N A T U R A L R E S O U R C E S

Regional Geology of Washington State

Raymond Lasmanis
and Eric S. Cheney,
Convenors

WASHINGTON
DIVISION OF GEOLOGY
AND EARTH RESOURCES

Bulletin 80
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WASHINGTON STATE DEPARTMENT OF
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Raymond Lasmanis—State Geologist
J. Eric Schuster—Assistant State Geologist

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Preface

The first symposium on the Regional Geology of Washington was held at the annual meeting of the Cordilleran Section of the Geological Society of America in April 1982, and a volume containing most of the proceedings was eventually published (Schuster, 1987). Bates McKee and Eric Cheney were the convenors of that symposium. The history of the symposium and volume are discussed in Schuster (1987).

Since 1982, so much additional progress has been made in the understanding of the regional geology of Washington that we thought the time ripe for a second symposium. This we convened in May 1992 at Eugene, Oregon, again in conjunction with the annual meeting of the Cordilleran Section of the Geological Society of America. The symposium was enormously successful; probably more than 400 people attended parts of it. Table 1 lists the authors and titles of their papers. A comparison of this table with the 1982 program (Schuster, 1987, p. viii-ix) reveals that only about 20 percent of the authors were repeat performers. Those of us with more than a few gray hairs look to the under-40 crowd in the other 80 percent to promote a third symposium in 2002.

Table 1. Symposium program, May 12, 1992. Asterisks indicate papers represented only by abstracts at the back of this book

The Second Symposium on the Regional Geology of the State of Washington, Part I

Raymond Lasmanis and Timothy J. Walsh, Presiding

Introduction	8:00 A
Darrel S. Cowan: Cordilleran tectonic setting of Washington*	8:10 A
Fred K. Miller: The Windermere Group and Late Proterozoic tectonics in northeastern Washington	8:30 A
Kenneth F. Fox, Jr.: Metamorphic core complexes within an Eocene extensional province in north-central Washington	8:50 A
Stephen E. Box: Detachment origin for Republic graben, northeastern Washington*	9:10 A
E. S. Cheney, M. G. Rasmussen, M. G Miller: Lithologies, structure, and stacking order of Quesnellia in north-central Washington	9:30 A
V. R. Todd, S. E. Shaw, H. A. Hurlow, R. J. Fleck: The Okanogan range batholith, north-central Washington—Root of a Late Jurassic(?)–Early Cretaceous continental-margin arc*	9:50 A
J. I. Garver, M. F. McGroder, D. Mohrig, J. Bourgeois: The stratigraphic record of mid-Cretaceous orogeny in the Methow basin, Washington and British Columbia*	10:25 A
Robert B. Miller, R. A. Haugerud, D. L. Whitney, S. A. Bowring, T. B. Housh: Tectonics of the NE margin of the Cascade crystalline core, Washington	10:45 A
E. H. Brown: Baric patterns in the Cascades crystalline core, Washington	11:05 A
Ralph A. Haugerud, Rowland W. Tabor, Charles D. Blome: Pre-Tertiary stratigraphy and multiple orogeny in the western North Cascades, Washington*	11:25 A
Wilbert R. Danner: Stratigraphy and biostratigraphy of the Paleozoic and Early Mesozoic rocks of San Juan Islands and northwestern Cascade Mountains, Washington*	11:45 A

The Second Symposium on the Regional Geology of the State of Washington, Part II

Park D. Snavely, Jr., and Brian S. Butler, Presiding

E. S. Cheney, R. J. Stewart: The Cenozoic interregional unconformity-bounded sequences of central and eastern Washington	1:00 P
Joseph A. Vance, Robert B. Miller: Another look at the Fraser River–Straight Creek fault (FRSCF)*	1:20 P
Samuel Y. Johnson, J. C. Yount: Toward a better understanding of the Paleogene paleogeography of the Puget Lowland, western Washington*	1:40 P
Donald A. Swanson, Russell C. Evarts: Tertiary magmatism and tectonism in an E–W transect across the Cascade arc in southern Washington*	2:00 P
R. S. Babcock, C. A. Suczek, D. C. Engebretson: Geology of the Crescent terrane, Olympic Peninsula, WA	2:20 P
Mark T. Brandon, Joseph A. Vance: Age and tectonic evolution of the Olympic subduction complex as inferred from fission-track ages for detrital zircons*	2:40 P
S. P. Reidel, K. R. Fecht, K. A. Lindsey, N. P. Campbell: Post-Columbia River Basalt structure and stratigraphy of south-central Washington	3:20 P
Stephen C. Porter: Alpine glaciation of western Washington*	3:40 P
Alan J. Busacca, Eric V. McDonald: Regional sedimentation of Late Quaternary loess on the Columbia Plateau—Source areas and wind distribution patterns	4:00 P
D. J. Easterbrook, G. W. Berger, R. Walter: Laser argon and TL dating of early and middle Pleistocene glaciations in the Puget Lowland, Washington	4:20 P
Derek B. Booth, Barry Goldstein: Patterns and processes of landscape development by the Puget Lobe ice sheet	4:40 P
Brian F. Atwater: Prehistoric earthquakes in Western Washington	5:00 P

Table 1 also illustrates that, as in the case of the first symposium, not all papers were submitted for publication. We thank the Geological Society of America for allowing abstracts of the unpublished papers to be republished here. We thank authors and reviewers for their expeditious efforts. We are grateful to the following, who provided additional reviews: Joe D. Dragovich; Kenneth F. Fox, Jr.; Nancy L. Joseph; William S. Lingley, Jr.; J. Eric Schuster; Patrick Spencer; and Timothy J. Walsh. Members of the publications staff of the Division of Geology and Earth Resources contributed to this volume as well: Nancy A. Eberle and Keith G. Ikerd assisted with illustrations, Jari Roloff designed and prepared layout, and Katherine Reed served as volume editor.

We believe that this volume will prove useful to a broad cross section of geological scientists working in the Pacific Northwest. For those who seek an overview of the geology of the state in a limited number of volumes, we recommend Schuster (1987), Joseph (1989), Galster (1989), and Lasmanis (1991, including other articles in that issue), and this volume. Geological maps that are particularly useful are the 1:250,000-scale maps compiled by the Division of Geology and Earth Resources: Walsh and others (1987), Stoffel and others (1991), and the southeast quadrant being compiled by J. E. Schuster and others. This series will be complete when the northwest quadrant map is published in about 1995; several articles in this volume discuss the regional geology of the northwest part of the state. *Cascadia* (McKee, 1972) can still be recommended for its lucid style, good descriptions, and fine photographs, but it was published before the plate tectonics era.

R.L. E.S.C.

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Contents

This 2-m-diameter Douglas-fir stump, discovered during excavations in Orting, Pierce County, was part of an old-growth forest inundated by the Electron Mudflow about 600 years ago. The mudflow began when more than 200 million m^3 of debris was dislodged from upper west flanks of Mount Rainier about 50 km away (channel distance), possibly by a steam explosion or regional earthquake. View is to the south-southeast. Photo by Patrick Pringle, 1993.

The Windermere Group and Late Proterozoic Tectonics in Northeastern Washington and Northern Idaho

Fred K. Miller
U.S. Geological Survey
W. 904 Riverside
Spokane, WA 99201-1087

ABSTRACT

The Late Proterozoic Windermere Group in northeastern Washington and northern Idaho is a sequence of varied thickness, consisting of conglomerate, diamictite, greenstone, quartzite, grit, argillite, limestone, and dolomite. Volumetrically, most Windermere rocks are sedimentologically immature, and at least part of the sequence is considered to be glacial or glacial-marine. In northeastern Washington, large lithostratigraphic variations within Windermere sequences in adjacent fault blocks are interpreted to be the result of syndepositional block faulting associated with initial continental-scale rifting and development of the Windermere depositional wedge. Strong Paleozoic(?) and Mesozoic tectonic overprints mask the Late Proterozoic structures, but they cannot fully account for differences in lithostratigraphy between adjacent fault blocks. The Late Proterozoic faults are both parallel and oblique to the northeast regional strike of the southern Kootenay arc; the regional strike may or may not parallel the Windermere depoaxis. These syndepositional faults appear to have caused numerous contrasting, but areally restricted, depositional environments. Thickness of a formation in adjacent fault blocks varies as much as 3,000 meters; differences in lithostratigraphy may accompany differences in thickness.

The Jumpoff Joe fault can be traced 130 kilometers from about latitude 48°N to latitude 49°N, at which point it is roughly aligned with the St. Mary fault 55 kilometers to the northeast in southern British Columbia. Both faults separate sequences that contain thick Windermere sections from sequences that do not. The faults may have formed a continuous fault system that was a major bounding structure for the Windermere depositional wedge.

INTRODUCTION

In northeastern Washington and northern Idaho (Fig. 1), some formations of the Late Proterozoic Windermere Group show extreme differences in thickness and lithofacies both parallel to and across the regional structural strike. These differences appear to have been caused by rift-related syndepositional faults within, and not adjacent to, the developing Windermere basin because they separate highly dissimilar, but nearby, sections of the same formation. The Jumpoff Joe fault (Miller and Clark, 1975) is an exception, and it may represent a major basin-bounding normal fault that was reactivated with reverse movement in post-Windermere time. Similar pre- or syndepositional faulting has been documented or proposed in Canada (Lis and Price, 1976; Eisbacher, 1981, 1985; Devlin and Bond, 1988; Jefferson and Parrish, 1989), where Windermere rocks are discontinuously found at least to latitude 65°N. Unlike the relatively widespread distribution of Windermere rocks in the Canadian Cordillera, all of the Windermere in the United States is restricted to a fairly narrow belt within the northeast-striking southern part of the Kootenay arc. The Kootenay arc is a 400-km-long, broad bend in the regional stratigraphic and structural framework of the Cordillera north and south of the U.S.–Canada boundary. All of the area west of the Jumpoff Joe fault in Figure 2 (and Fig. 13) is part of the arc.

Evidence for Late Proterozoic tectonism in the southern Kootenay arc is masked by a strong, pervasive Paleozoic(?) and Mesozoic compressional (and probably extensional [Miller and others, 1992]) tectonic overprint and, locally, a strong Eocene extensional overprint. Largely because of the magnitude and extent of these younger structural features in Washington and Idaho, it is difficult to recognize or document Windermere-age tectonic events such as the dual pulses of Windermere crustal thinning described by Pell and Simony (1987) 300 km to the north in south-central British Columbia. Despite these younger tectonic events, a number of geologic relations in the region strongly suggest that several faults cutting Windermere rocks are Late Proterozoic in age. These relations, chiefly extreme differences in formation thickness across some faults and extreme differences in lithologic character across others, suggest that rift-related fault control, similar to that found in Canada, extended southward into the U.S.

WINDERMERE GROUP

In Washington and northern Idaho, the Windermere Group is best exposed in two areas that are separated by about 40 km of Cretaceous plutonic rocks (Fig. 2). The southern area, in the southernmost part of the Kootenay arc, is herein referred to informally as the magnesite belt area. The northern area, in northeasternmost Washington and northern Idaho, is herein referred to informally as the Salmo–Priest

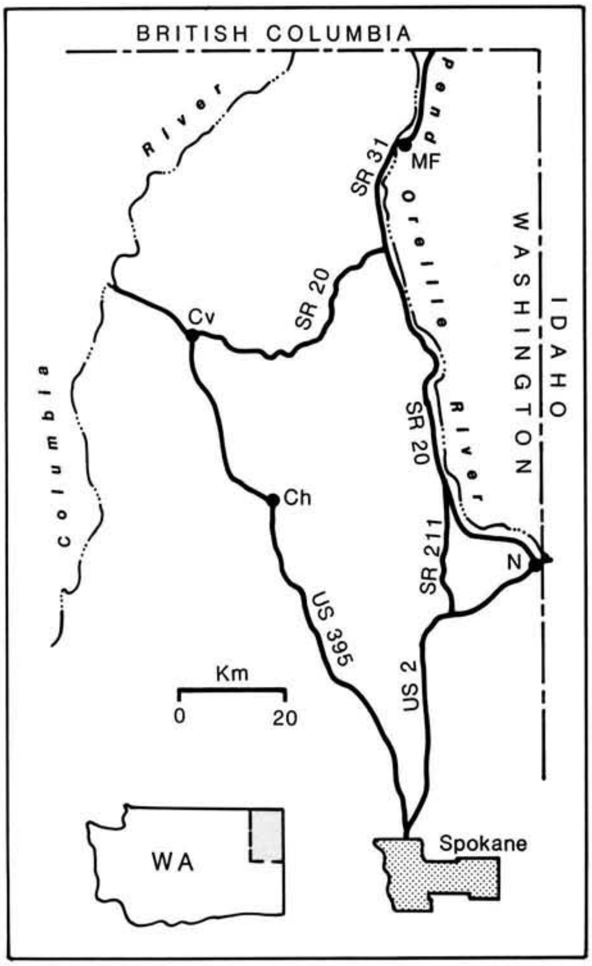

Figure 1. Index map and map showing major rivers, roads, and towns in the area of Figure 2. Cv, Colville; MF, Metaline Falls; N, Newport; Ch, Chewelah; SR, Washington State Route.

area. Within each area, the lithostratigraphy of the Windermere formations is nearly as varied as it is between these two areas; in many places, highly dissimilar Windermere sequences are separated by only a few kilometers. At least some of this variation is here interpreted as a result of Late Proterozoic syndepositional faulting. To illustrate the degree of variation, stratigraphic sections of each formation for several geographically separated sequences in each area are described later in this article.

Although early work in the region (Bennett, 1941; Campbell and Loofbourow, 1962; Park and Cannon, 1943) resulted in different names for lithologically similar units, rocks included in the Windermere Group in both areas essentially are made up of (in ascending order): (1) a lower unit that is mostly conglomerate, (2) a metavolcanic unit, and (3) a fine-grained, but highly varied clastic and carbonate unit. A fourth unit, coarse-grained clastic rocks[1], is found only in the Salmo–Priest area in the U.S. In southernmost Canada, the lower two units have different names than their U.S. counterparts, but they are in essence the same lithostratigraphic unit. Farther north, however, thickness and facies changes in Windermere rocks necessitate introduction of new formation names (Fig. 3).

As a whole, the Windermere Group is sedimentologically immature and constitutes a thick sequence of conglomerate, diamictite, greenstone (metavolcanic rocks), quartzite, grit, argillite, limestone, and dolomite. Some of the quartzite, argillite, and carbonate rocks, chiefly in the upper two units, are sedimentologically mature, but most of the quartzite is feldspathic or lithic, and much of the argillite and carbonate rocks contains millimeter-size quartz grains. With the exception of granitic boulders in the lower (conglomerate) unit and andesitic volcanic cobbles in one sequence of the fine-grained clastic and carbonate rocks unit, all conglomeratic clasts can be identified as derived from formations of the underlying Middle Proterozoic Deer Trail Group. Sparse, but ubiquitous, nearly spherical, millimeter-size sand grains appear to be a characteristic of all Windermere clastic units; this suggests that a small component of sediment came from a stable, long-term source. These grains contrast sharply with the bulk of Windermere sediment, which is angular, poorly sorted, and lithologically highly varied laterally and vertically in most sections.

Little is known about the orientation or configuration of the Windermere depositional wedge or basin in Washington and Idaho. Except for the upper coarse-grained clastic unit, the generally coarse clastic rocks of the Windermere Group display few sedimentary structures. They have undergone considerable deformation, so directions and processes of sediment transport are poorly understood. Even the fine-grained, largely argillitic parts of the sequence have few sedimentary structures that can be used to determine transport direction. Moreover, the Windermere Group is exposed only in a relatively narrow belt, so transport directions are not readily determined from east-west changes in lithofacies. On the basis of cross-stratification measurements in the upper coarse-grained clastic unit (Three Sisters Formation), Devlin and Bond (1988) suggest that between present-day latitudes 49°N and 50°N, the paleoshoreline trended northeast and that the paleocurrent direction for fluvial sedimentation was about due north. Some

[1] Devlin and Bond (1988) consider this coarse-grained clastic unit, the Three Sisters Formation (Walker, 1934), to be part of the Hamill Group (Evans, 1933), and the underlying fine-grained clastic and carbonate unit, the Monk Formation, to be the uppermost unit of what they call the Windermere Supergroup in southernmost British Columbia and northeastern Washington State. In making this reassignment, they cite the lithologic similarity of the Three Sisters to, and its gradational contact with, the Quartzite Range Formation (Walker, 1934) (equivalent to the Gypsy Quartzite in northeastern Washington), as well as its lithologic dissimilarity to, and its lack of gradation with, the Monk Formation. In the Salmo–Priest area, the lower part of the Three Sisters Formation (Canadian usage) appears to be gradational with the Monk Formation, so the Three Sisters is here retained in the Windermere as originally defined in the Salmo map-area of British Columbia by Walker (1934).

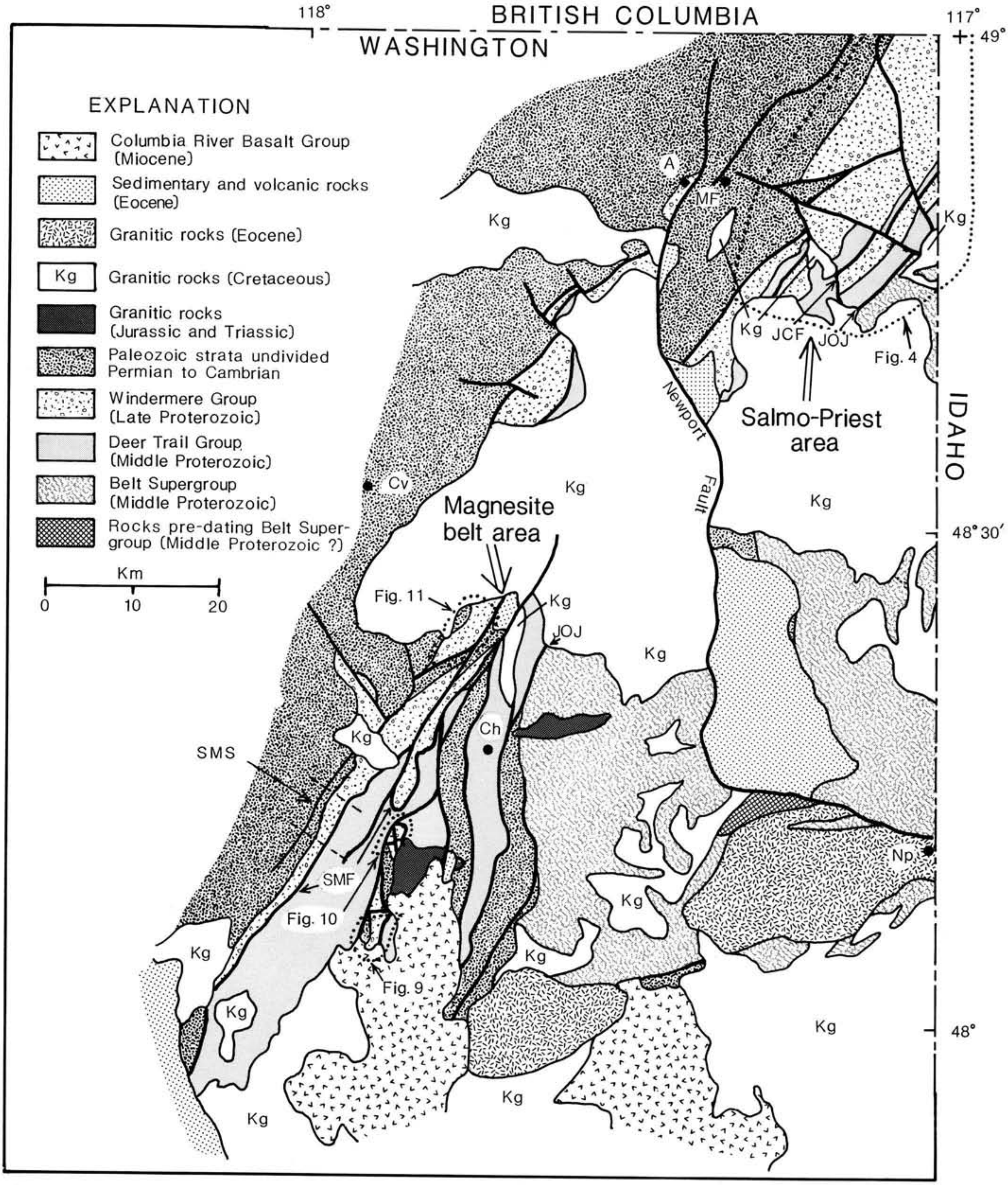

Figure 2. Generalized geologic map of area in Figure 1, showing distribution of the Windermere Group in Washington. Areas shown in Figures 4, 9, 10, and 11 are indicated by dotted outlines. SMF, Stensgar Mountain fault; JOJ, Jumpoff Joe fault; JCF, Johns Creek fault; SMS, Stensgar Mountain sequence; A (with solid diamond symbol), Abercrombie Mountain sequence. Cv, Colville; Ch, Chewelah; Np, Newport; MF, Metaline Falls. The northern 30 km shown as the Newport fault is the Flume Creek fault (Park and Cannon, 1943); the Flume Creek is the steepened northward extension of the Newport.

