)**—Maroon, pale-purple, and pale-green siltite, argillite, and quartzite. Bed thickness ranges from less than 1 cm to about 1 m. Unit is about 80 percent thin-bedded, coarse siltite, 10 percent thick beds of quartzite, and 10 percent thin beds and bedding-plane films of argillite. Mud cracks, mud-chip breccia, and ripple marks common. Miller (1974a) erroneously reported salt casts from this unit. North of Newport, thickness of unit as calculated from outcrop width is about 190 m. Unconformably overlain by Late Proterozoic and Cambrian Addy Quartzite

Mount Shields Formation (Middle Proterozoic)—Argillite, siltite, dolomite, and dolomitic siltite. Members 1 and 2 not present in quadrangle. Consists of:

Yms_5 **Member 5**—Dark-gray, parallel-planar, micro-laminated argillite. Almost no variation in unit except for gradational zone in lower part containing minor thin dolomitic seams. Upper contact is abrupt bed-by-bed gradation over 2-m interval. Thickness between 70 and 80 m

Yms_4 **Member 4**—Dolomite, chert, and minor argillite. Gray, tan, and white; weathers tan, chocolate brown, and gray. Bed thickness ranges from 1 cm to 1 m; averages about 20 cm. Bedding is uneven to parallel planar. All rock surfaces in uppermost and lowermost parts of unit are hematite stained. Stromatolites fairly common. Gray or white chert occurs as nodular and irregular masses ranging from a few centimeters to 50 cm in length. In lower part of member, massive breccia composed of dolomite and chert forms bed several meters thick; breccia is sedimentary, not tectonic. Member about 160 m thick

Yms_3 **Member 3**—Siltite, argillite, dolomite, and quartzite. Siltite, quartzite, and most argillite is pale green and tan weathering; some argillite in lower part is laminated and dark gray. Bedding wavy parallel and nonparallel; graded couplets common, locally cross laminated. Detrital mica common throughout. Most surfaces are hematite stained. Fluid-escape structures, mud cracks, mud-chip breccia, ripple marks, and salt casts in most of unit. Dolomite beds range from a few centimeters to 1 m thick and in lower part of member are stromatolitic; in upper part, dolomite beds are thicker and more numerous. In sequence north of Newport, member about 180 m thick. On Quartzite Mountain east of Chewelah and in Grouse Creek area east of Jumpoff Joe Lake (fig. 1), a few meters of member 3 preserved beneath unconformity at base of Late Proterozoic and Cambrian Addy Quartzite

Yhm **Argillite of Half Moon Lake (Middle Proterozoic)**—Dark- to medium-gray, laminated argillite, thin-bedded siltite, and thick- to thin-bedded quartzite. Lower third of unit is dark-gray, parallel-planar-laminated argillite. Middle 75 m of unit is parallel-planar, thin-bedded, gray and greenish-gray siltite interlayered with laminated argillite similar to that in lower part of unit. North of Newport, several 3- to 50-cm-thick, fine-grained quartzite beds form middle part of this siltite interval. East of Chewelah, middle part of unit is much more quartzitic and contains detrital mica. Above siltite zone in both sections, unit is argillite similar to that in lower part. Argillite unit characterized by small localized areas of soft-sediment deformation confined in sharply bounded beds. Lithostratigraphic equivalent of laminated argillite and siltite member of Wallace Formation at Clark Fork, Idaho (Harrison and Jobin, 1963). Unit is about 350 m thick north of Newport, and thickens westward to about 650 m near Chewelah

Yssh **Shepard Formation and Snowslip Formation, undivided (Middle Proterozoic)**—Schist, calc-silicate rock, dolomitic marble, argillite, siltite, and dolomite. Mapped as undivided unit only west of Pend Oreille River at north edge of quadrangle. Rocks are contact metamorphosed by large Cretaceous pluton located a few hundred meters north of quadrangle; range from amphibolite to greenschist facies. Most sedimentary features in northern third of unit destroyed by metamorphism. All rocks in this area contain anomalous amounts of sulfide minerals, chiefly pyrite. Sequence may contain unrecognized faults. Thickness unknown

Yssw **Shepard Formation, Snowslip Formation, and Wallace Formation, undivided (Middle Proterozoic)**—Argillite, siltite, and porous quartzite. Very poorly exposed. All rocks hydrothermally altered and bleached; carbonate minerals leached from quartzite and siltite. Most rocks are pale tan to pale red-orange. Sedimentary features characteristic of the three formations poorly preserved in limited areas. Unit appears to be depositionally overlain by Late Proterozoic and Cambrian Addy Quartzite. Mapped only in area between Fan Lake and Loon Lake Mountain. Thickness unknown

Ysh **Shepard Formation (Middle Proterozoic)**—Dolomite, dolomitic siltite, and siltite. In sequence north of Newport, white, tan, and pale-gray, tan-weathering stromatolitic and oolitic dolomite; interbedded with dark, chlorite-green siltite beds in lower 60 m of formation. Siltite beds may be volcaniclastic (Miller and Whipple, 1989). Ten-cm- to 8 m-thick intervals of dolomite separate six or eight siltite intervals. Upper part of formation is pale-green, white, gray, and tan dolomite interbedded with pale-green dolomitic siltite. Shepard section southeast of Chewelah is interbedded dolomite and dolomitic siltite, but dominant colors are maroon and tan rather than pale green and tan, and chlorite-green siltite beds in lower part of unit in Newport sequence are lacking. Newport section is about 360 m thick; in Chewelah sequence, unit may be internally faulted, but appears to be about 430 m thick

Ywr **Wallace Formation and Ravalli Group, undivided (Middle Proterozoic)**—Quartz-feldspar-muscovite-biotite schist and calc-silicate rocks; medium to coarse grained. Most calc-silicate rocks restricted to northwestern part of unit between Calispell and Little Calispell Peaks (fig. 1), but numerous pods and inclusions up to 100 m in length found at least as far north as Calispell Peak. Most common assemblage is quartz-plagioclase-actinolite-diopside, but locally scapolite, vesuvianite, epidote, and clinozoisite are abundant. Nearly all schist is plagioclase-muscovite-biotite-quartz schist. Concentrations of muscovite suggest retrograded aluminum silicate minerals. No primary sedimentary features preserved; thickness unknown

Yss **Snowslip Formation (Middle Proterozoic)**—Medium- and dark-gray argillite and siltite and minor quartzite. Wavy, nonparallel to near-parallel argillite and siltite beds in fining-upward couplets interbedded with 2- to 50-cm-thick siltite intervals that have same bedding characteristics. North of Newport, upper two-thirds of formation has 1- to 20-m-thick intervals of gray-green siltite alternating with even-parallel-laminated, dark-gray argillite. Fluid-escape, syneresis, and soft-sediment-deformation structures common throughout unit. Formation is lithostratigraphic equivalent of combined argillite, siltite, and limestone member and argillite member of Wallace Formation at Clark Fork, Idaho (Harrison and Jobin, 1963). Snowslip of Newport and Chewelah sequences may also contain carbonate rocks in numerous covered intervals. In section north of Newport, thickness as calculated from outcrop width is about 380 m. East of Chewelah, if not repeated by faults, unit is about 1,400 m thick

Yw **Wallace Formation (Middle Proterozoic)**—Carbonate-bearing siltite and quartzite and abundant thin interbeds of dark-gray argillite. Thick beds of relatively pure dolomite that contain irregularly shaped fillings of sparry calcite are scattered throughout unit, but most abundant in lower part. Some dolomitic limestone beds in lower 100 m. Bedding mostly wavy nonparallel. Unit characterized by tan, carbonate-bearing quartzite and siltite beds that decrease in thickness from as much as 2 m to less than a few centimeters over lateral distance of 2 to 3 m. Tan beds separated by thin, black, phyllitic-looking argillite beds and bedding-plane partings that show the same radical thickness changes as the tan beds. Argillite cut by numerous syneresis cracks. Soft-sediment deformation, load features, and fluid-escape structures are also abundant. In section north of Newport, thickness about 730 m; east of Chewelah, about 800 m

Ye **Empire Formation (Middle Proterozoic)**—Pale-green siltite, quartzite, argillite, dolomite, and carbonate-bearing siltite. Mostly thin, uneven, wavy bedding, but some parallel-planar beds of quartzite as much as 1 m thick. Irregularly shaped 2- to 10-cm-long dolomite pods commonly weather out to form voids in pale-green argillite. Much of unit, particularly dolomitic parts, contains sparsely disseminated sulfide minerals. Lower 30 m contains thin beds of purple siltite similar to St. Regis Formation, except some is dolomite-bearing. Empire Formation shown as separate map unit north of Newport where it has average thickness as calculated from outcrop width of 320 m. As mapped herein, some strata of this unit is included as part of upper part of St. Regis Formation (Ysr) east and south of Chewelah where Empire is estimated to be about 100 m thick