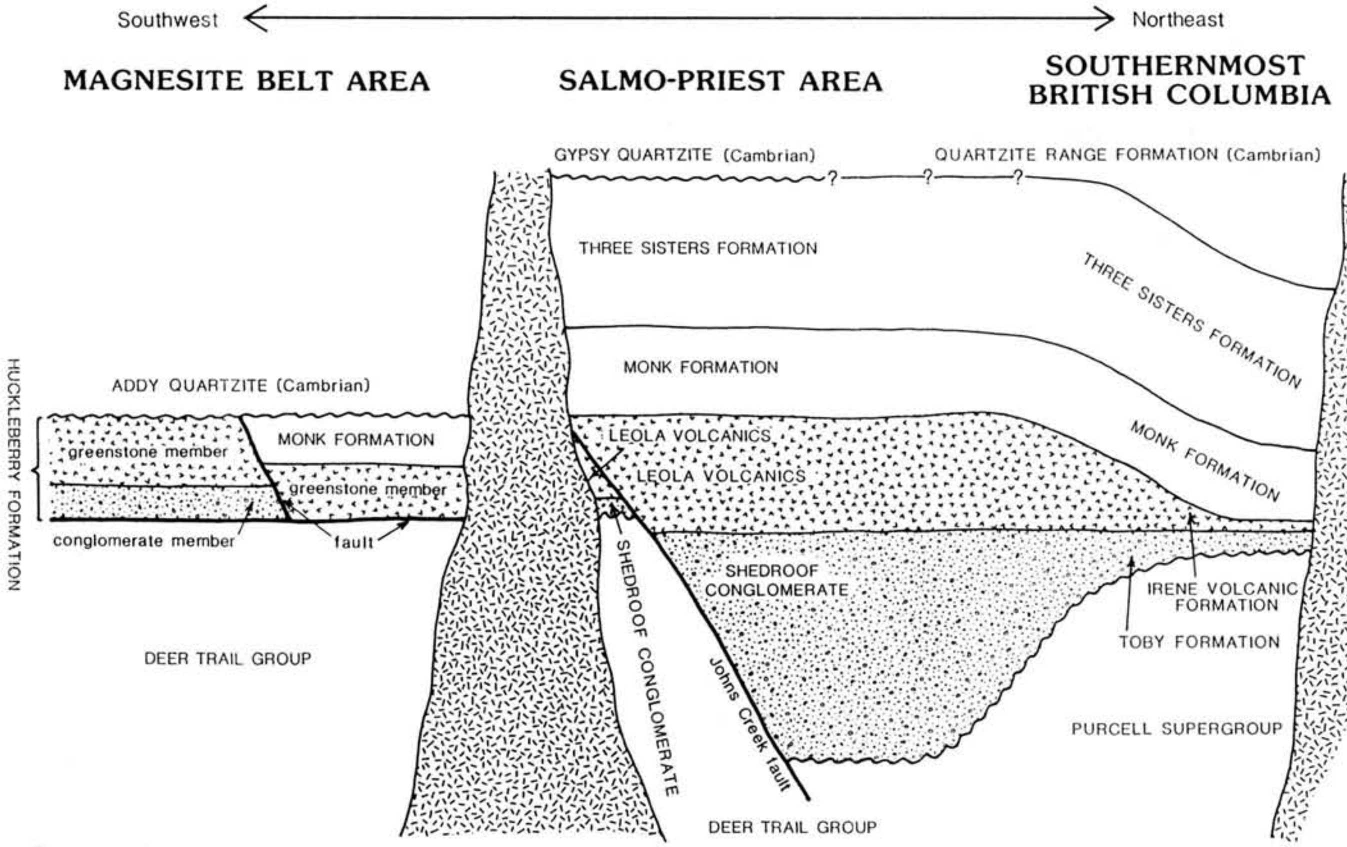

Figure 3. Diagram showing the relation between formational names of the Windermere Group in the U.S. and the Windermere Supergroup in British Columbia, Canada. Relative thicknesses of formations are shown diagrammatically. Random dash pattern, Cretaceous granitic rocks separating sections; random v pattern, main volcanic (greenstone) unit of the Windermere Group; mixed dot and circle pattern, lower conglomerate unit of the Windermere Group.

parts of the Windermere Group are considered to be of glacial or glacial-marine origin (Aalto, 1971; Eisbacher, 1981, 1985).

Shedroof Conglomerate and Conglomerate Member of the Huckleberry Formation

Salmo–Priest area

Sullivan Creek sequence. The Shedroof Conglomerate is the stratigraphically lowest formation of the Windermere Group in the Salmo–Priest area. It is thickest and best exposed 11 km east of Metaline Falls in the Sullivan Creek drainage, where the formation unconformably overlies the Middle Proterozoic Priest River Group[2]. The formation in the Sullivan Creek sequence (Fig. 4) is 3,250 m thick as calculated from outcrop width (Miller, 1983). True stratigraphic thickness may be less because of repetition by undetected reverse faults and by reverse slip along nearly ubiquitous microshears (Fig. 5). Matrix-supported conglomerate is by far the most abundant rock type, making up 70 to 90 percent of the formation at most places (Fig. 6); diamictite, feldspathic quartzite, sandy phyllite, sandy limestone, and metabasalt, in proportions that vary from place to place, form the remainder (Fig. 7).

Bedding is poorly developed or obscured, except where feldspathic quartzite beds are found. These quartzite beds are lensoidal and range from about 10 cm to 2 m in thickness. They are sparse and unevenly spaced, but are typically separated from each other by hundreds of meters of conglomerate. Clasts in the conglomerate are mostly quartzite, dolomite, and argillite derived from the underlying Deer Trail Group, but they also include some two-mica granitic rock of unknown provenance. Nearly all clasts are flattened, some to an extreme degree; the plane of flattening ranges from parallel to bedding to about 70 degrees from bedding (Fig. 5). Matrix of the conglomerate consists of quartz, feldspar, micas, chlorite, sand-size lithic grains, and carbonate and opaque minerals; the micas may be either detrital or recrystallized argillaceous material, or

[2] The Priest River Group is the same lithostratigraphic sequence as the Middle Proterozoic Deer Trail Group in the magnesite belt to the southwest (Miller and Whipple, 1989). In a paper currently in preparation on the revision of Middle Proterozoic stratigraphy in northern Washington and Idaho, I will recommend that the name "Priest River Group" be abandoned and the name "Deer Trail Group" be applied to all Middle Proterozoic rocks west of the Jumpoff Joe fault. Strictly for the purposes of this paper, I will informally apply these tentatively proposed stratigraphic revisions hereafter in the paper.

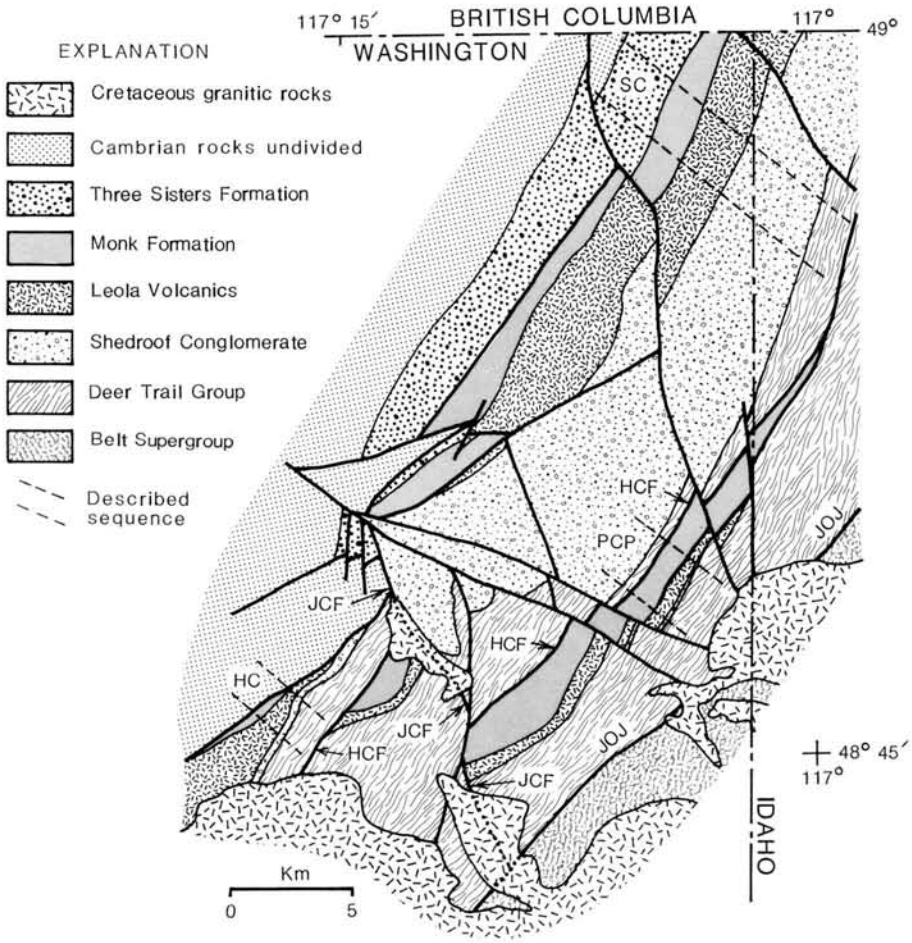

Figure 4. Generalized geologic map of the Salmo–Priest area showing the location of the described sequences; after Burmester and Miller (1983) and Miller (1982, 1983). Northeast strike is characteristic of the structural trend in the southern part of the Kootenay arc. The Belt Supergroup east of the Jumpoff Joe fault (including younger rocks south of the area shown in the figure) typically strikes north, abutting into the fault. The Jumpoff Joe fault appears to be the bounding structure for the southeastern edge of the southern Kootenay arc. The Jumpoff Joe and Johns Creek faults are shown dotted through Cretaceous granitic rocks only to indicate inferred pre-Cretaceous continuity. Except as noted, the same map patterns used here for geologic units are used in Figures 9, 10, and 11. SC, Sullivan Creek sequence; PCP, Pass Creek Pass sequence; HC; Harvey Creek sequence; JCF, Johns Creek fault; JOJ, Jumpoff Joe fault; HCF, Helmer Creek fault.

both. Sorting of matrix material is poor throughout the formation, as is degree of roundness and sphericity of most grains. However, even in rocks made up almost entirely of angular material, sparse to locally abundant well-rounded, nearly spherical, millimeter-size quartz grains are scattered through the matrix, suggesting more than one sediment source.

The lower one-third to one-half of the Shedroof Conglomerate is tan-weathering, matrix-supported, boulder and pebble conglomerate with sparse interbeds of feldspathic quartzite, some of which contain scattered pebbles. Clast size in most of this lower conglomerate ranges from small pebbles to boulders 1 m in length. Clasts make up from 10 to 70 percent of the rock, averaging about 30 percent; dolomite is by far the most abundant clast type. Carbonate minerals constitute from 5 to 30 percent of the matrix and probably are responsible for the color of the rock. This part of the formation contains little chlorite, especially in comparison to the upper part.

The upper one-half to two-thirds of the Shedroof Conglomerate differs from the lower part by having a lower clast to matrix ratio; much of it has only a few clasts per square meter of exposure. The color of the formation grades over several hundred meters from tan to predominantly green, but some tan intervals are found throughout the upper part of the formation. The color change appears to be the result of a decrease in carbonate minerals and an increase in chlorite in the matrix; it may reflect an increasing admixture of pyroclastic or volcaniclastic material, marking the inception of volcanism that predominated following Shedroof deposition. Scattered, discontinuous greenstone layers, some or all of which may be basalt flows, are found throughout the formation, but they are most abundant in the upper half. Poorly stratified green and gray phyllitic quartzite and metagraywacke beds are also abundant in the upper half of the formation.

North of the Johns Creek fault (Fig. 4), the middle part of the Shedroof Conglomerate contains a 2,000-m-thick northeastward-thinning wedge of clastic rocks in which conglomeratic clasts are sparse to extremely rare. This unit is predominantly gray-green phyllite and sandy phyllite; it contains two or more sandy, argillaceous, dolomitic limestone beds as much as 150 m thick that also appear to thin or be discontinuous northeastward.

Pass Creek Pass sequence. The Shedroof Conglomerate in the Pass Creek Pass sequence (Figs. 4 and 7) is only 20 to 40 m thick and appears to pinch out northeastward. It has characteristics of both the upper and lower parts of the Shedroof in the Sullivan Creek sequence (Miller, 1983). Clasts are mostly quartzite and subordinate dolomite and argillite, range from 1 to 20 cm in length, and are angular to well rounded. Matrix material is both green and tan, and all rocks contain the rounded millimeter-size quartz grains characteristic of sedimentary rocks of the Windermere Group. On the ridge north of the Pass Creek Pass road, the

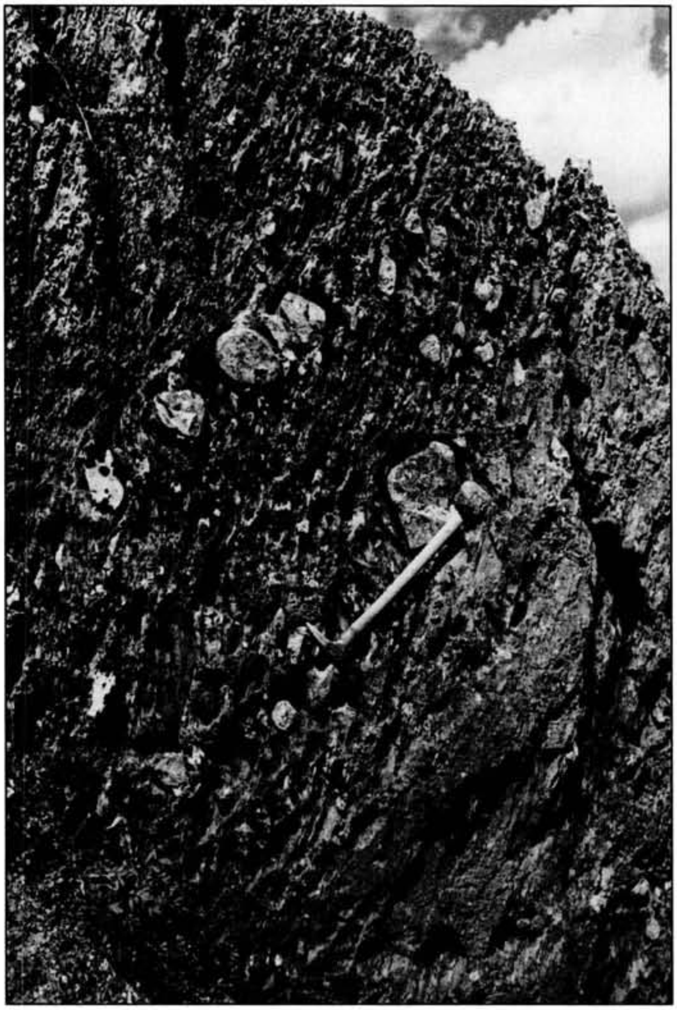

Figure 5. Shedroof Conglomerate; this exposure is near the middle part of the formation, about 3 km south of the U.S.–Canada boundary. Hammer handle is parallel to bedding, which is indicated by a finer grained, nonconglomeratic bed about 1 handle-length below the hammer. Many clasts are rotated parallel to the pervasive microshears, and some, particularly carbonate rocks such as the elongate dolomite clast 1 handle-length to the left of the hammer head, appear flattened. Oval clast 1½ handle-lengths above hammer is leucocratic two-mica granitic rock of unknown provenance.

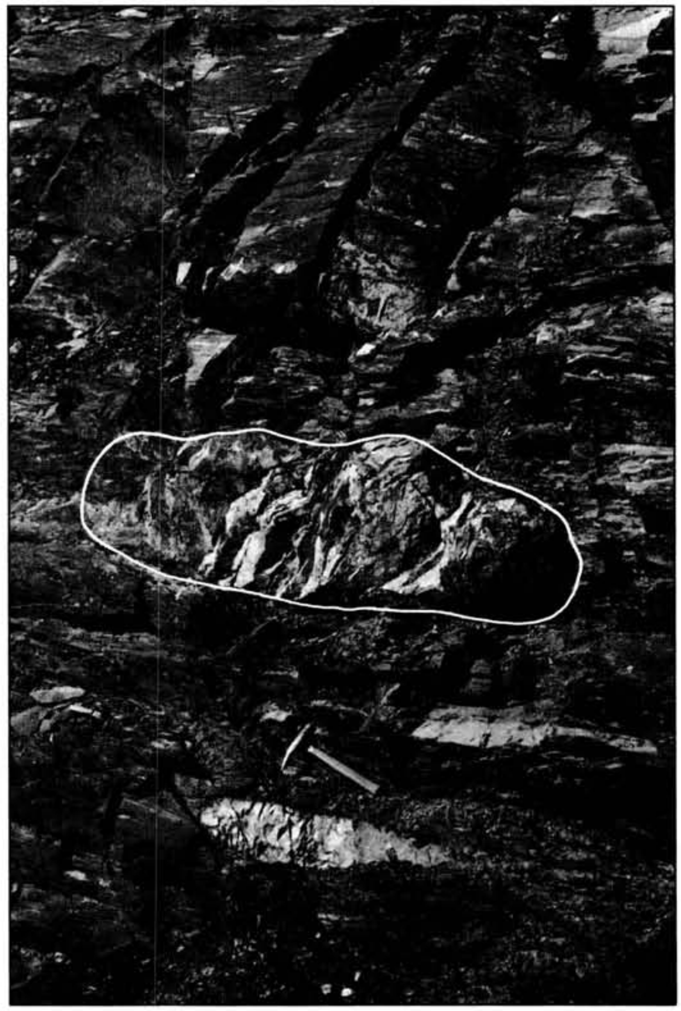

Figure 6. Shedroof Conglomerate; this exposure is in the upper part of the formation, about 4 km north of the Johns Creek fault in a roadcut on the Pass Creek Pass Road. Almost all of the sparse, matrix-supported, flattened clasts are dolomite, probably derived from the Stensgar Dolomite of the Deer Trail Group. The clast with ill-defined boundaries above the hammer is about 1.5 m long. Bedding, flattening, and microshears are approximately parallel at this location; compare to the rocks shown in Figure 5 where microshears and flattening are at a high angle to bedding.

unconformable contact with the underlying Buffalo Hump Formation (the stratigraphically highest formation of the Deer Trail Group) is well exposed and reveals about half a meter of relief on the pre-Windermere surface. The upper contact of the Shedroof is not exposed, but only about 2 m of cover separate the highest conglomerate outcrop from the lowest exposure of the Leola Volcanics. Neither unit is deformed, so the contact is considered depositional.

Harvey Creek sequence. The Shedroof Conglomerate in the Harvey Creek sequence (Figs. 4 and 7) is lithologically similar to that in the Sullivan Creek sequence, but it is only about 200 m thick (Miller, 1983). In the Harvey Creek sequence, tan, matrix-supported conglomerate averaging about 35 percent clasts makes up the lower part of the formation. The upper part is chlorite green, contains several 1- to 2-m-thick basalt flows, and grades into the overlying Leola Volcanics through a zone about 30 m thick. Unlike the upper part of the Shedroof in the Sullivan Creek sequence, clasts in much of the green upper part of the Shedroof in the Harvey Creek sequence are relatively abundant, and the clast to matrix ratio varies widely over small stratigraphic intervals. Clasts are angular to well rounded and range from 1 to 30 cm in length. Dolomite is the most abundant clast type throughout the formation in this sequence, but the proportion of quartzite and argillite clasts appears to increase upward in the section. In the Harvey Creek se-

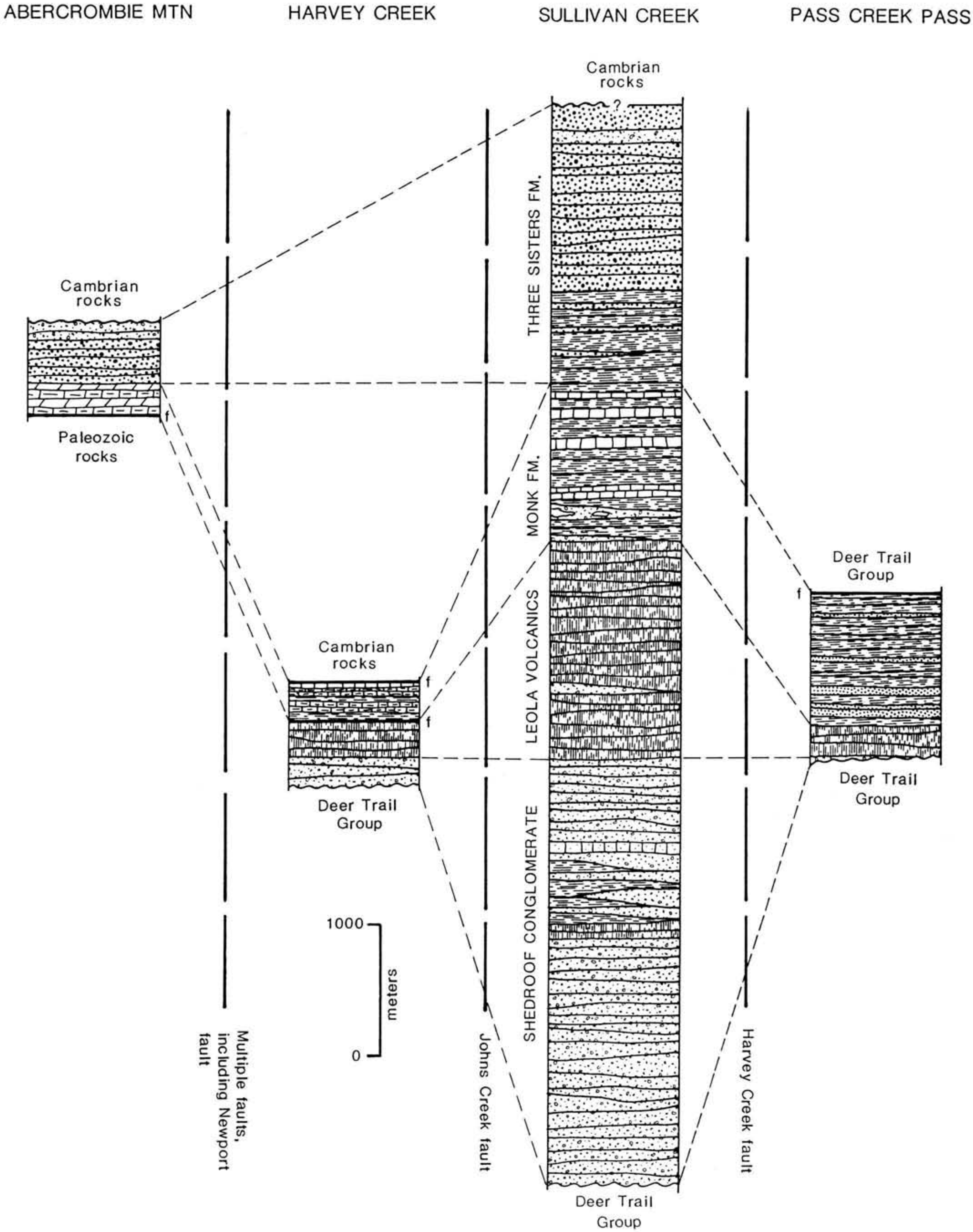

Figure 7. Generalized columnar sections of sequences in the Salmo–Priest area. Heavy lines within columnar sections are faults. Faults separating sequences are shown diagrammatically between columns.

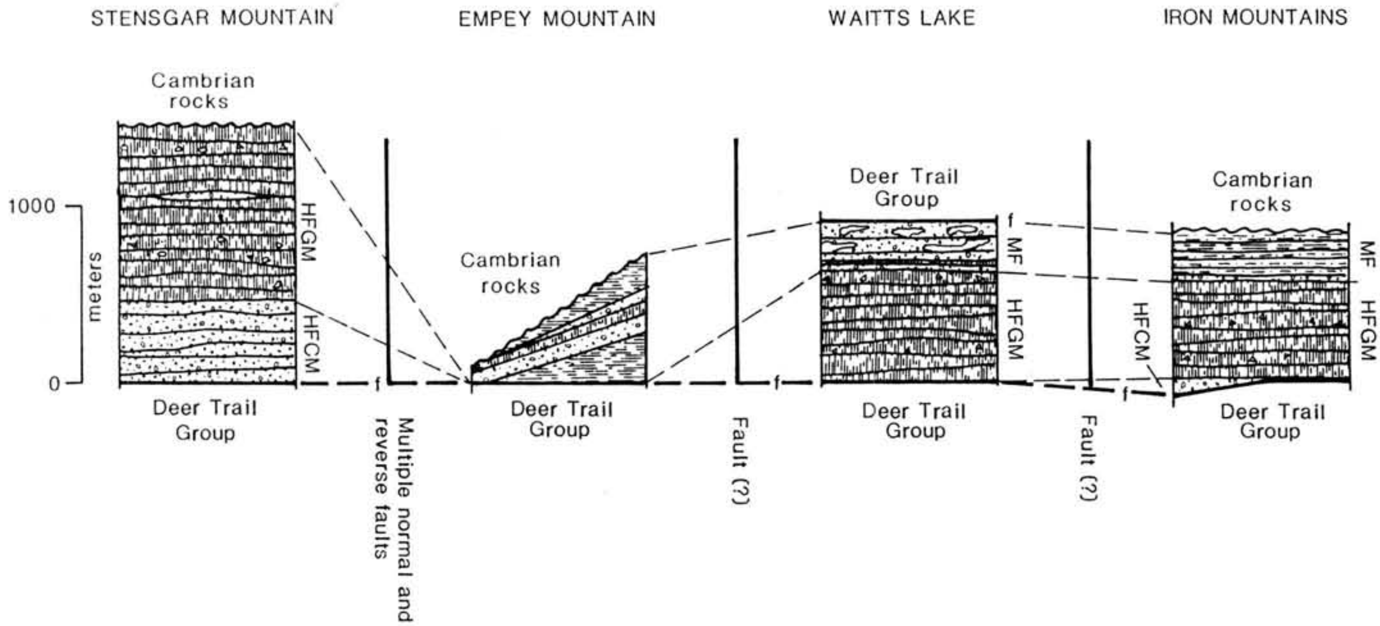

Figure 8. Generalized columnar sections of sequences in the magnesite belt area. Heavy lines within sections are faults. Except for the Stensgar Mountain fault (left), faults diagrammatically shown separating the sequences are unnamed. HFGM, greenstone member of the Late Proterozoic Huckleberry Formation; HFCM, conglomerate member of the Late Proterozoic Huckleberry Formation; MF, Late Proterozoic Monk Formation; f, fault.

quence, the Shedroof lies unconformably on the Stensgar Dolomite of the Deer Trail Group.

Magnesite belt area

Stensgar Mountain sequence. The conglomerate member of the Huckleberry Formation is the stratigraphically lowest unit of the Windermere Group in the Stensgar Mountain sequence (Figs. 2 and 8). It is considered to be correlative with the Shedroof Conglomerate (Fig. 3) on the basis of lithologic similarity and stratigraphic position (Campbell and Loofbourow, 1962; Aalto, 1971). It is best exposed in the Huckleberry Mountains about 20 km southwest of Chewelah, where it averages about 450 m in thickness. The base of the conglomerate member everywhere appears to be faulted (Evans, 1987), but before faulting, the unit is presumed to have been deposited unconformably on the Deer Trail Group. Most of the conglomerate member is composed of uniform pale-gray to pale-green matrix-supported conglomerate. In the central part of the magnesite belt, the lowest preserved 10–50 m of the unit has a medium-gray matrix and is about 30–40 percent clasts. Except in this lower 10–50 m, clasts are sparse and relatively small throughout the unit compared to the average clast concentration and size in the Shedroof Conglomerate. Clasts include dolomite, argillite, siltite, quartzite, and milky quartz and appear to vary randomly in relative proportions; all appear to be derived from Deer Trail formations. Unlike the Shedroof Conglomerate, however, no granitic clasts were identified in the conglomerate member. With respect to the clast to matrix ratio and color, most of the conglomerate member of the Huckleberry Formation resembles clast-poor parts of the upper part of the Shedroof Conglomerate. Matrix material consists of sand- or silt-size quartz, feldspar, sericite, carbonate and opaque minerals, and lithic material, and most rocks in the unit contain millimeter-size, rounded, nearly spherical quartz grains. All of the rock is phyllitic, and all conglomerate clasts are flattened parallel to the phyllitic foliation. Stratification is poorly developed or obscured by the penetrative foliation and, as in the Shedroof Conglomerate, is best defined by sparse, lensoidal quartzite beds that are most abundant in the lower part. Where stratification is identifiable, it is both parallel and oblique to foliation. At most places, the conglomerate member of the Huckleberry Formation grades upward over a stratigraphic interval of less than 5 m into the greenstone member of the Huckleberry.

Empey Mountain sequence. The conglomerate member of the Huckleberry Formation is not present in the Empey Mountain sequence (Figs. 8 and 9). Units of the Windermere Group on Empey Mountain are in fault contact with the underlying Deer Trail Group, so the absence of the conglomerate member may be due to faulting or non-deposition.

Waitts Lake sequence. The conglomerate member of the Huckleberry Formation is not present in the Waitts Lake sequence. As in the Empey Mountain sequence, the lower contact of the Windermere Group here is a fault (Fig. 10). About 6 km north of the Waitts Lake sequence, but separated by a number of faults, about 5–10 m of the conglomerate member is found between the fault at the base of the Windermere and the depositionally overlying greenstone member of the Huckleberry. If these relations extend southward, it would mean that the conglomerate member was deposited in the Waitts Lake sequence but was removed by faulting.

Iron Mountains sequence. In the Iron Mountains, 3 km south of the area shown in Figure 11, about 60 m of the conglomerate member of the Huckleberry Formation is in fault contact with the Togo Formation, the stratigraphically lowest unit of the Deer Trail Group, and is depositionally overlain by the greenstone member of the Huckleberry. The conglomerate is clast poor, pale gray green, and matrix supported, and very similar in most respects to the conglomerate that forms the conglomerate member of the Huckleberry in the Stensgar Mountain sequence.

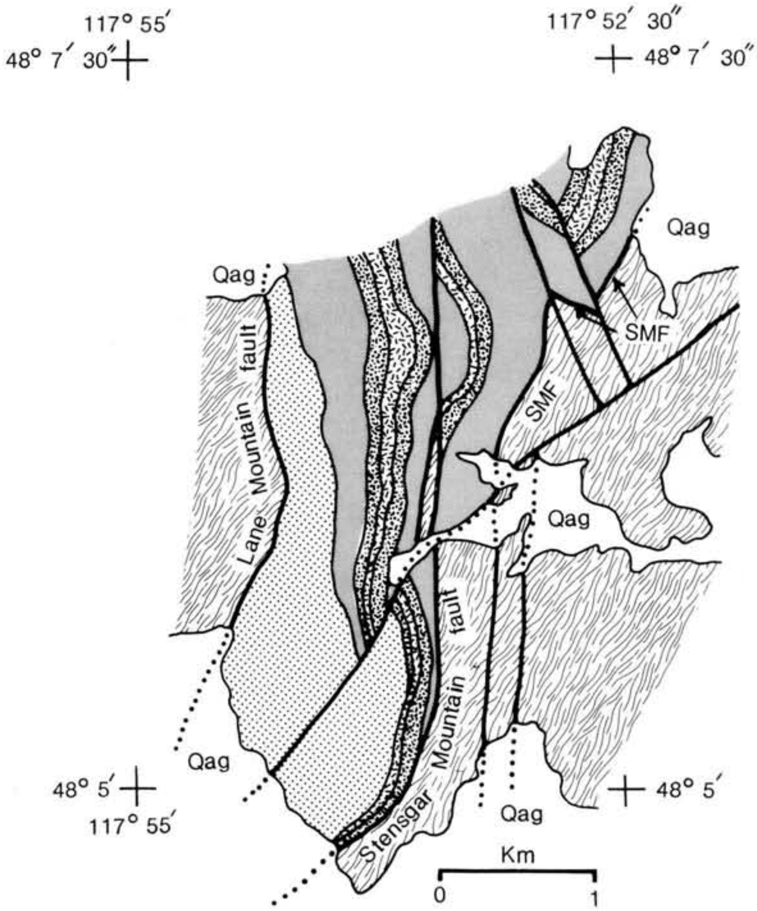

Figure 9. Geologic map of the Empey Mountain sequence showing the relation of the Monk Formation to other units. Except for local variations, all rocks dip westward. The unconformity at the base of the Cambrian Addy Quartzite cuts progressively deeper into units of the Windermere Group southward. An inferred fault, about 1 km north of the area shown in the figure, separates the Empey Mountain Monk section from the Monk on Lane Mountain. All map patterns are the same as in Figure 4, except the dense dot pattern, which delineates conglomerate units in the Monk Formation, and the randomly oriented dash pattern, which denotes greenstone in the Monk Formation, not the Huckleberry Formation as it does in Figures 4, 10, and 11. Qag, Quaternary alluvium and glacial deposits; SMF, Stensgar Mountain fault.

Leola Volcanics and Greenstone Member of the Huckleberry Formation

Salmo–Priest area

Sullivan Creek sequence. In the Sullivan Creek sequence (Fig. 4), the Shedroof Conglomerate is gradationally overlain by the Leola Volcanics (Park and Cannon, 1943), a thick accumulation of metabasalt (greenstone) having an average thickness of about 1,650 m (Fig. 7) as measured from outcrop width. The Leola is composed almost entirely of greenstone except for a few discontinuous lenses of conglomerate and volcaniclastic rocks up to a few kilometers long and 10 m thick. Massive flows, as much as 25 m thick, make up the lower 300–400 m of the unit. These exhibit few primary or secondary structures except for well-developed pillows in the lower 100–150 m and flattened vesicles scattered through all of the formation. The greenstone is composed of millimeter-long pyroxene and plagioclase crystals in a microcrystalline groundmass altered to chlorite, quartz, albite, calcite, epidote, and opaque minerals. All plagioclase is altered to albite and calcite, but most pyroxene is conspicuously unaltered. The upper two-thirds of the unit is about half massive flow rocks and half moderately foliated greenstone probably derived from tuffaceous and volcaniclastic rocks and finely brecciated flow rocks. Texture and primary structures in much of this volcaniclastic and pyroclastic interval, as well as in some of the interlayered flow rocks in this part of the formation, are obscured or obliterated by the foliation (Miller, 1983).

Pass Creek Pass sequence. The Leola Volcanics in the Pass Creek Pass sequence (Figs. 4 and 7) is 250 m thick. The upper and lower contacts appear to be depositional. Here, greenstone of the Leola rests depositionally on the Shedroof Conglomerate, but 3 km to the north, it rests on quartzite of the Buffalo Hump Formation. Except for the difference in thickness and the absence of interbedded conglomerate lenses, the Leola in the Pass Creek Pass sequence is indistinguishable from the Leola in the Sullivan Creek sequence.

Harvey Creek sequence. Lithologically the Leola Volcanics at Harvey Creek (Figs. 4 and 7) is also indistinguishable from the Leola in the Sullivan Creek and Pass Creek Pass sequences. At Harvey Creek the formation is about 275 m thick, but 4 km to the southwest, it is about 650 m thick. The contact with the Shedroof Conglomerate is gradational and depositional, but it is not clear whether the upper contact is depositional or faulted. Younger rocks adjacent to this upper contact are deformed, but not noticeably more so than those in the general area that are away from the contact. However, the progressive southwestward thickening of the formation may result from up-section mi-

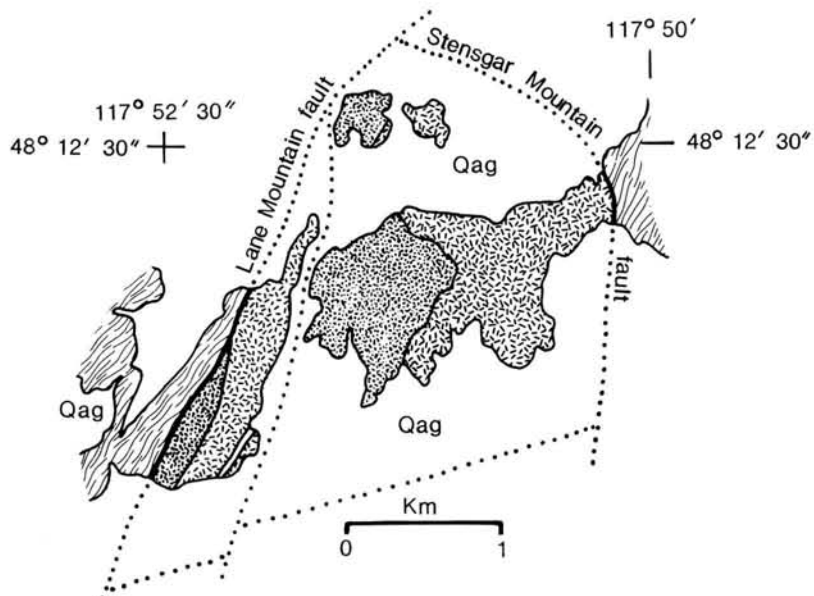

Figure 10. Geologic map of the Waitts Lake sequence showing the relation between rocks of the Windermere Group and surrounding units. All units dip westward. The east-northeast-trending inferred fault separates the Waitts Lake sequence from the Monk Formation on Lane Mountain; the direction of offset on this fault by the younger fault is inferred because the dip of the older fault is unknown. Northward, the Stensgar Mountain fault emerges from beneath the younger west-dipping Lane Mountain fault and does not separate the Waitts Lake and Iron Mountains sequences, both of which lie west of the Stensgar Mountain fault. See Figures 4 and 9 for explanation of map-unit symbols and patterns. The stripe with no pattern is interflow conglomerate.

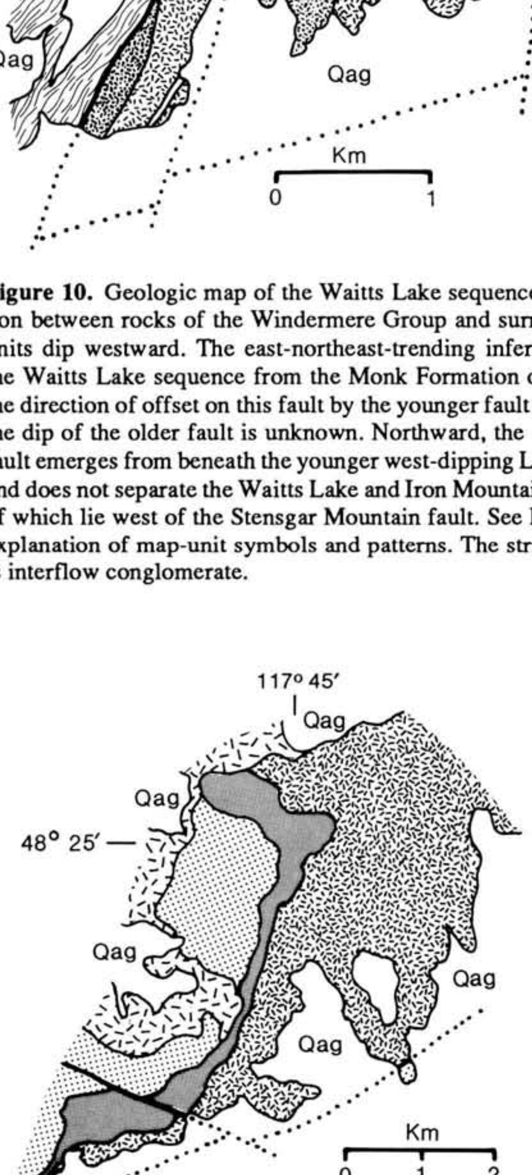

Figure 11. Geologic map of the Iron Mountains sequence showing the relation between rocks of the Windermere Group and surrounding units. All units dip westward. See Figures 4 and 9 for explanation of map unit symbols and patterns.

gration of a fault between the Leola and the overlying Monk Formation.

Magnesite belt area

Stensgar Mountain sequence. The greenstone member of the Huckleberry Formation in the Stensgar Mountain sequence (Figs. 2 and 8) is similar to the Leola Volcanics in almost all respects. Development of the foliation in the flow breccia and pyroclastic part of the greenstone member is about the same as in the Leola. The greenstone member gradationally overlies the conglomerate member of the Huckleberry; it is overlain by the Cambrian Addy Quartzite. The contact with the Addy is an unconformity that cuts progressively deeper into the greenstone member southwestward; at the northeastern end of the magnesite belt area, the greenstone member is about 1,000 m thick, and at the southwestern end it is completely removed by pre-Addy erosion.

Empey Mountain sequence. The greenstone member of the Huckleberry Formation is not present in the Empey Mountain sequence. The Monk Formation is in fault contact with formations of the Deer Trail Group (Fig. 9).