Ysr **St. Regis Formation (Middle Proterozoic)**—Maroon to purple siltite, argillite, and lesser amounts of quartzite. Bedding characteristically wavy nonparallel, but some almost parallel planar. Bed thickness averages about 6 cm, but much argillite forms thin laminations and some siltite and quartzite beds are as much as 0.5 m thick. Ripple marks, mud cracks, mud-chip breccia, cross lamination, and fluid-escape structures common throughout formation. As mapped herein, St. Regis south and east of Chewelah includes pale-green beds of Empire Formation, but section north of Newport does not. Average thickness of St. Regis north of Newport is 275 m; St. Regis of Chewelah sequence averages 450 m as mapped with included pale-green Empire strata and 350 m without them

Yr **Revett Formation (Middle Proterozoic)**—Quartzite and minor siltite; white, tan, light-gray, pink, and maroon. Average bed thickness 0.5 m; ranges from a few centimeters to about 2 m. Large- and small-scale crossbedding common, but owing to metamorphism and poor exposures not obvious everywhere; ripple marks common, especially in upper part of unit. Most of formation is uniform white, tan, or gray fine-grained quartzite. North of Newport, upper 200 m of formation is pink or maroon quartzite; color similar to St. Regis Formation, but lithology typical of Revett. East and south of Chewelah, upper 100 m of formation is pale lavender and 200 m near middle part is mostly siltite. Formation north of Newport averages 750 m thick and, east and south of Chewelah, 950 m thick

Ybk **Burke Formation (Middle Proterozoic)**—Siltite with minor amounts of argillite and quartzite. Most of formation extremely uniform medium- to pale-gray siltite; characteristically has weathering rind that is distinctly lighter gray. Quartzite beds sparsely scattered throughout formation but more abundant in upper 300 m. Bed thickness of siltite averages 20 cm; ranges from a few centimeters to 3 m. Bedding mostly parallel planar to slightly wavy parallel. Oscillation and current ripple marks common, especially in upper part; much siltite is finely cross laminated. Upper part of formation contains about 100 m of maroon siltite, argillite, and quartzite virtually identical to St. Regis Formation; has all bedding features and sedimentary structures that characterize latter unit. In section north of Newport, formation averages about 850 m thick and, east and south of Chewelah, about 1,100 m

Yd **Mafic sills (Middle Proterozoic)**—Medium- to fine-grained sills of diabase composition that intrude Prichard Formation. All are recrystallized; composed of hornblende, biotite, plagioclase, quartz, and opaque minerals. Hornblende forms over 50 percent of most rocks. Sills range in thickness from about 1 m to 460 m. Thicker sills have noticeably coarser grained, relatively leucocratic zones in upper middle parts. Concentration and thickness of sills variable over distances of a few to tens of kilometers. North of Newport, high concentration of thick sills thins northward. East of Chewelah, sills are sparser and thinner than in Newport sequence. Some intrusions discordant, especially in lower part of Prichard Formation. Zircon from sill near Bonners Ferry, Idaho, gives uranium-lead age of 1,433 Ma (Zartman and others, 1982)

Yp **Prichard Formation (Middle Proterozoic)**—Interbedded quartzite, siltite, and argillite; colors range from white and pale gray for quartzite, pale to medium gray for siltite, and medium to dark gray for argillite. Entire formation contains pyrite, but highest concentration is in argillites; oxidation of pyrite causes almost all rock surfaces in Prichard to be iron-oxide stained. Much quartzite relatively pure; complete gradation from siltite to argillite. Most bedding and lamination in Newport and Chewelah sequences is parallel planar, and graded bedding on all scales characterizes rocks. Upper 500 to 1,500 m of formation mainly dark-gray, parallel-planar-laminated argillite and some interbedded quartzite in uppermost part. Below that, quartzite, siltite, and argillite vary in relative proportions, but internal stratigraphy not studied in enough detail to quantify or subdivide formation. Cressman (1989) assigned most of the Prichard north of Newport to what he termed its quartzite member, in which quartzite forms about 65 percent of section and individual quartzite beds are turbidites. He assigned middle 1,300 m of Prichard east of Chewelah to his quartzite member; upper 3,000 m to his upper laminated argillite member; and all rock below quartzite member to his member F, characterized by thin-bedded and laminated siltite and argillite. Prichard in Newport sequence is about 5,200 m thick as calculated from outcrop width and in Chewelah sequence about 4,100 m thick, but undetected faults are probably present in both sequences. Base of formation nowhere exposed in quadrangle. In Chewelah sequence, metamorphism of Prichard Formation progressively increases eastward. Includes:

Ypm **Metamorphosed part (Middle Proterozoic)**—Medium- to coarse-grained schist and hornfels intruded by two-mica granitic rocks; proportion of igneous rocks in unit generally increases eastward. Contact with main part of Prichard Formation is gradational zone several hundred meters wide that is generally placed where bedding in Prichard cannot be distinguished from metamorphic foliation. Metamorphic facies range from middle to upper greenschist to upper amphibolite; coarseness of recrystallized grains increases progressively eastward, but increase in metamorphic grade is not constant and may peak several km west of Newport Fault. Most of unit Ypm within 4 km of Newport Fault is mylonitized

Ynl **Newman Lake Gneiss (Proterozoic?)**—Biotite-quartz-plagioclase-potassium feldspar gneiss. Contains traces of muscovite; opaque minerals, allanite, zircon, and apatite present as accessories. Foliation and lineation in most of unit caused by intense ductile deformation, which decreases westward. Numerous pods of highly recrystallized rocks of Prichard Formation (**Yp**) and mafic sill rocks (**Yd**) found enclosed within easternmost 300 to 500 m of unit

sgg **Schist, gneiss, and leucocratic granitic rocks (age unknown)**—Coarse-grained quartz-feldspar-muscovite-biotite schist and gneissic rocks that include amphibolite bands and pods. Intruded by concordant and discordant, texturally and compositionally heterogeneous leucocratic granitic rocks. Schist and gneiss could be either metamorphosed rocks of Belt Supergroup, or could be pre-Belt crystalline rocks

REFERENCES CITED

Archibald, D.A., Krogh, T.E., Armstrong, R.L., and Farrar, Edward, 1984, Geochronology and tectonic implications of magmatism and metamorphism, southern Kootenay arc and neighbouring regions, southeastern British Columbia. Part II: mid-Cretaceous to Eocene: Canadian Journal of Earth Science, v. 21, p. 567-583.

Armstrong, R.L., Parrish, R.R., van der Heyden, Peter, Reynolds, S.J., and Rehrig, W.A., 1987, Rb-Sr and U-Pb geochronology of the Priest River complex—Precambrian X basement and its Mesozoic-Cenozoic plutonic-metamorphic overprint, northeastern Washington and northern Idaho, *in* Schuster, J.E., ed., Selected papers on the geology of Washington: Washington Division of Geology and Earth Resources Bulletin 77, p. 15-40.

Becraft, G.E., and Weis, P.L., 1963, Geology and mineral deposits of the Turtle Lake quadrangle, Washington: U.S. Geological Survey Bulletin 1131, 73 p., scale 1:62,500.

Burmester, R.F., and Miller, F.K., 1983, Preliminary geologic map of the Abercrombie Mountain area: U.S. Geological Survey Open-File Report 83-600, 10 p., scale 1:48,000.

Campbell, A.B., and Raup, O.B., 1964, Preliminary geologic map of the Hunters quadrangle, Stevens and Ferry Counties, Washington: U.S. Geological Survey Mineral Investigations Field Studies Map MF-276, scale 1:48,000.

Campbell, Ian, and Loofbourow, J.S., 1962, Geology of the magnesite belt of Stevens County, Washington: U.S. Geological Survey Bulletin 1142-F, p. F1-F53, scale 1:36,000.

Carter, Claire, 1989a, A Middle Ordovician graptolite fauna from near the contact between the Ledbetter Slate and the Metaline Limestone in the Pend Oreille Mine, northeastern Washington state: U.S. Geological Survey Bulletin 1860-A, p. A1-A23.

Carter, Claire, 1989b, Ordovician-Silurian graptolites from the Ledbetter Slate, northeastern Washington state: U.S. Geological Survey Bulletin 1860-B, p. B1-B29.