Waitts Lake sequence. In the Waitts Lake sequence, about 600 m of the greenstone member of the Huckleberry Formation is in fault contact with the Stensgar Dolomite of the Deer Trail Group (Figs. 8 and 10) and is gradationally overlain by the Monk Formation; it is not known how much of the lower part of the greenstone member is faulted out in this area. The greenstone member here is about half flow rocks and half volcaniclastic, pyroclastic, and flow breccia rocks. Well-developed pillow structures in the flow rocks are scattered, but common. In most respects, the unit is very similar to the greenstone member in the Stensgar Mountain sequence. About 10–30 m of interflow conglomerate (Fig. 10), in which all clasts are volcanic rock, is found locally about 150 m below the top of the Huckleberry.

Iron Mountains sequence. The greenstone member of the Huckleberry Formation is about 500-600 m thick in the Iron Mountains (Figs. 8 and 11). It is about 65 percent flow rocks; the rest is tuffaceous, pyroclastic, and flow-breccia rocks. The unit is nearly indistinguishable from the greenstone member in the Stensgar Mountain sequence, except that, in the Iron Mountains, it contains no conglomerate and the pyroclastic and flow-breccia parts show relatively little foliation. In the southern part of the Iron Mountains, 3 km south of the area shown in Figure 11, the greenstone member depositionally overlies the conglomerate member of the Huckleberry Formation, but the upper contact of the greenstone member is a fault. In the northern Iron Mountains, in the area shown in Figure 11, the greenstone member is depositionally overlain by the Monk Formation, and its lower contact is a fault.

Monk Formation

Salmo–Priest area

Sullivan Creek sequence. An apparently complete section of the Monk Formation is exposed in the Sullivan Creek sequence (Figs. 4 and 7), where the unit reaches a thickness of about 1,200 m as measured from outcrop width (Miller, 1983). Lithologically, it is the most varied of all formations in the Windermere Group, with respect to both number of lithologies it contains and lateral changes in lithofacies it exhibits. Dark-gray, carbonaceous argillite is the most abundant rock type, but green, laminated argillite, limestone, and diamictite are also abundant; carbonate-bearing quartzite is rare. Approximately the lower 150 m of the formation is fissile, dark-gray, carbonaceous argillite. In this part of the section, a discontinuous bed of green-matrix diamictite as much as 100 m thick and containing a large proportion of volcanically derived material and dolomite clasts is locally found between the argillite and the Leola Volcanics. In the thickest and best exposed Monk section, about 5 km south of the U.S.–Canada boundary, the 150 m of dark-gray carbonaceous argillite is overlain by about 50 to 100 m of fining-upward diamictite that probably was a mud flow or massive slump deposited on the lower part of a submarine slope. Some blocks in this diamictite are tens of meters in length. The diamictite grades bed-by-bed upward into dark-gray carbonaceous argillite lithologically similar to that below; the argillite in turn grades into about 50 m of parallel-planar laminated green argillite. About 20 m of pale-gray thin-bedded limestone containing ½- to 2-m-thick beds of intraformational limestone conglomerate (Fig. 12) overlies the green argillite and grades into more dark-gray carbonaceous argillite. Limestone beds, a few carbonate-bearing quartzite beds, and one pebble-conglomerate bed are scattered through this upper argillite zone, which is about 500 m thick.

Pass Creek Pass sequence. The Monk Formation in the Pass Creek Pass sequence (Figs. 4 and 7) is about 900 m thick as calculated from outcrop width. It differs markedly from the Sullivan Creek sequence because it lacks most of the carbonate rocks and has only about half the carbonaceous argillite of that sequence. Argillite and phyllitic argillite and ½- to 2-m-thick interbeds of distinctive pale-tan feldspathic quartzite make up most of the formation. About 80 percent of the Monk Formation of the Pass Creek Pass sequence is laminated and thin-bedded dark-gray to pale-green argillite and phyllitic argillite; about 50 percent of that is carbonaceous and pyritic. Much of the argillite has very sparsely scattered, millimeter-size, rounded, nearly spherical quartz grains in it. The feldspathic quartzite is distinctive because it has 1- to 3-cm-long, 1-mm-thick chips of dark-gray argillite scattered sparsely through it. Except for a few 1- to 3-m-thick beds, this rock type is not found in the Sullivan Creek sequence. None of the coarse diamictite zones in the Monk Formation of the Sullivan Creek sequence have been found in the Pass Creek Pass sequence, and, except for a 1-m-thick dolomite bed near the middle part of the Monk, no zones or beds of carbonate rocks have been found at Pass Creek Pass.

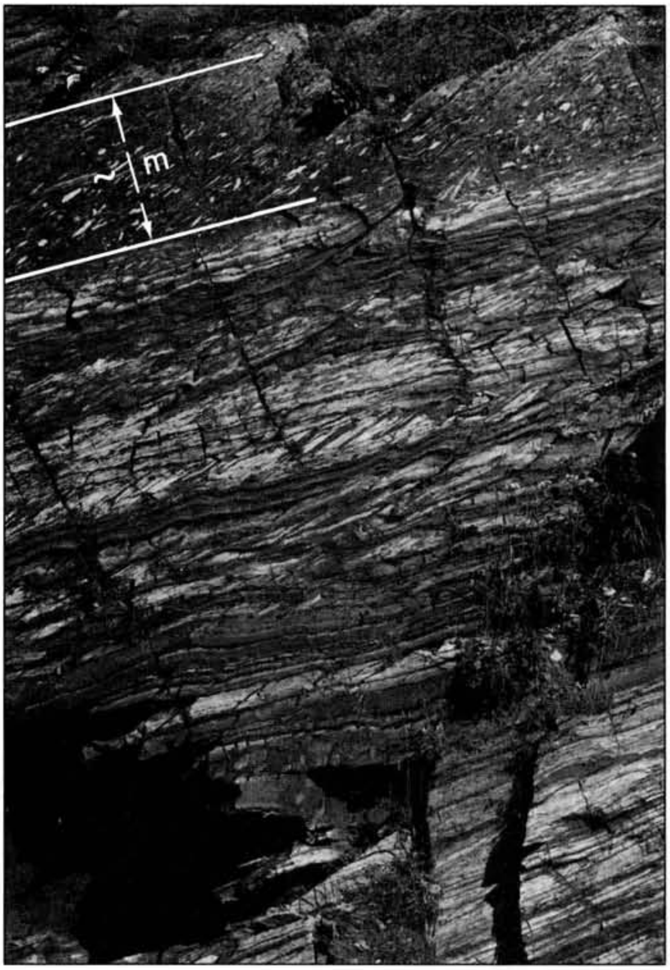

Figure 12. Limestone with intraformational limestone conglomerate (upper part of photograph) in the Monk Formation of the Salmo–Priest area, about 4 km south of the U.S.–Canada boundary in a roadcut along the Salmo Mountain Road. The conglomerate bed is about 1 m thick.

Harvey Creek sequence. The Monk Formation in the Harvey Creek area (Figs. 4 and 7) is mostly thin-bedded, limy, carbonaceous argillite and thin-bedded argillaceous limestone. A few thick beds of fairly pure limy dolomite are found in the upper part, and 1- to 2 m-thick quartzite beds are interlayered with phyllitic argillite in the lower part. The upper contact is a normal fault of probable Eocene age that places the Cambrian Maitlen Phyllite against the Monk. The lower contact may be a fault; if it is, it pre-dates 100-m.y.-old plutons 2 km to the south (Fig. 2).

Magnesite belt area

Stensgar Mountain sequence. The Monk Formation is not present in the Stensgar Mountain sequence. In this area, the greenstone member of the Huckleberry Formation is directly overlain by the Cambrian Addy Quartzite.

Empey Mountain sequence. In the Empey Mountain sequence (Figs. 8 and 9), the Monk Formation is unconformably overlain by the Addy Quartzite, and the lower contact is a fault. The Monk here is divisible, in ascending order, into five informally designated units: (1) argillite-siltite-quartzite, (2) conglomerate-arkose, (3) greenstone, (4) conglomerate, and (5) sandy, phyllitic argillite. Rocks similar to those making up the lower three units are not found in any other Monk section. The argillite-siltite-quartzite unit is a cyclic series of thin-bedded, fining-upward siltite-argillite couplets; many cycles have a thin fine-grained quartzite bed at their base. These rocks are dark gray with submillimeter, parallel-planar laminations in the argillite and some siltite beds. The upper part of this unit is greenish-gray argillite; this has few internal laminations and contains a single 1-m-thick diamictite bed with angular, centimeter-long, matrix-supported dolomite clasts. This unit is conspicuously non-phyllitic compared to nearly all other argillitic rocks in the magnesite belt area. It is sharply overlain by the conglomerate-arkose unit, which consists of repeated tan and brown, parallel-planar, fining-upward cycles of coarse-grained arkose to fine-grained siltite. The cycles are interrupted by one 5-m-thick matrix- and clast-supported conglomerate bed and two others less than 10 cm thick. Clasts are boulders and cobbles derived from rocks of the Deer Trail Group and range from moderately well to very well rounded; most matrix material is moderately angular. Southward in the Empey Mountain area, the arkose and conglomerate beds in this unit decrease in number as diamictite containing angular 1- to 5-cm dolomite and argillite clasts increases. The greenstone unit, which appears to thin southward, is mostly flow rocks, some vesicular, that are lithologically indistinguishable from flow rocks in the greenstone member of the Huckleberry Formation. It is overlain by green matrix-supported conglomerate of the conglomerate unit that contains a high proportion of millimeter-size rounded quartz grains in the matrix. This unit appears to thin southward to about the same degree as the greenstone unit. The sandy, phyllitic argillite unit forming the upper part of the Monk Formation of the Empey Mountain sequence is tan and thin bedded and contains scattered centimeter-thick beds and millimeter-thick lenses of quartzite and siltite. Nearly all rocks of this unit are moderately to highly phyllitic. The unit thins southward; some of this thinning is owing to pre-Addy erosion, some to up-section migration of the fault at the base of the sequence, and some to differences in deposition. Thickness of the Monk on the north slope of Empey Mountain is about 700 m and, in the southernmost exposures on Empey Mountain, about 90 m.

Waitts Lake sequence. The Monk Formation gradationally overlies the greenstone member of the Huckleberry Formation in the Waitts Lake sequence (Figs. 8 and 10); its upper contact is a fault. Unlike the Monk in other areas, here it is almost entirely matrix-supported conglomerate and sedimentary breccia. Angular blocks of dolomite, one nearly 200 m long, are randomly oriented in a matrix of siltite and sandy conglomeratic siltite. Most blocks fall within the size range from ½ to 2 m. Matrix sediments are wrapped around blocks as if the blocks slid into water-saturated sediment; in at least part of this unit, the blocks clearly came into the enclosing sediment and were not deposited with it. Rounded 1- to 6-cm clasts that were deposited with the silty matrix sediments, or as dropstones into it, are mostly quartzite and are very sparse. Many of these clasts bow down laminations in the matrix sediments, but are not accompanied by backsplash features. Nearly spherical millimeter-size quartz grains are sparsely scattered throughout the matrix. The lowermost part of the Monk Formation in the Waitts Lake sequence is boulder and cobble conglomerate overlain by several meters of greenstone. Although the conglomerate is not accompanied by arkose as it is in the Empey Mountain sequence, these lowermost beds and flows may be rough lithostratigraphic equivalents of the two lower Monk units in the Empey Mountain sequence 9 km to the south. The Waitts Lake Monk section is about 300 m thick as calculated from outcrop width.

Iron Mountains sequence. The Monk Formation depositionally overlies the greenstone member of the Huckleberry Formation in the Iron Mountains, and it is unconformably overlain by the Cambrian Addy Quartzite (Figs. 8 and 11). The lower 15 m is dark-gray conglomerate that is both matrix and clast supported. Clasts in the conglomerate are well rounded, many nearly spherical, and they range in size from 2 to 15 cm. Unlike those in any other conglomerate in the region, the clasts are mostly andesite-like volcanic rock whose provenance is unknown. The conglomerate is overlain by 2–10 m of dolomite, which grades upward into thin-bedded, limy siltite. Carbonate content in the siltite diminishes progressively up-section and argillite beds increase in number, so that the upper half of the Monk Formation in the Iron Mountains sequence is interlayered thin-bedded siltite and laminated argillite. Thickness of the Monk is varied due to pre-Addy erosion; the thickest preserved section is about 300 m (calculated from outcrop width).

Three Sisters Formation

Salmo–Priest area

Sullivan Creek sequence. The Three Sisters Formation is composed of quartzite, conglomeratic quartzite, conglomerate, phyllitic quartzite, and phyllite, and in the Sullivan Creek sequence (Figs. 4 and 7) is about 2,100 m thick as calculated from outcrop width (Miller, 1983). The unit, which was named by Walker (1934), derives its name from the Three Sisters Peaks located south of Nelson, B.C.; its principal reference locality is here designated as encompassing the slopes of the Three Sisters Peaks (Salmo 82 F/3 E 1:50,000 quadrangle, British Columbia). Miller (1982) informally extended the Three Sisters into the U.S. at the expense of part of the Cambrian Gypsy Quartzite. (See Park and Cannon, 1943.) Lindsey and others (1990) formalized and refined this extension, and their usage is adopted here. The age of the Three Sisters is considered to be Late Proterozoic on the basis of its regional stratigraphic relations. Rocks comprising this formation were previously included as part of the Cambrian Gypsy Quartzite by Park and Cannon (1943). Approximately the lower 700 m of the Three Sisters Formation is composed of 80 or 90 percent brownish-gray, laminated to massively bedded, phyllitic argillite. Interlayered with the argillite are beds of quartzite and conglomeratic quartzite as much as 5 m thick. These rocks grade upward over a short interval into a 1,200-m-thick zone of multi-hued, thin- to thick-bedded quartzite, conglomeratic quartzite, and conglomerate. The uppermost part of this zone consists of about 100 m of massive to poorly bedded green cobble and boulder conglomerate, in which the clasts range from well rounded and moderately spherical to angular and non-equant. Within the green cobble and boulder conglomerate is at least one 1-m-thick lava flow (greenstone) and about 10–20 m of dark-green chlorite-rich rocks of probable volcaniclastic origin. Above the conglomerate, the upper 200 m of the formation is quartzite and conglomeratic quartzite lithologically similar to that in the thick quartzite zone below the conglomerate.

Pass Creek Pass sequence. Phyllitic argillite, diamictite, quartzite, and conglomerate above the Monk Formation in the Pass Creek Pass sequence are here considered to be part of the Three Sisters Formation. All of the Three Sisters here is intensely deformed, and its upper contact is one of the largest reverse faults in the region. About 100 m of the Three Sisters, as calculated from outcrop width, is preserved in this sequence, but the figure has little meaning owing to the strong deformation. The lower contact is not exposed.

Harvey Creek sequence. The Three Sisters Formation is not present in the Harvey Creek sequence. If originally present, it has been faulted away, as has most of the Monk Formation.

Abercrombie Mountain sequence. West of the Flume Creek fault (on Fig. 2 the Flume Creek fault is not labeled; it is included as the northern 30 km of the Newport fault) on the lower east flank of Abercrombie Mountain (Figs. 1 and 7), the Three Sisters Formation depositionally overlies coarse-grained dolomite of the Monk Formation and is overlain by the Cambrian Gypsy Quartzite (Burmester and Miller, 1983). It is not clear if either contact is conformable or unconformable. About 5 m of vitreous white quartzite overlain by about 2 m of dolomite lithologically similar to that in the upper part of the Monk forms the basal part of the Three Sisters. Above the dolomite is about 350 m of interbedded phyllite, quartzite, and pebbly quartzite. About 120 m of green cobble and boulder conglomerate, quartzite, minor phyllite, and a single basalt flow form the upper part of the formation. The green cobble and boulder conglomerate here is indistinguishable from the conglomerate comprising the 100-m-thick zone of green cobble and boulder conglomerate 200 m below the top of the Three Sisters in the Sullivan Creek sequence. Total thickness of the Three Sisters Formation in the Abercrombie Mountain sequence is about 475 m.

Magnesite belt area

The Three Sisters Formation is not present in any of the magnesite belt sequences.

COMPARISON WITH THE WINDERMERE SUPERGROUP IN SOUTHERN BRITISH COLUMBIA

About 15 km north of the U.S.–Canada boundary, the Toby Formation (lateral equivalent of the Shedroof Conglomerate in the U.S.) thins over a distance of a few kilometers from about 3,000 m to only a few hundred meters (Rice, 1941). Within the range of variation of the formation at the international boundary, the lithologic characteristics of the Toby are largely unchanged for at least 100 km north of where it thins, but the thickness varies several tens of meters (Reesor and LeClair, 1983). If a 40-km apparent offset of the Toby across the Purcell trench shown by Reesor and LeClair (1983) is included, then the Toby extends a strike length of at least 115 km from where it thins.

The Irene Volcanic Formation (lateral equivalent of the Leola Volcanics in the U.S.) thins from about 1,500 m to less than 300 m at about the same place as the Toby Formation (Glover, 1978), but it continues northward for only about 50 km to where it intersects the Purcell trench (Reesor and LeClair, 1983). About 30 km north of the international boundary, the Monk Formation thins from about 950 m to about 650 m and continues northward for about the same distance as the volcanic unit (Reesor and LeClair, 1983). Neither the Irene nor the Monk is found east of the Purcell trench, whereas the Toby is. There appears to be little lithologic variation in the volcanic unit, except that it contains a thin zone of carbonate rocks not found in the Salmo–Priest area and its metamorphic grade increases northward (Rice, 1941). Compared to the Monk Formation in the Salmo–Priest area, the Monk a few tens of kilometers north of the international boundary appears to have a higher proportion of quartzite beds, and more of the argillitic rocks are green and presumably not carbonaceous (Little, 1960; Glover, 1978). Although the Monk here appears to be in the same structural block as in the Salmo–Priest section, some of the lithologic characteristics resemble those of the Pass Creek Pass Monk section east of the Helmer Creek fault.

The lithology and thickness of the Three Sisters Formation just north of the international boundary appear to be similar to those of the Three Sisters in the Salmo–Priest area. Whether due to faulting or nondeposition, the formation does not occur north of Midge Creek, 45 km north of the boundary (Reesor and LeClair, 1983). East of the Purcell trench, a thick sequence of schist, quartzite, and carbonate rocks overlies the Toby Formation (Rice, 1941; Reesor and LeClair, 1983). Höy (1977) assigned the lower part of the sequence east of the trench to the Horsethief Creek Group and the upper part to the Hamill Group. The Horsethief Creek Group there is chlorite and chlorite-muscovite schist with rare beds of quartzite and quartz-pebble conglomerate. The lower 1,600 m of the overlying Hamill

Group is feldspathic quartzite and quartzite. Devlin and Bond (1988) correlated the Horsethief Creek Group and the feldspathic quartzite of the lower part of the Hamill east of the trench with the Monk and Three Sisters Formations, respectively, in a section west of the trench about 12 km north of the boundary. The approximately 700 m of phyllitic argillite with scattered quartzite beds that form the lower part of the Three Sisters Formation in the Salmo–Priest area apparently do not occur or do not crop out in the section 12 km north of the boundary. This lower part of the Three Sisters in the Salmo–Priest area may be the lateral equivalent of the upper part of Horsethief Creek Group east of the Purcell trench described by Höy (1977).

Devlin and Bond (1988) exclude the Three Sisters Formation from their Windermere Supergroup in Canada, emphasizing the change in depositional conditions represented by the difference between (1) sediment gravity-flow deposits interbedded with argillite and carbonate rocks of the Monk Formation and (2) coarse clastic rocks of the Three Sisters Formation which were deposited by persistent current activity in a shallow marine or fluvial environment. They also point out that, in Canada, the transition from the coarse clastic rocks of the Three Sisters to the more uniform and mature quartzites of the overlying Cambrian rocks is gradational.

In the Salmo–Priest area, the contact between the Three Sisters Formation and the Monk Formation appears to be gradational. Rather than the abrupt change from fine-grained clastic and carbonate rocks found a fairly short distance to the north, the Monk–Three Sisters transition in the Salmo–Priest area is from carbonaceous argillite with limestone interbeds to phyllitic argillite with quartzite interbeds. Also, southwestward from the Salmo–Priest area, the major erosional unconformity in the Late Proterozoic through Cambrian sequence appears to be not at the top of the Monk Formation, but at the base of the Cambrian quartzite. Until the significance and extent of the gradational Monk–Three Sisters contact is understood, I recommend that the Three Sisters Formation in the U.S. be considered part of the Windermere Group (following the original usage of Walker [1934] north of the international boundary).