Clark, L.D., and Miller, F.K., 1968, Geology of the Chewelah Mountain quadrangle, Stevens County, Washington: Washington Division of Mines and Geology Geologic Map GM-5, 6 p., 2 sheets, scale 1:62,500.

Cressman, E.R., 1989, Reconnaissance stratigraphy of the Prichard Formation (Middle Proterozoic) and the early development of the Belt basin, Washington, Idaho, and Montana: U.S. Geological Survey Professional Paper 1490, 80 p.

Devlin, W.J., and Bond, G.C., 1988, The initiation of the early Paleozoic Cordilleran miogeocline—Evidence from the uppermost Proterozoic-Lower Cambrian Hamill Group of southeastern British Columbia: Canadian Journal of Earth Sciences, v. 25, no. 1, p. 1-19.

Devlin, W.J., Bond G.C., and Brueckner, H.K., 1985, An assessment of age and tectonic setting of volcanics near the base of the Windermere Supergroup in northeastern Washington—implications for latest Proterozoic-earliest Cambrian continental separation: Canadian Journal of Earth Sciences, v. 22, p. 829-837.

Ellis, M.A., 1986, Structural morphology and associated strain in the central Cordillera (British Columbia and Washington)—Evidence of oblique tectonics: Geology, v. 14, p. 647-650.

Evans, J.G., 1987, Geology of the Stensgar Mountain quadrangle, Stevens County, Washington: U.S. Geological Survey Bulletin 1679, 23 p., scale 1:24,000.

Evans, J.G., 1988, Deformation in the Stensgar Mountain quadrangle, Stevens County, Washington: U.S. Geological Survey Bulletin 1820, 30 p.

Gager, B.R., 1982, Stratigraphy, sedimentology, and tectonic implications of the Tiger Formation, Pend Oreille and Stevens Counties, Washington: Seattle, University of Washington, M.S. thesis, 176 p.

Gager, B.R., 1984, The Tiger Formation: Paleogene fluvial sediments deposited adjacent to a deforming Cordilleran metamorphic core-complex, northeastern Washington: Sedimentary Geology, v. 38, p. 393-420.

Groffman, L.H., 1986, Stratigraphy, sedimentation, and structure of the upper Proterozoic Three Sisters Formation and lower Cambrian Gypsy Quartzite, northeast Washington: Pullman, Washington State University, M.S. thesis, 208 p., scale 1:12,000.

Harms, T.A., and Price, R.A., 1992, The Newport Fault: Eocene listric normal faulting, mylonitization, and crustal extension in northeast Washington and northwest Idaho: Geological Society of America Bulletin, v. 104, p. 745-761.

Harrison, J.A., and Jobin, D.A., 1963, Geology of the Clark Fork quadrangle, Idaho-Montana: U.S. Geological Survey Bulletin 1141-K, 38 p.

Hogge, C.E., 1982, Geology of the Paleozoic black shale sequence, Moberg Hill, Stevens County, Washington: Pullman, Washington State University, M.S. thesis, 80 p.

Hosterman, J.W., 1969, Clay deposits of Spokane County, Washington: U.S. Geological Survey Bulletin 1270, 96 p.

Joseph, N.L., 1990, Geologic map of the Colville 1:100,000 quadrangle, Washington-Idaho: Washington Division of Geology and Earth Resources Open-File Report 90-13, 78 p., scale 1:100,000.

Lindsey, K.A., Gaylord, D.R., and Groffman, L.H., 1990, Geology of the Upper Proterozoic to Lower Cambrian Three Sisters Formation, Gypsy Quartzite, and Addy Quartzite, Stevens and Pend Oreille Counties, northeastern Washington: Washington Division of Geology and Earth Resources, Report of Investigations 30, 37 p.

Lis, M.G., and Price, R.A., 1976, Large-scale block faulting during deposition of the Windermere Supergroup (Hadrynian) in southeastern British Columbia: in Report of activities, part A, Geological Survey of Canada, Paper 76-1A, p. 135-136.

Lucas, H.E., 1980, Geology of part of the Dunn Mountain and Arden 7-1/2 minute quadrangles, Stevens County, Washington: Cheney, Eastern Washington University, M.S. thesis, 36 p., scale 1:12,000.

Ludwig, K.R., Nash, J.T., and Naeser, C.W., 1981, U-Pb isotope systematics and age of uranium mineralization, Midnight mine, Washington: Economic Geology, v. 25, no. 4, p. 514-519.