STRATIGRAPHIC EVIDENCE FOR LATE PROTEROZOIC FAULTING

Salmo–Priest Area

In the Salmo–Priest area, the Johns Creek fault (Figs. 2 and 4) separates two sections of the Shedroof Conglomerate of radically different thicknesses. The fault is oriented at a fairly high angle to the Kootenay arc structural trends and to the sedimentary wedge of the Windermere Group itself, if the arc trends are parallel or nearly parallel to the depoaxis of the wedge. In the Sullivan Creek sequence, the Shedroof Conglomerate is more than 3,000 m thick, but in the Harvey Creek sequence, 3 km to the southwest and on the other side of the Johns Creek fault, it is only 200 m thick (Figs. 4 and 7); the upper and lower contacts of the formation in both sequences are depositional and are clearly unfaulted. Detailed knowledge of the internal stratigraphy of the >3,000-m-thick section is not adequate to preclude some thickness changes by bedding-parallel faults, but it is adequate to document that the Sullivan Creek section is at least an order of magnitude thicker than the Harvey Creek section.

This large difference in the thicknesses of two nearby, lithologically similar sections of the Shedroof Conglomerate that meet at the Johns Creek fault is interpreted to indicate that the fault was active at the time of Shedroof deposition, progressively down-dropping the thick northeastern side during all or most of Shedroof time. Thickness of the overlying Leola Volcanics also differs across the fault, but to a much lesser degree, implying that for at least part of its history, the Johns Creek fault was a growth fault. Similar contrasts in unit thicknesses across the Johns Creek fault are not found in overlying Cambrian units, which are also cut by the fault. Displacements of the Phanerozoic units show an opposite sense of throw and resulted from reactivation of the fault in Paleozoic(?) and Mesozoic time.

Geologic relations suggest that a pre-Shedroof-deposition age for the Johns Creek fault is unlikely. Much of the rock making up the thick section northeast of the fault does not reflect the nearby presence of a high-relief source terrane, let alone a 3,000-m-high scarp, at the time of deposition of the Shedroof Conglomerate. Conglomerate adjacent to, and several kilometers away from, the fault is not anomalously coarse or angular relative to the rest of the formation, as would be expected adjacent to a high-relief source area. In fact, the previously described, fine-grained wedge of clast-poor phyllitic rocks in the middle part of the formation appears to be thickest near the fault. Younger faulting southwest of the Johns Creek fault precludes comparison of post-Leola Windermere units across the fault.

In the eastern part of the Salmo–Priest area, the Helmer Creek fault (Fig. 4) separates lithostratigraphically dissimilar sequences of the Windermere Group, but this fault, unlike the Johns Creek fault, parallels the structural strike of the southern Kootenay arc. Windermere rocks in the Pass Creek Pass sequence on the east side of the Helmer Creek fault include essentially unfaulted sections of the Shedroof Conglomerate and Leola Volcanics and a possibly unfaulted section of the Monk Formation. Compared to the Sullivan Creek sequence, the Shedroof in the Pass Creek Pass sequence is less than 1 percent as thick, the Leola is less than one-sixth as thick, and the Monk is about three-fourths as thick. Lithologically, the Shedroof and Leola are similar in both sequences, but, as indicated in descriptions earlier in this paper, the Sullivan Creek Monk section differs conspicuously from the Pass Creek Pass Monk section. Although the Helmer Creek fault separating the two sections is largely a post-Windermere thrust fault because it duplicates the entire Windermere and part of the Deer Trail Group, the pronounced thickness and lithofacies differences between corresponding formations in each sequence strongly suggest that (1) the distance between the two unlike sequences has been severely telescoped by the Helmer Creek fault, or (2) although Phanerozoic telescoping took place, it may have been only moderate in amount, and that the Helmer Creek fault reactivated a Late Proterozoic syndepositional normal fault that had down-dropped the thick Sullivan Creek sequence relative to the Pass Creek Pass sequence.

Stratigraphic and structural relations argue both for and against the Helmer Creek fault being a reactivated Late Proterozoic fault. Thickness of the Shedroof Conglomerate in the Sullivan Creek sequence exceeds by nearly an order of magnitude the thickness of any other basal Windermere conglomerate units in Washington and southernmost British Columbia, indicating an anomalously rapid and localized down-dropping of the Sullivan Creek block. Both the contrast in thickness between like formations in the Sullivan Creek and Pass Creek Pass sequences and the progressive thickening of successive units in the Pass Creek Pass sequence would be expected if, after an initially rapid down-dropping of the Sullivan Creek block and an initially high rate of deposition, the rate of deposition progressively diminished relative to the rate of fault movement.

For at least part of its strike length, the Shedroof Conglomerate in both sequences lies depositionally on nearly the same stratigraphic unit of the Deer Trail Group. For this reason, one or more of the large Late Proterozoic(?) normal faults in the footwall section east of the Windermere formations, rather than the Helmer Creek fault, was probably responsible for the relief that produced the coarse detritus in both sections. Differences in the rate of movement between these normal faults and the Helmer Creek fault could account for the two Shedroof sections of disparate thicknesses being deposited on the same Deer Trail formation, if both were deposited subaqueously. If a rising footwall block of the Helmer Creek fault were the chief source for detritus of the thick Sullivan Creek Shedroof section, then the thin Shedroof section eventually deposited on this footwall block should lie on a Deer Trail formation older than the one that the thick Shedroof section in the hanging wall lies on. The fact that the basal part of the Windermere of the Harvey Creek footwall block lies on erosionally resistant orthoquartzite—and does so for a strike-length of at least 25 km—should also be taken into account.

The Three Sisters Formation on the lower slopes of Abercrombie Mountain is of interest because it is the southwesternmost exposure of the unit and because it differs in both thickness and lithofacies from the Three Sisters in the Sullivan Creek sequence, 11 km to the east. In the Abercrombie Mountain sequence, about 475 m of the Three Sisters Formation overlie a poorly exposed partial section of the Monk Formation, and the Three Sisters is overlain there by the Gypsy Quartzite. Lithologically, the Abercrombie Mountain Three Sisters section strongly resembles the upper part of the Three Sisters in the Sullivan Creek sequence, but it is highly attenuated, and the phyllitic lower part of the unit found in the Sullivan Creek sequence is conspicuously absent. In addition, the overlying Gypsy lies directly on the green cobble and boulder conglomerate unit common to both sequences, but without the intervening 200 m of conglomeratic quartzite found in the Sullivan Creek sequence. The differences between these two relatively close Three Sisters sections may or may not be attributable to Late Proterozoic faulting. Because of the incomplete exposure and the deformation associated with the Newport fault (called the Flume Creek fault by Park and Cannon [1943] in this area), it is not clear if the Three Sisters–Gypsy contact is an unconformity or if the Three Sisters–Monk contact is a fault.

Magnesite Belt Area

In the magnesite belt area, the Windermere Group is considerably thinner than it is in the Salmo–Priest area, and most of the preserved sequences are bounded by faults and incomplete. In addition, southwestward from the Salmo–Priest area, the unconformity at the base of the Cambrian section cuts progressively deeper into the Late Proterozoic rocks. Because most of the sequences are incomplete, interpretations as to the cause for differences in lithostratigraphy and thickness between Windermere units within the magnesite belt area, even large differences between nearby sections, are far weaker than similar interpretations in the Salmo–Priest area. However, the fact that similar processes undoubtedly were active in both areas offers some justification for extending interpretations made in the Salmo–Priest area southwestward.

Using structure sections across the magnesite belt area to restore Phanerozoic offsets of sequences of the Windermere Group does not result in a geometric arrangement of sedimentary sequences that can be interpreted to be parts of a single integrated sedimentary prism, such as a deep distal facies outboard from a shallow or near-sediment-source facies. Neither do the restored sequences seem to indicate consistent directions of deepening or shallowing. Instead, differences in some formations from one sequence to another, particularly the Monk Formation and the conglomerate member of the Huckleberry Formation, suggest a chaotic pattern of differing sedimentary environments through at least part of Windermere time. Post-Windermere faulting could, and probably did, contribute to this chaotic pattern, but cannot fully account for it. In all instances, faults, most of which cut across the dominant structural trend of the region, divide adjacent Windermere sequences having dissimilar lithostratigraphy and (or) thickness. These faults are considered to be syndepositional (although some were reactivated) and responsible for development of the adjacent, dissimilar sedimentary environments. Many of these faults are preserved only as segments bracketed by the younger post-Windermere faults that control the dominant structural trends of the region.

In contrast to the conglomeratic units above and below it, the greenstone member of the Huckleberry Formation is lithologically uniform throughout the magnesite belt area. Slightly varied proportions of flow rocks or volcanic breccia are the only noticeable differences in the formation from sequence to sequence. The unit's thickness also seems to be about the same throughout the area, even though its lower contact is a fault at many places. Why this unit maintains relative uniformity throughout the area, whereas the units above and below it do not, is not understood. The extrusion of the greenstone protolith may have taken place in a fairly short time compared to depositional duration of the other units.

In the magnesite belt area, the Stensgar Mountain sequence differs from all other sequences, but it is not clear if the differences are the result of Late Proterozoic faulting. The Stensgar Mountain sequence is the only sequence in which the lower (conglomerate) unit of the Windermere Group is present. In addition, it is the only sequence from which the Monk Formation is absent; Cambrian quartzite lies unconformably on the greenstone member of the Huckleberry Formation everywhere in this sequence. The lower (conglomerate) unit, for its entire strike length of several tens of kilometers, is fairly uniform with respect to thickness and lithofacies, even though its lower contact is everywhere a fault. This uniformity contrasts with the extreme variations in thickness and lithofacies from one fault-bounded Monk section to another, even over short distances. Comparison of Monk sections in the Iron Mountains, Waitts Lake, and Empey Mountain sequences illustrates the magnitude of these variations.

The Monk Formation exposed in the Iron Mountains overlies the greenstone member of the Huckleberry Formation and is unconformably overlain by Cambrian quartzite. Because of the unconformity, thickness of the Monk at this locality ranges from 300 m to less than 10 m. Overall, the Monk in the Iron Mountains is lithologically unlike the Monk in any other sequence, although the upper siltite and argillite part resembles part of the formation in the Empey Mountain sequence and in the Pass Creek Pass sequence in the Salmo–Priest area. The conglomerate that makes up the lower part of the formation in the Iron Mountains is unique because of its clast composition. Nowhere else, including the nearest Monk section 20 km to the southwest, does the Monk Formation contain clasts of andesite-like volcanic rock. Most of these volcanic clasts are rounded, many nearly spherical, a characteristic not known elsewhere in the Monk Formation.

The Monk Formation in the Waitts Lake sequence lies depositionally on the greenstone member of the Huckleberry Formation and consists of sandy siltite containing sparse, rounded, matrix-supported cobbles. This bimodal grain-size matrix encloses large angular blocks of dolomite of the Middle Proterozoic Deer Trail Group. Some of the rounded clasts in the fine-grained matrix may be glacial dropstones, but the large angular blocks have disrupted and distorted the siltite matrix material, and they clearly entered it with a lateral component of motion after the matrix sediment was deposited. Similar rocks are also present in isolated, fault-bounded exposures 6 km to the north; both occurrences probably represent submarine debris flows, possibly seismically triggered, that entered and disrupted sandy silt accumulating near or below wave base.

On Lane Mountain, 2 km to the south, about 200 m of fault-bounded rocks of the Monk Formation consist chiefly of matrix- and clast-supported cobble conglomerate and interbedded sandy conglomerate; sandy mudstone containing large angular clasts, such as those in the Waitts Lake sequence, is absent. The Waitts Lake section is separated from the one on Lane Mountain by an inferred fault concealed by alluvium.

About 5.5 km south of the Lane Mountain section of the Monk Formation, and also separated from it by an inferred fault concealed by alluvium, the Monk on Empey Mountain differs from both of the previously described sections. Here, it is about 700 m thick and unconformably overlain by the Cambrian Addy Quartzite (lateral equivalent of the Gypsy Quartzite); the lower part is in fault contact with the Deer Trail Group. The thinly interbedded argillite, siltite, and quartzite in multiple fining-upward couplets that make up the lower part of the formation are unlike any other nearby Monk rocks. Along with the multiple fining-upward conglomerate-arkose beds, these strata probably formed below wave base and reflect cyclic changes in current strength and rare turbidity flows. This section is the only Monk section in the U.S. that contains extrusive rocks (greenstone). Conglomerate above the greenstone is somewhat similar to the sandy conglomerate that makes up the Monk on Lane Mountain, but the only Monk rock in the magnesite belt area resembling the uppermost argillitic unit on Empey Mountain is the argillitic part of the formation in the Iron Mountains sequence, 30 km to the north.

The southern two of these three aligned sections of the Monk Formation are only about 9 km apart; the northern section is 20 km north of the middle one. The lithostratigraphy of each section is unlike that of the others, and as a group, they do not form expected facies transitions. All these sections are faulted and incomplete, but if they were faulted segments of a single section, it would be reasonable to expect some overlap among the three partial sections. The dissimilarity of the three partial sections indicates no apparent overlap and that each is basically different from the others. Although there may be other interpretations explaining the proximity of such dissimilar sections, two of the more likely are: (1) the rocks making up each were formed in different sedimentary environments controlled by syndepositional faulting, and (2) very different Monk sections were juxtaposed by later faulting. Intervening valleys and Quaternary sedimentary cover preclude proving either, but cross-cutting fault relations in younger and older rocks bounding the three sections make the latter interpretation unlikely.

The unconformity at the base of the Cambrian section that cuts deeper into rocks of the Windermere Group southwestward does not do so at a uniform rate; that suggests that Late Proterozoic faulting continued at least until after deposition of the Monk Formation. Dissimilar Monk sections in the Empey Mountain and Iron Mountains sequences are bounded by presumed syndepositional faults, and they appear to have been preserved in a series of subsequently tilted fault blocks. Each block was tilted northward after deposition of the Monk Formation and before deposition of the Cambrian Addy Quartzite. In both sequences, down-dropped fairly thick Monk sections at the north ends of the blocks are progressively thinned southward by pre-Addy erosion.

Jumpoff Joe and St. Mary Faults

In southern British Columbia, 60 km northeast of the Salmo–Priest area, a sequence southeast of the St. Mary fault in which Lower Cambrian rocks unconformably lie on the Purcell Supergroup (lateral equivalent of the Belt Supergroup in the U.S.) is separated from a sequence 25 km northwest of the fault, in which 9,000 m of the Windermere Supergroup unconformably overlie the Purcell and are unconformably overlain by Lower Cambrian rocks (Rice, 1941). 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 down-dropped basin of accumulation on its northwest side. The fault was reactivated with an opposite sense of throw in post-Windermere time. (See Fig. 13.)

Unlike the faults described in this paper, which separate dissimilar Windermere sequences and are internal to the Windermere basin, the St. Mary fault in British Columbia separates a sequence that contains a thick section of the Windermere Supergroup from one in which no Windermere is present; in the Late Proterozoic, the fault appears to have been a major bounding structure for the basin. The same relations between Belt (Purcell) Supergroup–Windermere Group–Cambrian rocks are found across the Jumpoff Joe fault near Chewelah (Fig. 2), about 150 km southwest of the St. Mary fault. East of the Jumpoff Joe, Lower Cambrian quartzite lies unconformably on a variety of Belt formations (Miller and Clark, 1975), and west of the fault, Windermere rocks lie between the Lower Cambrian quartzite and the Deer Trail Group (lateral equivalent of the upper part of the Belt Supergroup). Here, however, the thickest preserved Windermere section is only about 1,450 m thick, compared to 9,000 m of the Windermere west of the St. Mary fault.

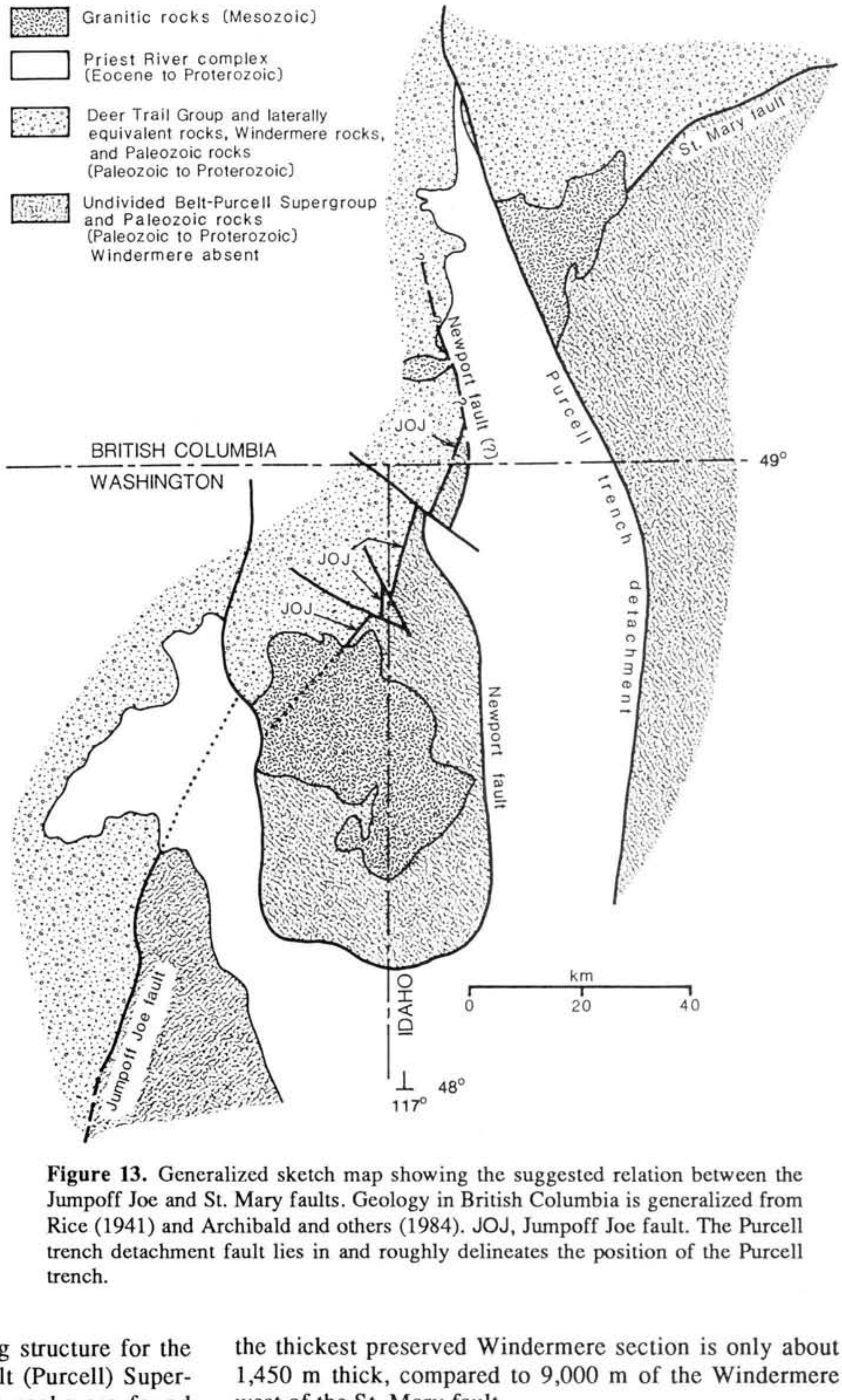

Figure 13. Generalized sketch map showing the suggested relation between the Jumpoff Joe and St. Mary faults. Geology in British Columbia is generalized from Rice (1941) and Archibald and others (1984). JOJ, Jumpoff Joe fault. The Purcell trench detachment fault lies in and roughly delineates the position of the Purcell trench.

Recent mapping has extended the Jumpoff Joe fault northeastward to the international boundary. At that point, the fault is roughly aligned with the St. Mary fault 55 km to the northeast, and it separates a sequence on the east side, in which Cambrian quartzite unconformably lies directly on rocks of the Belt Supergroup, from a sequence on the west in which about 8,000 m of the Windermere Group lie between the Cambrian and Belt-equivalent rocks. The area between the Jumpoff Joe and St. Mary faults is largely underlain by Cretaceous granitic rocks (Archibald and others, 1984) that post-date latest movements on both faults and by a complexly deformed and metamorphosed assemblage of Proterozoic to Mesozoic sedimentary rocks. The Purcell trench and possible Eocene (Rehrig and others, 1987; Armstrong and others, 1987) and older (Archibald and others, 1984) faults that may be concealed in it also intervene between the Jumpoff Joe fault and the St. Mary fault. Restoration of Eocene offset on an east-dipping normal fault in the Purcell trench would move the northeast end of the Jumpoff Joe fault in the direction to more closely align it with the southwest end of the St. Mary fault, but the amount of movement to be restored is unknown. Because of the similarity of structural and stratigraphic relations across the two faults and their approximate alignment, the Jumpoff Joe and St. Mary faults may once have been either a single continuous structure or parts of a single fault system, and that fault or fault system represented a major Late Proterozoic bounding structure for the Windermere sedimentary wedge.

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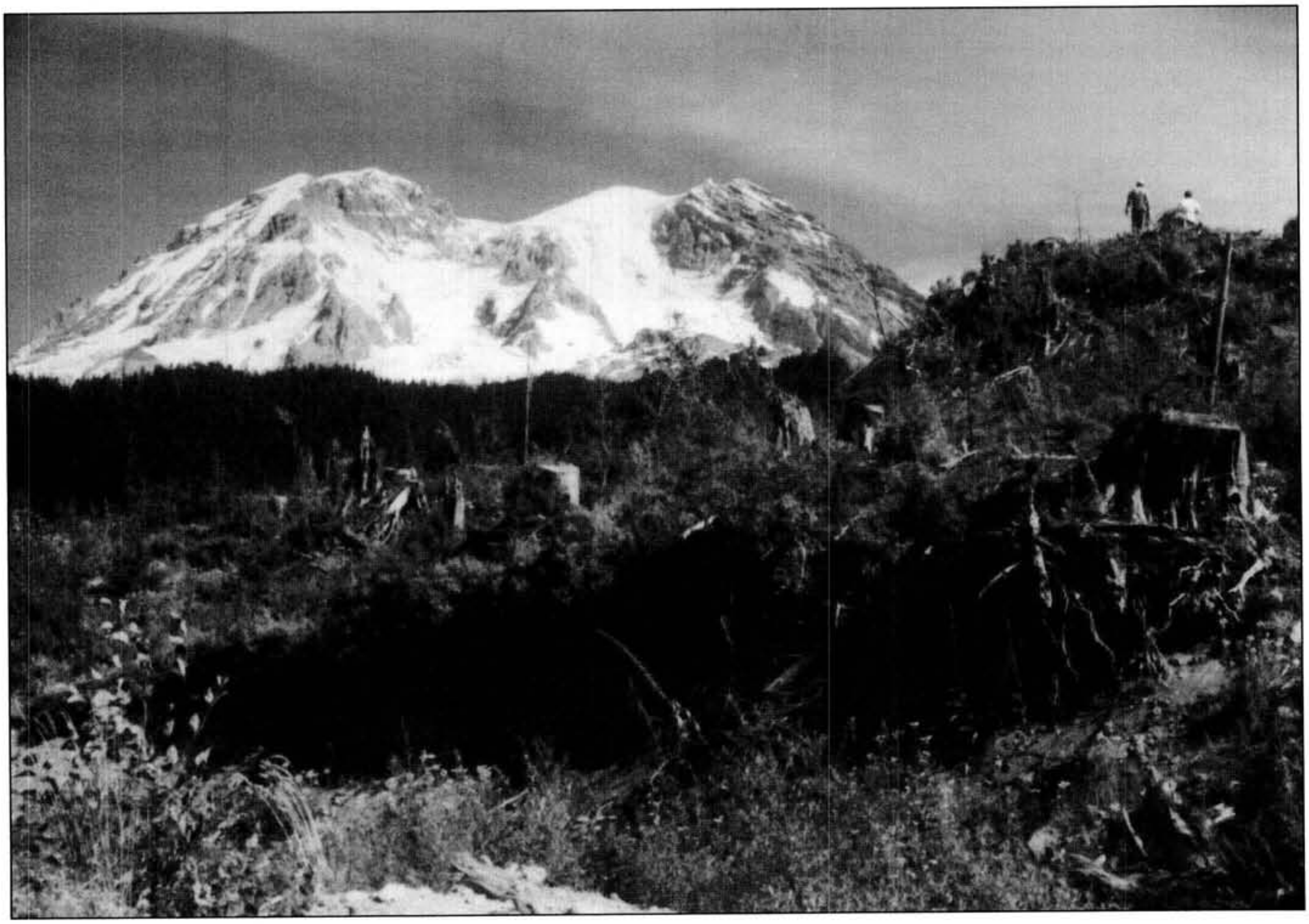


Mount Rainier from the west-southwest, 14 km from the summit. Geologists Dwight R. Crandell and David Frank stand on a megaclast of Mount Rainier rock that was rafted down about 2,600 radiocarbon years ago in the Round Pass Mudflow. The block was stranded on the west wall of the South Fork Puyallup valley about 250 m above the current river level. (The east wall of the valley is the darker forested area on the left side of the photo.) The Round Pass Mudflow is one of several large, relatively clay-rich "sector collapses" that have occurred within the past 3,000 years and cannot be correlated with any known eruptive activity. These failures may have been triggered by earthquakes. Columbia Crest cone, the present summit of Mount Rainier, was constructed some time after about 2,300 years ago. Photo by Patrick Pringle, 1991.

Geology of Metamorphic Core Complexes and Associated Extensional Structures in North-Central Washington

Kenneth F. Fox, Jr.
U.S. Geological Survey
P.O. Box 25046, MS 913
Denver, CO 80225-0046

ABSTRACT

Metamorphic core complexes form much of the bedrock in north-central and northeastern Washington north of the Columbia Plateau. The three complexes described here are the two large northwardly elongated gneiss domes—Okanogan, 75 kilometers long, on the west, and Kettle, 115 kilometers long, on the east—and the more equant and much smaller Vulcan Mountain gneiss dome between.

The gneiss domes consist chiefly of varied proportions of layered paragneiss, schist, quartzite, marble, amphibolite, orthogneiss, and syntectonic granite of the Colville batholith. Rocks in the domes are typically, but not universally, of higher metamorphic grade than adjacent rocks outside the domes. Unlike rocks outside the domes, however, those inside are penetrated at outcrop and commonly hand-specimen scale by uniformly west-northwest-trending lineation and by mylonitic foliation. The foliation is a fluxion structure formed through ductile flow, typically dips gently, and parallels compositional layering. Within each dome, the foliation and layering are subhorizontal in the center and dip gently to moderately outward along the margins.

The domes are bounded on the east and west by normal (detachment) faults dipping away at low angles. Mylonite locally forms a thick carapace on the domes adjacent to these faults, particularly along the west side of the Okanogan and east side of the Kettle dome. Contacts of the deformed rocks of the domes with undeformed leucocratic granite of the Colville batholith and low-grade metamorphic rock overlying the domes are gradational. The detachments and gradational zones separate a lower plate composed of the gneiss domes from an upper plate composed of Eocene volcanic and epiclastic strata, undeformed granite of the Colville batholith, and folded and thrust-faulted Paleozoic and lower Mesozoic strata and Mesozoic plutonic rocks.

The Eocene strata are rotated to moderate to steep and locally vertical dips. They commonly dip into the detachment faults bounding the domes or into north- to northeast-trending low- to high-angle normal faults that splay from the detachments. The north- to northeast-trending faults thus bound tilt-block grabens. These include the Toroda Creek, Republic, and Keller grabens, which occupy the medial zone between the Okanogan and Kettle domes, and the First Thought graben formed along the eastern side of the Kettle dome. Thick accumulations of Eocene rocks are also present fringing the western side of the Okanogan dome near Oroville, Tonasket, and Omak.

The Eocene rocks at Oroville and near Republic are detached from the low-grade metamorphic rocks that formed the floors of the Eocene basins. The Eocene basin fill slid westward along these detachments toward the shifting axial deeps of the basins as the basin floors tilted. At Republic, folds in the Eocene basin fill that formed as a consequence of this movement were subsequently overlain by monolithologic breccia and younger Eocene lavas.

Contact relations and published potassium-argon, fission-track, and uranium-lead age determinations indicate that ductile flow of high-grade metamorphic rock in the core of the gneiss domes was under way shortly after, if not before, intrusion of earliest leucocratic plutons of the Colville batholith in late Paleocene time. Growth of the domes continued concurrently with intrusion of syntectonic granite and eruption of the volcanic rocks in the early to middle Eocene. Early Eocene plutonic rocks that cut the rocks of the core complexes are locally mylonitized; foliation and lineation are conformable with those of the core complexes, indicating that although the complexes had largely formed by this time, ductile flow of their interior was still under way. Thus, although Late Cretaceous ductile deformation in the complexes cannot be entirely ruled out, the complexes probably formed in the Paleocene and early Eocene, concurrently with eruption of the volcanic rocks of the grabens and extension of the crust.

The domal aspect of the fluxion structure, its conformability with bounding detachment faults, and fault contacts with tilted Eocene volcanic and epiclastic strata indicate that the ductile basement bulged upward and, through structural attenuation of superjacent rocks and upward movement of the ductile front, locally intersected nearly coeval basin fills. The domes appear to be part of a Paleocene and Eocene extensional province structurally analogous to the contemporary Basin and Range extensional province. However, in Washington, the Paleocene and Eocene province has been eroded deeply enough to expose the then-ductile basement (the gneiss domes) and remnants of overlying but detached basal parts of brittle crust and synextensional epiclastic and volcanic basin fill.

INTRODUCTION

Five metamorphic core complexes have been identified in north-central and northeastern Washington. These are the rocks in the Okanogan gneiss dome (Fox and Rinehart, 1971), Kettle dome (Cheney, 1976, 1980; Wilson, 1981; Fox and Wilson, 1989), Vulcan Mountain gneiss dome (type area of the metamorphic rocks of Tenas Mary Creek of Parker and Calkins, 1964), Lincoln dome (Atwater and Rinehart, 1984), and Spokane dome (Cheney, 1980; Rhodes and Hyndman, 1984) (Fig. 1). The complexes consist chiefly of penetratively lineated and foliated, variously mylonitic metamorphic rocks forming structural domes or half domes. Most of the complexes are partially enveloped by a mylonitic carapace.

Similar rocks composing the Shuswap Complex of southern British Columbia were described and categorized as gneiss domes by Reesor (1965). These rocks and the analogous rocks composing the Okanogan and Kettle gneiss domes were identified as part of a class whose members included metamorphic complexes scattered along the length of the North American Cordillera (Fox and others, 1977). The complexes are now referred to as metamorphic core complexes (Crittenden and others, 1978). Throughout the Cordillera, the complexes are typically associated with low-angle extensional (detachment) faults and coeval volcanic and plutonic rocks, which thus are likely to be coproducts themselves of extensional processes (Davis and Coney, 1979).

The core complexes of north-central Washington are elements of a Paleocene and Eocene extensional province that also includes highly attenuated low-grade upper crustal metamorphic rocks, Eocene volcanic rocks and associated grabens, and Paleocene and Eocene plutonic rocks (the Colville batholith) that are in part the intrusive equivalents of the volcanic rocks (Fox and Beck, 1985, p. 334; Holder and Holder, 1988) (Fig. 2). The purpose of this report is to summarize information pertaining to three of the complexes—the Okanogan, Kettle, and Vulcan Mountain gneiss domes—and the associated extensional structures and coeval plutonic and volcanic rocks. The geology of the gneiss domes was previously summarized by Orr and Cheney (1987) and the extensional features of the region by Parrish and others (1988).

This report is organized as follows: first the geologic setting is summarized, next the general characteristics of rocks composing the Colville batholith and the gneiss domes are described. The external contacts of the gneiss domes are then discussed. These contacts include detachment faults and associated structures bounding the western flank of the Okanogan gneiss dome and the eastern flank of the Kettle gneiss dome as well as detachment faults and associated normal faults bounding the grabens forming the medial zones between the domes. The external contacts also include gradational zones locally present between the domes and remnants of their former roofs. Structural relations in the Oroville (Fig. 3), Curlew (Fig. 4), and Republic (Fig. 5) areas exemplify features common to the gneiss domes and are therefore described in somewhat greater detail. The concluding summary focuses on three questions: (1) origin of the complexes, (2) timing of extension, and (3) origin of detachment and normal faults associated with the complexes.

K-Ar age determinations cited in this paper have been revised to International Union of Geological Sciences (IUGS) constants using correction factors as given by Dalrymple (1979). Nomenclature of quartzose plutonic rocks follows that of Aramaki (1973).

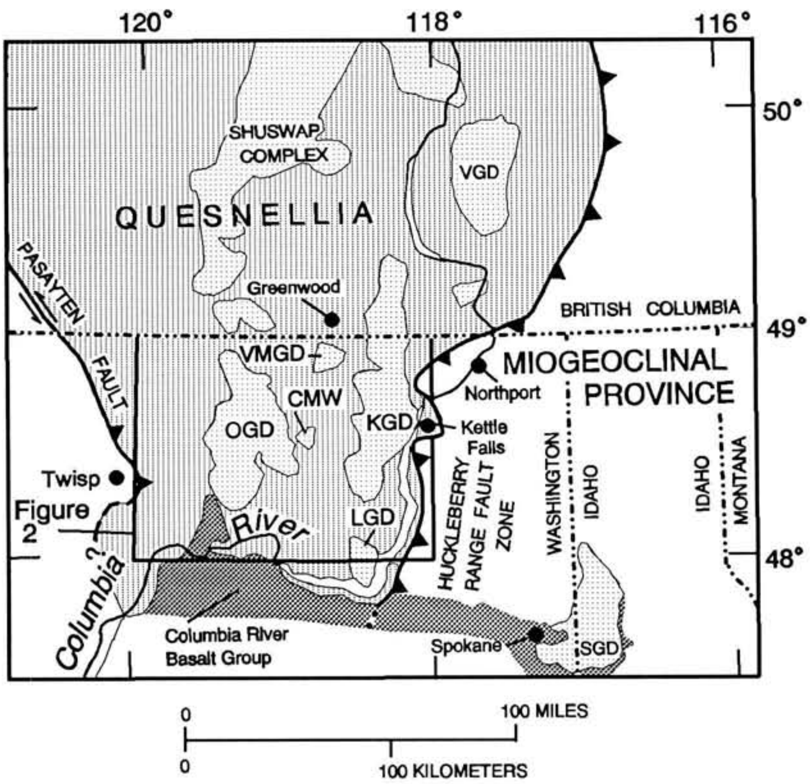


Figure 1. Index map, showing locations of gneiss domes, tectonostratigraphic terranes, and major structural features in north-central Washington and southern British Columbia. CMW, Coco Mountain window; KGD, Kettle gneiss dome; LGD, Lincoln gneiss dome; OGD, Okanogan gneiss dome; SGD, Spokane gneiss dome; VMGD, Vulcan Mountain gneiss dome; VGD, Valhalla gneiss dome.

GEOLOGIC SETTING

Tectonostratigraphic Terranes

Northeastern Washington encompasses parts of two depositional provinces, a miogeoclinal province on the east, and a eugeosynclinal province on the west. The eugeosynclinal province corresponds to a potentially allochthonous microplate (Fox, 1977) or tectonostratigraphic terrane referred to as Quesnellia (Monger and others, 1982) (Fig. 1). Quesnellia is the easternmost of a collage of tectonostratigraphic terranes, some or many of which were accreted to the craton during the Mesozoic or Cenozoic (Beck and others, 1980). Whether Quesnellia is in fact allochthonous is uncertain.

The boundary between the miogeoclinal province and Quesnellia in Washington is arcuate, extending southwestward from near Northport through Kettle Falls and thence southward to where it is concealed by the Columbia River Basalt Group (Cady and Fox, 1984, p. 3). Although the core complexes described in this paper are confined to Quesnellia, other core complexes (for example, Spokane gneiss dome) are present in the craton to the east.

In Washington, the sedimentary rocks of the miogeoclinal province consist of miogeoclinal Precambrian rocks of the Belt Supergroup, Deer Trail Group, and Windermere Group overlain by Cambrian to Mississippian shelf deposits. These rocks are intruded by granitic rocks ranging in age from Jurassic to Tertiary.

Pre-Eocene basement rocks of Quesnellia outside the core complexes in Washington consist chiefly of stratified eugeosynclinal deposits and Jurassic and Cretaceous plutonic igneous rocks. The eugeosynclinal rocks can be grouped into two assemblages, a northern one of Pennsylvanian(?), Permian, Triassic, and Jurassic ages, and a southern one chiefly of Ordovician age. Permian strata are most widespread. They include weakly to moderately metamorphosed, complexly interfingering deposits of sharpstone (chert-pebble) conglomerate, argillite, siltstone, graywacke, limestone, and lava (greenstone) forming the lower part of the Anarchist Group (the Spectacle Formation) in the Loomis area and the Churchill unit of Bowman (1950) in the Orient area (Fig. 2). Similar rocks of Permian age are also present in the Mission Argillite of Weaver (1920) in the Kettle Falls area (Mills, 1985, p. 5). Thickness in the Loomis area is estimated to be at least 4.5 km (Rinehart and Fox, 1972, p. 8).

In the Loomis area, the strata containing Late(?) Permian fossils (Spectacle Formation) are conformably overlain by the Bullfrog Mountain Formation, a 1.5-km-thick sequence of chert-pebble (sharpstone) conglomerate, wacke, and argillite that forms the upper part of the Anarchist Group (Rinehart and Fox, 1972, p. 10). The Bullfrog Mountain unit is unfossiliferous. The lower part of the Middle Triassic Brooklyn Formation (Little, 1983, p. 13) of the Greenwood area of British Columbia is lithologically similar to the Bullfrog Mountain and apparently occupies a similar stratigraphic position.

In the Orient area, the Churchill unit of Bowman (1950) is lithologically similar to the Spectacle, but it includes an apparently conformable succession of Permian through Lower Triassic strata (Kuenzi, 1965; Little, 1978).

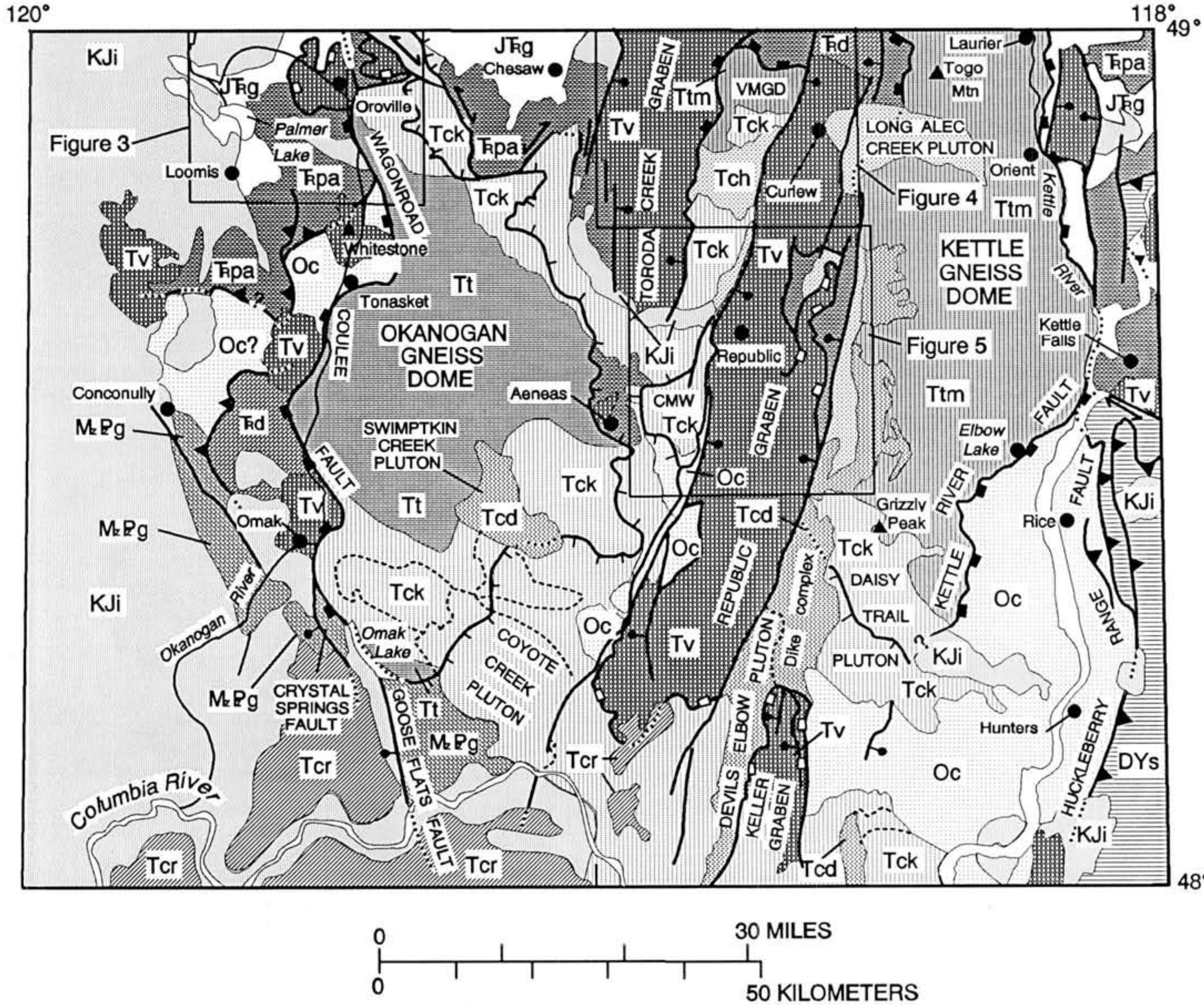
The Permian through Lower Triassic—and perhaps some Middle(?) Triassic—strata are unconformably overlain by Upper Triassic limestone, dolomite, argillite, and minor siltstone and sandstone. The Upper Triassic strata have been recognized in the Conconully (Rinehart and Fox, 1976, p. 9), Curlew (Parker and Calkins, 1964, p. 30), and Orient areas (Fox, 1981). They are well exposed in the Conconully area, where they reach a thickness of 1.3 km (Rinehart and Fox, 1976, p. 9).