McCarthy, P.T., Miller, D.M., Nakata, J.K., and Hoisch, T.D., 1993, Structures adjacent to the Purcell trench south of Sandpoint, Idaho: Geological Society of America Abstracts with Programs, vol. 25, no. 5, p. 117.

McMechan, M.E., Höy, Trygve., Price, R.A., 1980, Van Creek and Nicol Creek Formations (new); A revision of the stratigraphic nomenclature of the middle Proterozoic Purcell Supergroup, southeastern British Columbia: Bulletin of Canadian Petroleum Geology, v. 28, no. 4, p. 542-558.

Miller, F.K., 1971, The Newport fault and associated mylonite, northeastern Washington: U.S. Geological Survey Prof. Paper 750-D, p. D77-D79.

Miller, F.K., 1974a, Preliminary geologic map of the Newport number 1 quadrangle, Pend Oreille County, Washington, and Bonner County, Idaho: Washington Division of Geology and Earth Resources Geologic Map GM-7, 6 p., 1 pl., scale 1:62,500.

Miller, F.K., 1974b, Preliminary geologic map of the Newport number 2 quadrangle, Pend Oreille and Stevens Counties, Washington: Washington Division of Geology and Earth Resources Geologic Map GM-8, 6 p., 1 pl., scale 1:62,500.

Miller, F.K., 1974c, Preliminary geologic map of the Newport number 3 quadrangle, Pend Oreille, Stevens, and Spokane Counties, Washington: Washington Division of Geology and Earth Resources Geologic Map GM-9, 7 p., 1 pl., scale 1:62,500.

Miller, F.K., 1983, Preliminary geologic map of the Gypsy Peak area: U.S. Geological Survey Open-File Report 83-601, 28 p., scale 1:48,000.

Miller, F.K., 1994, The Windermere Group and Late Proterozoic tectonics of northeastern Washington and northern Idaho: *in* Lasmanis, Raymond and Cheney, E.S., eds., Selected papers on the geology of Washington: Washington Division of Geology and Earth Resources Bulletin 80, p. 1-19.

Miller, F.K., 1996a, Geologic map of the Empey Mountain area, Stevens County, Washington: U.S. Geological Survey Miscellaneous Investigation Series Map I-2493, scale 1:62,500.

Miller, F.K., 1996b, Geologic map of the Addy area, Stevens County, Washington: U.S. Geological Survey Miscellaneous Investigation Series Map I-2492, scale 1:62,500.

Miller, F.K., Burmester, R.F., Miller, D.M., and Powell, R.E.; *digital compilation by* Pamela D. Derkey, 1999, Digital geologic map of the Sandpoint 1° X 2° quadrangle, Washington, Idaho, and Montana: U.S. Geological Survey Open-File Report 99-144, scale 1:250,000.

Miller, F.K., and Clark, L.C., 1975, Geology of the Chewelah-Loon Lake area, Stevens and Spokane Counties, Washington: U.S. Geological Survey Professional Paper 806, 74 p., scale 1:62,500.

Miller, F.K., and Engels, J.C., 1975, Distribution and trends of discordant ages of the plutonic rocks of northeastern Washington and northern Idaho: Geological Society of America Bulletin, v. 86, no. 4, p. 517-528.

Miller, F.K., McKee, E.H., and Yates, R.G., 1973, Age and correlation of the Windermere Group in northeastern Washington: Geological Society of America Bulletin, v. 84, p. 3,723-3,730.

Miller, F.K., and Whipple, J.W., 1989, The Deer Trail Group—Is it part of the Belt Supergroup?, *in* Joseph, N.L., and others, eds., Geologic guidebook for Washington and adjacent areas: Washington Division of Geology and Earth Resources Information Circular 86, p. 1-21.

Okulitch, V.J., 1951, A Lower Cambrian fossil locality near Addy, Washington: Journal of Paleontology, v. 25, no. 3, p. 405-407.

Park, C.F., Jr., and Cannon, R.S., Jr., 1943, Geology and ore deposits of the Metaline quadrangle, Washington: U.S. Geological Survey Professional Paper 202, 81 p, scale 1:125,000.

Pearson, R.C., and Obradovich, J.D., 1977, Eocene rocks in northeastern Washington—Radiometric ages and correlation: U.S. Geological Survey Bulletin 1433, 41 p.