In the Conconully, Oroville, and Chesaw areas, dolomite at the base of the Upper Triassic sequence locally grades to magnesite and magnesitic dolomite containing sparse flakes of bright-green fuchsite (Fox and Rinehart, 1968), which forms a useful marker bed. At some localities where the Upper Triassic beds are absent or perhaps very thin and unrecognized, the magnesitic rocks are in conformable contact with the superjacent Kobau Formation. Because of this association, the magnesitic rocks were formerly thought to be part of the basal strata of the Kobau Formation (for example, Fox, 1978). However, in the Conconully area, the magnesitic bed is located at the base of the Upper Triassic carbonate sequence (Rinehart and Fox, 1976, p. 13).

The Permian and Triassic rocks described above are overlain by weakly metamorphosed, stratiform, marine andesite and basalt containing interlayers of quartzite (metachert), volcanic conglomerate, tuff, and locally at the base, sandstone, siltstone, and conglomerate composed of chert and limestone pebbles and cobbles. These rocks have been assigned to the Kobau Formation in the Loomis and Chesaw areas and to the Rossland Group of Little (1982) in the Kettle Falls area. Older units are extensively cut by dikes and sills of mafic intrusive rocks, thought to be the intrusive equivalents of the Kobau and Rossland units.

Rocks assigned to the Rossland Group are chiefly of Early Jurassic age (Little, 1982). Fossils have not been found in strata assigned to the Kobau Formation; hence its age is not known with certainty. Okulitch (1973, p. 1514) concluded that the Kobau was pre-Permian in age because it contained rootless isoclinal folds that he believed were related to a deformation event that predated the Permian Spectacle Formation. However, the Kobau (whose rocks were formerly included in the Anarchist Series of Daly, 1912) stratigraphically overlies the Bullfrog Mountain Formation in the Loomis area (Waters and Krauskopf, 1941, p. 1363; Rinehart and Fox, 1972, p. 22); hence, it cannot be older than Late Permian. In the Chesaw area, McMillan (1979) postulated that the Kobau was thrust over the Permian rocks, forming the Chesaw thrust. Moderate deformation in the Chesaw area is ubiquitous; extreme deformation is common. Furthermore, exposure is much diminished by an extensive cover of glacial deposits. With these caveats in mind, the contact appears to be depositional (Fox, 1978).

The age of the Kobau is bracketed between approximately 198 Ma (Early Jurassic, according to the time scale of Harland and others, 1982), the hornblende K-Ar age of the Loomis pluton that intrudes it (Rinehart and Fox, 1972), and the Permian age of the Spectacle Formation unconformably underlying it. More speculatively, the age of the Kobau may be Late Triassic or Early Jurassic (assuming that the magnesitic marker bed below the Kobau in the Oroville and Chesaw areas is correctly correlated with that at the base of the Upper Triassic rocks in the Conconully area).



EXPLANATION

Contacts

Sharp contact — dashed between plutons within suites

Gradational contact — between penetratively foliated and lineated rocks of gneiss domes and less foliated and lineated rocks outside of gneiss domes; ticks toward less foliated rocks; dotted where engulfed by younger plutonic rocks

Faults — queried where uncertain; dotted where concealed by lake waters or younger rocks and where engulfed by younger plutonic rocks

? Low-angle extensional (detachment) fault bounding gneiss dome — blocks on upper plate

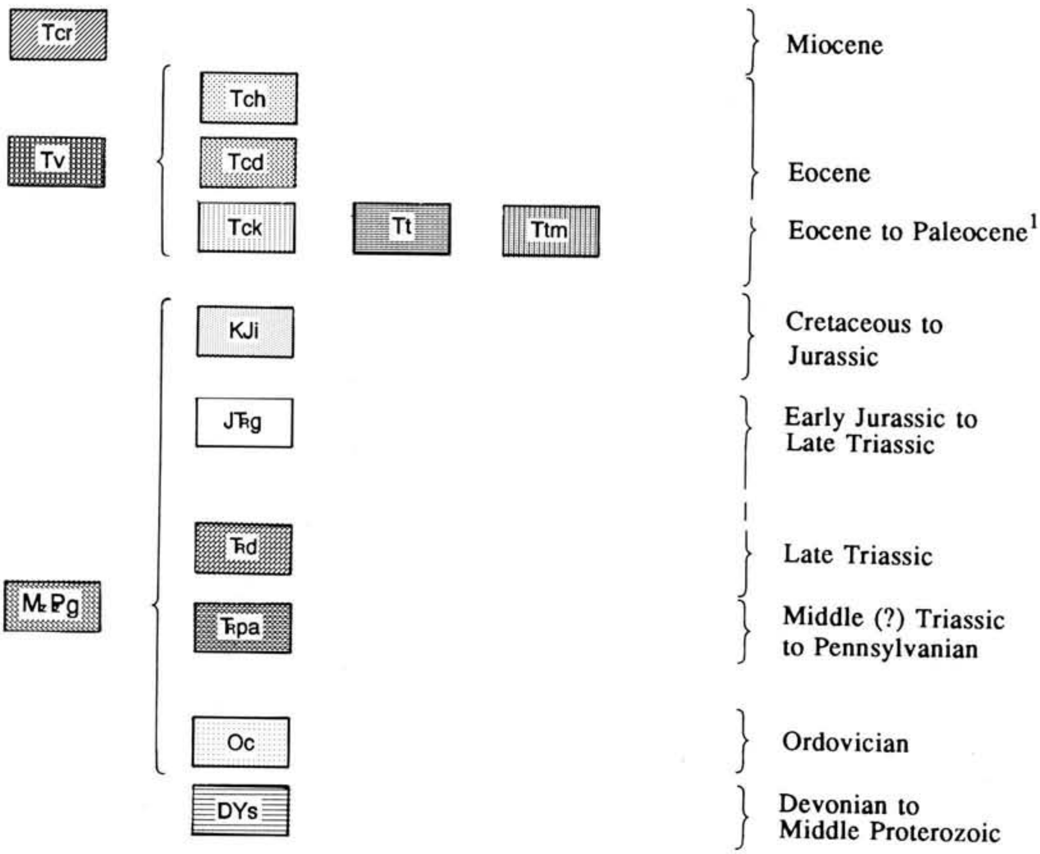
Low-angle extensional (detachment) fault at base of Paleocene to Eocene basin — blocks on upper plate

Thrust fault — sawteeth on upper plate

? Normal fault — ball on downthrown side

Transform(?) fault — high- to low-angle faults subparallel to direction of Eocene crustal extension

Figure 2. Geologic map of the Okanogan 2-degree quadrangle. Modified from sources listed in Stoffel and others (1991). VMGD, Vulcan Mountain gneiss dome; CMW, Coco Mountain window. Low-angle normal (detachment) fault north of Long Alec Creek pluton from Orr and Cheney (1987). Figure continues on following page.



[1] With respect to units Tt and Ttm, refers to age of penetrative dynamic metamorphism. Age of protolith uncertain.

GEOLOGIC UNITS

Tcr Columbia River Basalt Group

Tv Volcanic and sedimentary rocks — includes O'Brien Creek Formation, Sanpoil Volcanics (and related hypabyssal intrusive rocks), Klondike Mountain Formation, and correlative rocks

Colville batholith

Tch Herron Creek suite — fine-to medium-grained hornblende (pyroxene) biotite granite

Tcd Devils Elbow suite — mafic fine-grained hornblende (pyroxene) biotite monzodiorite; locally gneissic

Tck Keller Butte suite — coarse- to fine-grained biotite (muscovite) granite; commonly megacrystic; lineated and gneissose in gneiss domes

Tt Tonasket Gneiss — biotite (hornblende) orthogneiss, mafic amphibole gneiss and schist, amphibolite, and biotite-sillimanite schist

Ttm Metamorphic rocks of Tenas Mary Creek and correlative rocks — chiefly orthogneiss, paragneiss, amphibolite, marble, sillimanite quartzite, and sillimanite schist

KJi Granodiorite, tonalite, quartz monzonite, monzonite, monzodiorite, and syenite

JTRg Greenstone (weakly metamorphosed mafic to felsic lavas and pyroclastic rocks) with interlayers of quartzite (metachert), metasiltite, phyllite, and argillite — includes Kobau Formation and Rossland Group of Little (1982)

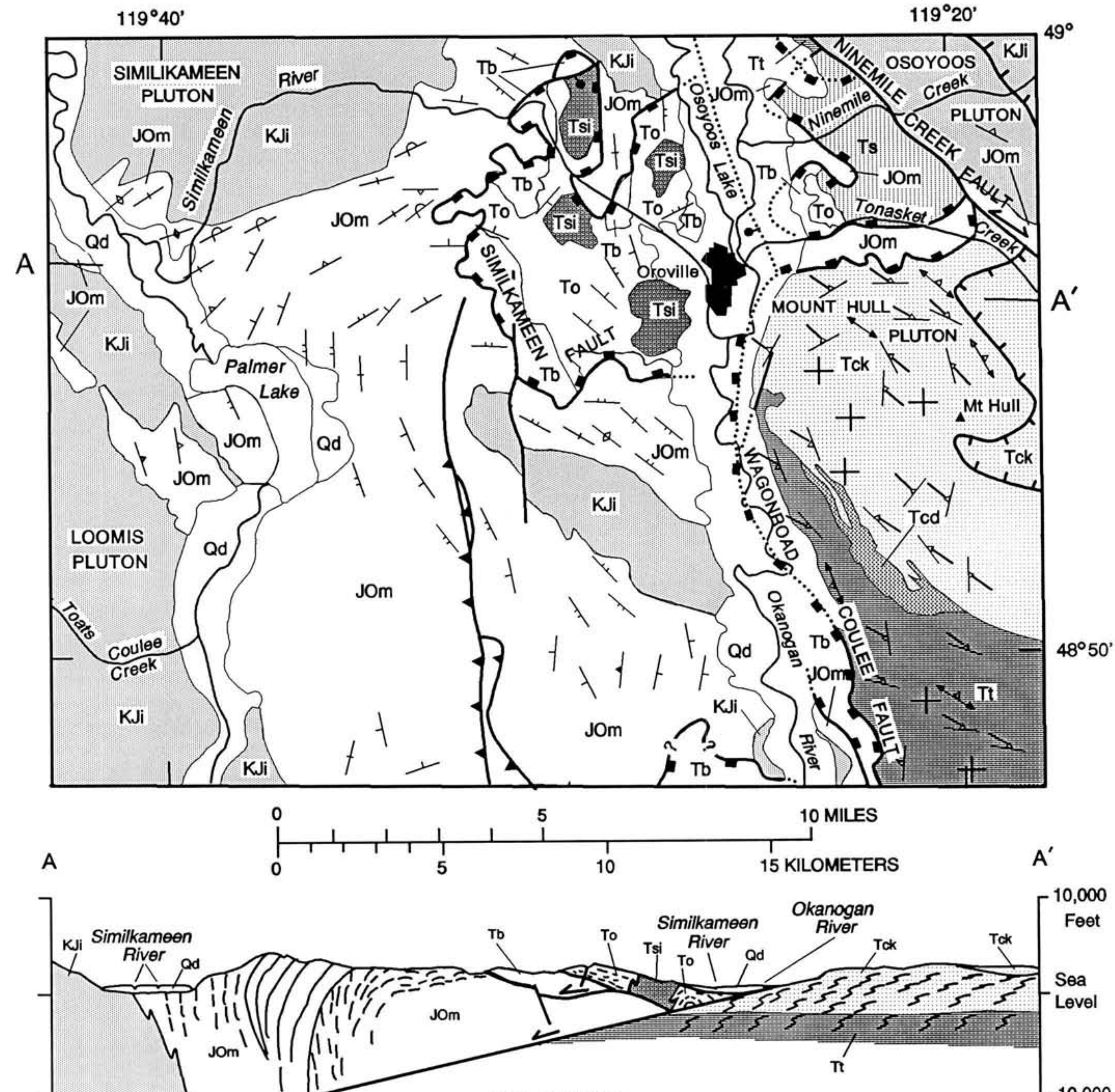
TRd Weakly metamorphosed dolomite, limestone, sandstone, siltstone, and chert conglomerate — includes Cave Mountain and correlative rocks

TRpa Weakly metamorphosed chert conglomerate, argillite, wacke, pyroclastic rocks, and mafic lava — includes Anarchist Group (Bullfrog Mountain and Spectacle Formations) and correlative rocks

Oc Weakly to moderately metamorphosed quartzite, argillite, limestone, pillowed basalt, and K-feldspar-bearing arkose and wacke — includes Covada Group and metamorphic complex of Conconully

DYs Cratonic sedimentary rocks — deposited in a miogeoclinal or continental shelf setting

MzPzg Gneiss and schist



EXPLANATION

Contacts

Sharp contact

Gradational contact — between penetratively foliated and lineated rocks of gneiss domes and less foliated and lineated rocks outside of gneiss domes; ticks toward less foliated rocks

Faults — dotted where concealed; arrows show direction of movement on cross sections

?... Low-angle extensional (detachment) fault — blocks on upper plate; queried where uncertain

Thrust fault — sawteeth on upper plate

Normal fault — ball on downthrown side

Transform(?) fault — high- to low-angle faults subparallel to direction of Eocene crustal extension

Fold

Syncline — dotted where concealed

Bedding — showing dip

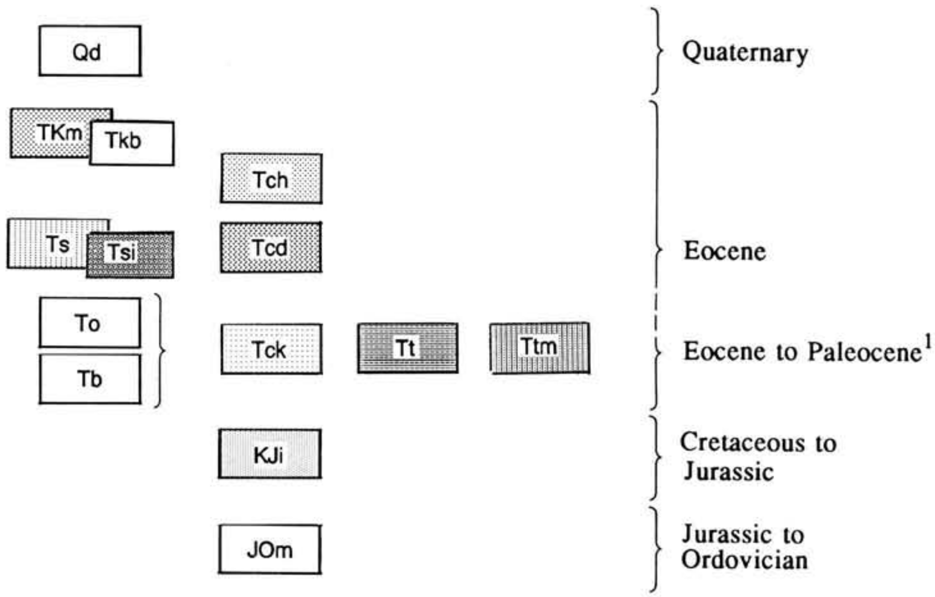
15° - 45° | 45° - 85° | 85° - 90° | Overturned

Foliation — showing dip as well as strike of lineation lying in plane of foliation

0° - 15° | 15° - 45° | 45° - 85° | 85° - 90° | with lineation

Lineation

Inclined | Horizontal



[1] With respect to units Tt and Ttm, refers to age of penetrative dynamic metamorphism. Age of protolith uncertain.

GEOLOGIC UNITS

Qd — Alluvial deposits

Klondike Mountain Formation

TKm — Glassy and lithoidal silicic lava flows — in Republic and Curlew Lake areas includes underlying and interlayered sandstone, mudstone, and volcanic conglomerate and breccia

Tkb — Monolithologic breccia and slide masses (greenstone, granite, marble, serpentine) — in Toroda Creek area includes interlayered conglomerate, sandstone, and shale

Sanpoil Volcanics

Ts — Lithoidal lava flows — typically rhyodacite, but also locally includes quartz latite and andesite

Tsi — Hypabyssal intrusive rocks — typically rhyodacite porphyry; in Toroda Creek area includes trachyte and andesite

To — O'Brien Creek Formation and related rocks (tuffaceous sandstone, tuff, siltstone, shale, and conglomerate) — in Oroville area includes granite conglomerate and conglomerate composed of low-grade metamorphic rocks

Tb — Monolithologic breccia and slide masses (greenstone, granite) — in Oroville area

Colville batholith

Tch — Herron Creek suite — fine- to medium-grained hornblende (pyroxene) biotite granite

Tcd — Devils Elbow suite — mafic fine-grained hornblende (pyroxene) biotite monzodiorite, locally gneissic

Tck — Keller Butte suite — coarse- to fine-grained biotite (muscovite) granite; commonly megacrystic; lineated and gneissose within gneiss domes

Tt — Tonasket Gneiss — biotite (hornblende) orthogneiss, mafic amphibole gneiss and schist, amphibolite, and biotite-sillimanite schist

Ttm — Metamorphic rocks of Tenas Mary Creek and correlative rocks — chiefly orthogneiss, paragneiss, amphibolite, marble, sillimanite quartzite, and sillimanite schist

KJi — Granodiorite, tonalite, quartz monzonite, monzonite, monzodiorite, and syenite

JOm — Weakly metamorphosed argillite, wacke, mafic lava flows, limestone, quartzite, and chert conglomerate — includes Bullfrog Mountain and Spectacle Formations (of the Anarchist Group), Kobau Formation, Covada Group, and correlative rocks

Figure 3. ***(previous page)*** Geologic map and cross section, Oroville area. Modified from Rinehart and Fox (1972) and Fox (1970). Explanation applicable to Figures 3, 4, and 5.