Reesor, J.E., 1957, Lardeau (east half) map area, British Columbia: Geological Survey of Canada Map 12-1957.

Rehrig, W.A., Reynolds, S.J., and Armstrong, R.L., 1987, A tectonic and geochronologic overview of the Priest River crystalline complex, northeastern Washington and northern Idaho, *in* Schuster, J.E., ed., Selected papers on the geology of Washington: Washington Division of Geology and Earth Resources Bulletin 77, p. 1-14.

Rhodes, B.P., and Hyndman, D.W., 1984, Kinematics of mylonites in the Priest River "metamorphic core complex," northern Idaho and northeastern Washington: Canadian Journal of Earth Sciences, v. 21, no. 10, p. 1161-1170.

Rice, H.M.A., 1941, Nelson map-area, east half, British Columbia: Canadian Geological Survey Memoir 228, Pub. 2,460, 86 p.

Roback, R.C., and Walker, N.W., 1989, Stratigraphic relationships in late Paleozoic to middle Mesozoic rocks of Quesnellia, northeast Washington and southeast British Columbia: Geological Society of America Abstracts with Programs, vol. 21, no. 5, p. 134.

Russell, L.C., 1989, Geochemistry of plutonic rocks of the Loon Lake batholith and associated volcanic and hypabyssal rocks: Geological Society of America Abstracts with Programs, vol. 21, no. 5, p. 137.

Russell, L.C., 1993, Geochemistry of Tertiary igneous rocks in the eastern half of the Chewelah quadrangle, Pend Oreille County, Washington: Cheney, Eastern Washington University, M.S. thesis, 323 p.

Schroeder, M.C., 1952, Geology of the Bead Lake district, Pend Oreille County, Washington: Washington Division of Mines and Geology Bulletin 40, 57 p.

Schuster, J.E., 1976, Geology of the Clugston Creek area, Stevens County, Washington: Washington Division of Geology and Earth Resources Open-File Report 76-8, 26 p.

Steiger, R.H., and Jaeger, E., 1977, Subcommission on geochronology—Convention on the use of decay constants in geo- and cosmo-chronology: Earth and Planetary Science Letters, v. 36, no. 3, p. 359-362.

Streckeisen, A.L., 1973, Plutonic rocks—Classification and nomenclature recommended by the IUGA Subcommission on Systematics of Igneous Rocks: Geotimes, vol. 18, p. 26-30.

Swanson, D.A., Wright, T.L., Hooper, P.R., and Bentley, R.D., 1979, Revisions in stratigraphic nomenclature of the Columbia River Basalt Group: U.S. Geological Survey Bulletin 1457G, 59 p.

Waggoner, S.Z., 1990, Geologic map of the Chewelah 1:100,000 quadrangle, Washington-Idaho: Washington Division of Geology and Earth Resources Open-File Report 90-14, 63 p.

Wasson, R.J., Smith, G.I., and Agrawal, D.P., 1984, Late Quaternary sediments, minerals, and inferred geochemical history of Didwana Lake, Thar Desert, India: Palaeogeography, Palaeoclimatology, Paleaoecology, v. 46, p. 345-372.

Watkinson, A.J., and Ellis, M.A., 1987, Recent structural analysis of the Kootenay arc in northeastern Washington, *in* Schuster, J.E., ed., Selected papers on the geology of Washington: Washington Division of Geology and Earth Resources Bulletin 77, p. 41-53.

Whitehouse, M.J., Stacey, J.S., and Miller, F.K., 1992, Age and nature of the basement in northeastern Washington and northern Idaho—Isotopic evidence from Mesozoic and Cenozoic granitoids: Journal of Geology, v. 100, p. 691-701.

Yates. R.G., and Engels, J.C., 1968, Potassium-argon ages of some igneous rocks in northern Stevens County, Washington, in Geological Survey research 1968: U.S. Geological Survey Professional Paper 600-D, p. D242-D247.

Zartman, R.E., Peterman, Z.E., Obradovich, J.D., Gallego, M.D., and Bishop, D.T., 1982, Age of the Crossport C sill near Eastport, Idaho, *in* Reed, R.R., and others, eds., Society of Economic Geologists, Coeur d'Alene Field Conference: Idaho Bureau of Mines and Geology Bulletin 24, p. 61-69.