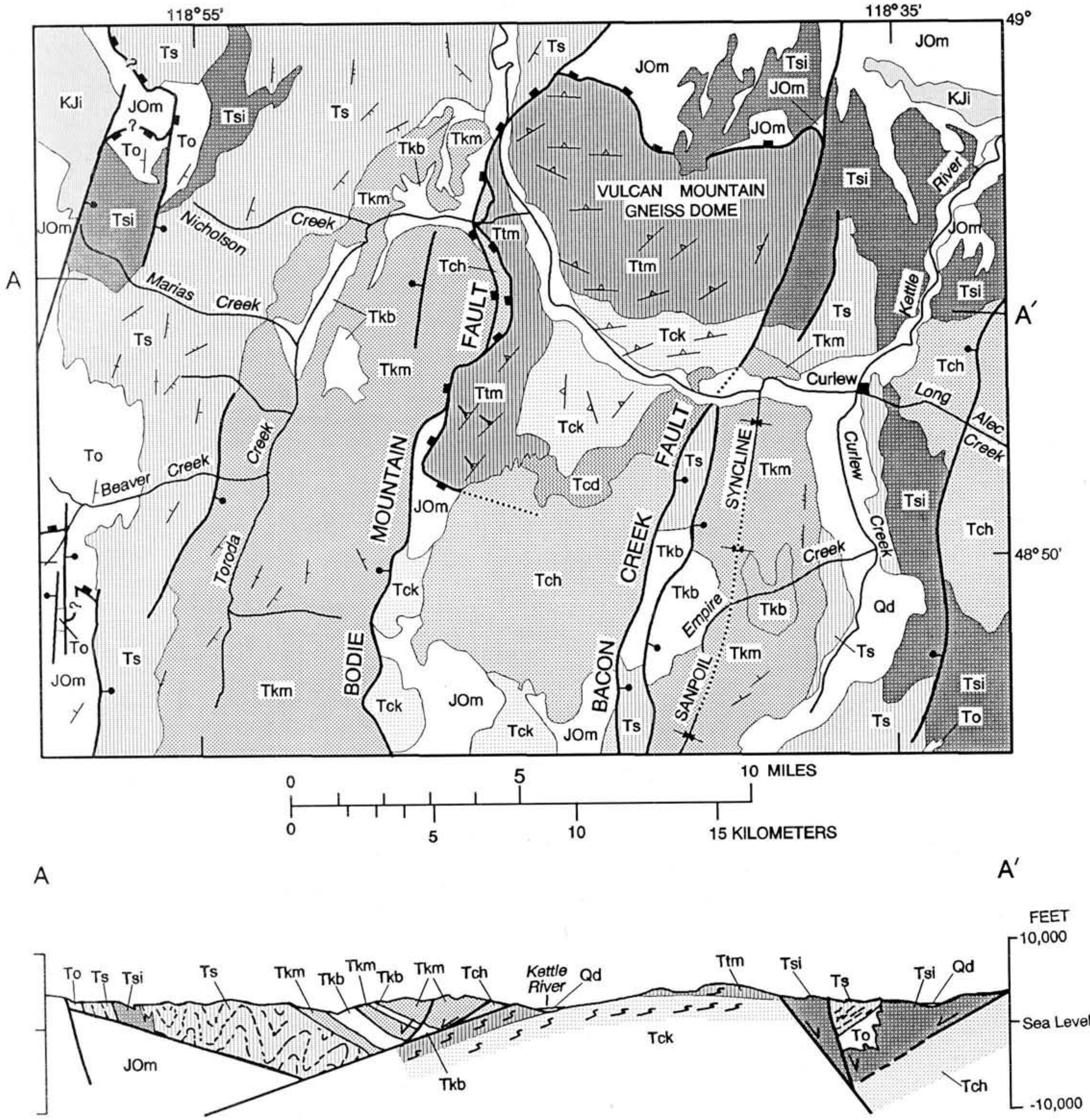
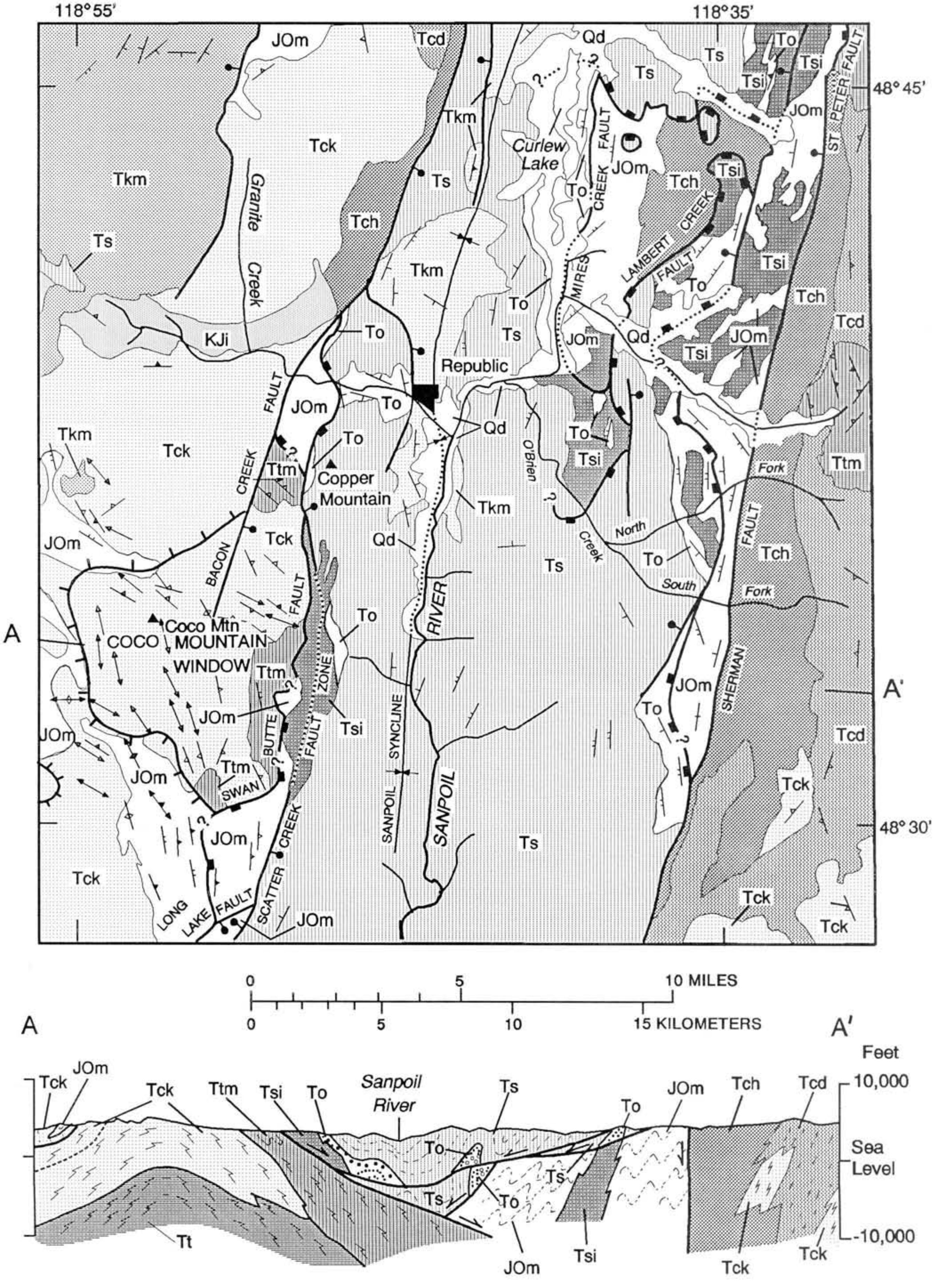


Figure 4. *(above)* Geologic map and cross section, Curlew area. Modified from Pearson (1967), Parker and Calkins (1964), and Holder (1990). Explanation same as that for Figure 3.

Figure 5. *(facing page)* Geologic map and cross section, Republic area. Modified from Muessig (1967), Rinehart and Greene (1988), Parker and Calkins (1964), Staatz (1964), and Holder (1990). Swan Butte fault modified from Orr and Cheney (1987). Explanation same as that for Figure 3.


118°55'
118°35'
48°45'
48°30'
JOm
Tcd
Qd
Ts
To
Tsi
Tkm
Tck
Tch
Curlew Lake
Granite Creek
MIRES CREEK FAULT
LAMBERT CREEK FAULT
ST. PETER FAULT
KJi
Republic
Qd
FAULT
CREEK
BACON
Ttm
Copper Mountain
O'Brien
North Fork
Creek
South Fork
Ttm
Tcd
Coco Mtn
COCO MOUNTAIN WINDOW
SCATTER CREEK FAULT ZONE
BUTTE
SWAN
SANPOIL SYNCLINE
SANPOIL RIVER
SHERMAN FAULT
LONG LAKE FAULT
A
A'
0
5
10 MILES
0
5
10
15 KILOMETERS
Sanpoil River
Feet
10,000
Sea Level
-10,000
Tt

The southern assemblage consists chiefly of weakly to moderately metamorphosed, interlayered graywacke, arkose, quartzite, siltstone, argillite, and basaltic to andesitic lava flows assigned to the Covada Group[1]. The fine-grained clastic rocks typically consist of thick sequences of thinly laminated, graded beds of black argillite and medium- to light-gray siltstone and fine sandstone. These features suggest deposition in part by turbidity currents in a bathyal marine setting. The interlayered lava flows commonly are pillowed, indicating underwater deposition. Other distinctive lithologies include quartzite and wacke. The quartzite is commonly dark bluish gray to black and forms thick mappable layers. The wacke is composed of angular to poorly rounded sand- to coarse-grit-size grains of quartz, sodic plagioclase, potassium feldspar, mica, orthoquartzite, and, rarely, granitic rock. Neither the black quartzite nor K-feldspar-bearing wacke are known to be present in the Anarchist Group or its correlatives, and chert-pebble conglomerate so characteristic of the Anarchist and the Permian and Lower Triassic strata in the Kettle Falls area (Fig. 2) is apparently absent from the Covada (Fox and Rinehart, 1974).

Rocks now correlated with the Covada in the Tonasket area (Fig. 2) were in part previously assigned to the Anarchist (Fox, 1970; Rinehart and Fox, 1972, 1976), but the presence of mappable layers of dark bluish-gray quartzite and K-feldspar-rich arkose indicates that these rocks are more likely part of the Covada.

Fossils from a locality east of Lake Roosevelt indicate that the Covada Group is Early Ordovician in age (Snook and others, 1981, p. 6). Detrital muscovite separated from coarse grits in the Covada at two localities yielded K-Ar ages of 1,060 Ma and 945 Ma (Table 1), the latter a minimum age because of data acquisition problems. Detrital zircons from a locality near Rice yielded $^{207}Pb/^{206}Pb$ apparent ages ranging from 1,762 to 2,549 Ma (Smith and Gehrels, 1991, p. 1278). The Covada was apparently derived, in part, through erosion of a source terrain consisting dominantly of potassium-rich Precambrian granite. No such terrane is known to have been exposed to erosion during the Ordovician in contiguous regions of the craton to the east.

The contact between the Permian to Middle Triassic rocks (chiefly the Anarchist Group and its correlatives) and the Ordovician rocks (Covada Group) is obliterated over most of its inferred extent by the core complexes or covered by Eocene volcanic rocks and Quaternary deposits. The contact is, however, apparently exposed 8 km northwest of Tonasket (Fig. 2). At this locality, locally overturned light-gray limestone of the Spectacle Formation is thrust over contorted dark-gray phyllite assigned to the Covada. Coarser grained clastic rocks of the Covada—arkose and wacke—are commonly penetratively mylonitized within a zone several kilometers in width adjacent to this fault.

Table 1. Analytical data and calculated K-Ar ages of muscovite in samples from the Covada Group (Analyses by J. C. VonEssen, J. Saburomaru, S. Neil, and D. Vivit)

Field no.	K_2O (%)	$^{40}Ar_{rad}$ (moles/g x 10^{-8})	Radiogenic Ar (%)	Calculated age (Ma)
78KF34A1	10.085	2.09545	96.69	1060.37 ± 6.38
78KF37	10.185	1.82027[1]	97.37[1]	944.61 ± 5.68[2]

[1] Minimum value (data channel overflow) [2] Minimum age

Sample locations and descriptions:

78KF34A1: Inchelium 15-minute quadrangle, 48°26.17'N, 118°10.17'W, collected by K. F. Fox, Jr. Sample is detrital muscovite (-60 +140 mesh) separated from arkosic sandstone of the Covada Group. The muscovite was not detectably recrystallized after compaction of the protolithic sediments, although a chlorite-like mineral is present in the matrix. Other samples from this area show incipient recrystallization of muscovite and contain nontronite, suggesting minor alteration.

78KF37: Inchelium 15-minute quadrangle, 48°19.39'N, 118° 05.95' W, collected by K. F. Fox, Jr. Sample is detrital muscovite (-60 +140 mesh) separated from arkosic sandstone of the Covada Group. The sample is similar in all respects to sample 78KF34A1 (described above) except that the muscovite shows incipient recrystallization.

Black to dark-gray argillite, phyllite, and subordinate greenish-gray amphibolite assigned to the metamorphic complex of Conconully are thrust eastward over the Cave Mountain Formation. The argillite lithologically resembles the black argillite forming a major facies of the Covada Group. This similarity, coupled with the presence of the Covada to the north near Tonasket, suggests that the protolith of rocks in the metamorphic complex of Conconully could be a western continuation of the Covada. The argillite and phyllite grade westward and southward to extensive areas of granitoid gneiss and schist, which are intruded by the Mesozoic Okanogan composite batholith (of Daly, 1906) composed of tonalite, trondhjemite, and granodiorite. These crystalline rocks extend westward to the Pasayten fault (Fig. 1). The metamorphic rocks are of similar metamorphic grade as rocks in the core complexes but lack the domal foliation, mylonitization, and penetrative lineation characteristic of the core complexes.

Terrane Boundaries

In Washington, the contact between Quesnellia and the miogeoclinal province lies in a belt of northwest-vergent thrust faults centered approximately on the Columbia River between the 49th parallel and Kettle Falls (Figs. 1 and 2). Lower Paleozoic rocks of the craton (Metaline Formation, Gypsy Quartzite, Ledbetter Formation) are thrust over Permian through Upper Triassic (and Lower(?) Jurassic) strata of Quesnellia. South of Kettle Falls, the thrust zone (Huckleberry Range fault of Snook and others, 1990) stacks Precambrian cratonic rocks over Paleozoic cratonic rocks (Mills, 1985). The northwest-vergent polarity of the thrusts

[1] Near Hunters (Fig. 2), the Covada Group has been divided into two units by Smith and Gehrels (1992), the Daisy sequence, composed of fine- to coarse-grained clastic rocks and minor limestone, and the conformably overlying Butcher Mountain sequence, composed chiefly of basalt, tuff, and limestone.

north of Kettle Falls implies that movement on the Huckleberry Range thrust zone was also northwest vergent (compare Snook and others, 1981, 1990).

Quesnellia could be allochthonous and displaced right-laterally (as proposed, for example, by Fox, 1977) from the cratonic source areas of the first-cycle granitic detritus present in strata of the Covada Group. However, Yates (1973) concluded that the contact between the eugeosynclinal province (that is, Quesnellia) and the miogeoclinal province does not represent a plate junction. This view has been reiterated by Smith and Gehrels (1991, p. 1282), who conclude, on the basis of U-Pb geochronology of zircons from the Covada Group and possible correlative units in British Columbia, that these strata "...did not undergo significant postdepositional northward latitudinal displacement during Mesozoic time."

If allochthonous, Quesnellia was accreted to the pre-Jurassic craton (miogeoclinal province) in the Middle Jurassic (Monger and others, 1982), probably forming the Huckleberry Range thrust zone at that time.

The western contact of Quesnellia in Washington is formed by the Pasayten fault (Fig. 1). North of Twisp, this fault appears to be a high-angle left-lateral strike-slip fault with a linear north-northwest-trending trace (Lawrence, 1978). However, to the south, the fault plane rolls to a moderate eastward dip, and near Twisp, the trace bends sharply westward. This segment of the fault is likely a west-vergent thrust (Menzer, 1983, fig. 16).

Plutonic Rocks

Two composite batholiths are exposed in north-central Washington. The western of the two is the Okanogan composite batholith (Daly, 1906). It underlies the Okanogan Range and consists of Jurassic and Cretaceous plutonic rocks (Fig. 2). The eastern batholith, traditionally referred to as the Colville batholith (Pardee, 1918; Waters and Krauskopf, 1941) forms bedrock in much of the central part of the map area (Fig. 2) and consists of Paleocene and Eocene plutonic rocks. The Colville batholith is both spatially and temporally associated with the core complexes and Eocene volcanic and epiclastic deposits (correlation diagram, Fig. 2).

Okanogan composite batholith

The Okanogan composite batholith consists chiefly of large interlocking or nested calc-alkalic mesozonal plutons composed of tonalite, trondhjemite, and granodiorite. The plutons are locally separated by septa of high-grade metamorphic rocks, but more commonly they are in mutual contact. The plutons range in age from Early Jurassic near Loomis to mid-Cretaceous near Conconully (Menzer, 1970; Rinehart and Fox, 1976). Plutons forming the western margin of the batholith are also mid-Cretaceous (Hurlow and Nelson, 1993). The batholith is locally overlain by Eocene volcanic rocks (Fig. 2).

Early Jurassic to mid-Cretaceous plutons are also sparsely distributed as discrete bodies sharply cutting the low-grade Paleozoic and early Mesozoic basement throughout the remainder of the province east of the Okanogan Range. Eruptive equivalents of the Cretaceous plutonic rocks are absent, presumably removed by pre-Eocene erosion. The concentration of mesozonal plutonic rocks and high-grade metamorphic rocks along the western flank of the province and their exposure prior to the Eocene probably indicates (1) that pre-Eocene erosion cut deeper into the crust in that area than in areas to the east, and (2) that the Okanogan Range was the locus of successive magmatic and volcanic arcs during the Mesozoic.

Colville batholith

The Colville batholith consists of three lithologically, chemically, and probably temporally distinct suites of epizonal plutonic rocks, (from oldest to youngest) the informally named Keller Butte, Devils Elbow, and Herron Creek suites of Holder and Holder (1988) (Fig. 2). A similar classification of rocks in the Colville batholith has been presented by Carlson and Moye (1990). Their leucocratic granite suite is equivalent to the Keller Butte suite, the plutonic rocks included in their Sanpoil Suite are equivalent to the Devils Elbow suite, and their Deadhorse Suite is equivalent to the Herron Creek suite.

Plutons of the Keller Butte suite are the most extensive. They consist chiefly of leucocratic, medium- to coarse-grained granite and granodiorite containing biotite or biotite plus muscovite and accessory allanite, monazite, titanite, apatite, zircon, and garnet. Feldspars include orthoclase (commonly megacrystic) and oligoclase. Color index commonly ranges from 5 to 10. Discounting features related to the overprint of dynamic metamorphism (foliation, mylonitization, and lineation), three textural varieties are present: megacrystic, medium-grained granitic, and seriate. The seriate variety combines, at hand-specimen scale, intergrading fine-grained granitic and pegmatitic textures (Fig. 6), and this texture characterizes bodies of rock as large as several square kilometers within larger masses of megacrystic or medium-grained granitic rock. The Keller Butte suite comprises large bodies of lineated and mylonitic granite and granodiorite in the gneiss domes, as well as partially lineated and mylonitic bodies mantling the gneiss domes.

K-Ar, fission-track, and U-Pb age determinations on various minerals from samples of the Keller Butte suite (including the informally designated Mission Creek gneiss of Carlson and Moye, 1990) range from 66 to 47 Ma. (See compilations by Fox and others, 1977; Atwater and Rinehart, 1984, and Carlson and Moye, 1990.) Taken together, the data suggest intrusion and crystallization of these rocks between 60 and 55 Ma and subsequent cooling through the blocking temperatures of least retentive minerals (biotite and apatite) by 47 Ma. Local reheating prior to 47 Ma cannot be ruled out.

The Devils Elbow suite consists of biotite- and hornblende-bearing diorite, quartz monzodiorite, and granodiorite (Holder and Holder, 1988, p. 1972). These rocks are finer grained and distinctly more mafic (color index typically 20–25) than rocks of the Keller Butte suite.

Figure 6. Seriate texture common in plutonic rocks of the Keller Butte suite of Holder and Holder (1988).

Rocks of the Devils Elbow suite commonly crosscut the paragneiss of the gneiss domes and the rocks of the Keller Butte suite. Locally, contact relations are more complex. For example, inclusions in the Swimptkin Creek pluton (Fig. 2) include blocks of lineated augen gneiss similar to that composing the nearby Okanogan gneiss dome, along with blocks of graphitic hornfels, quartzite, and tactite presumably stoped down from the low-grade metamorphic rock forming the roof of the dome (Fox and others, 1976, p. 1221). In places, the Swimptkin interfingers with the rocks of the dome and is penetratively foliated and lineated (Singer, 1984). Attitudes of foliation and lineation conform with those of the gneiss dome.

K-Ar age determinations of coexisting biotite and hornblende from the Swimptkin Creek pluton are concordant at 49.2 ± 0.5 and 49.4 ± 1.5 Ma, respectively (Fox and others, 1976, p. 1220). Coexisting biotite and hornblende from another body in the suite, the Devils Elbow pluton, are slightly discordant at 45.1 ± 1.1 and 47.7 ± 1.2 Ma (Atwater and Rinehart, 1984). Coexisting hornblende and biotite from a pluton 5.8 km west of Curlew have K-Ar ages of 55.1 ± 2.7 and 53.5 1.7 Ma, respectively (sample A14, Pearson and Obradovich, 1977, p. 40). K-Ar ages of biotite from other samples of these plutons range from 49.5 ± 1.2 to 49.6 ± 1.2 Ma (Fox and others, 1976, p. 1220; Atwater and Rinehart, 1984).

The Herron Creek suite consists of two phases, an older phase composed of medium- to coarse-grained, gray to pinkish-gray hornblende-biotite quartz monzonite to monzogranite, and a younger phase composed of fine-grained, gray to brown hornblende-biotite monzogranite (Holder and Holder, 1988, p. 1973). These rocks intrude rocks of the Keller Butte and Devils Elbow suites. They also intrude extensive swarms of rhyodacite dikes coeval with the Sanpoil Volcanics (Parker and Calkins, 1964), as well as the paragneisses of the Kettle dome, and are themselves cut by boundary faults of the Republic graben. Pearson and Obradovich (1977, p. 40) reported two K-Ar age determinations of the unit as follows: Biotite from a sample of a pluton east of Republic has an age of 52.4 ± 1.6 Ma, and biotite from a sample of the Long Alec Creek pluton east of Curlew, 53.0 ± 1.6 Ma. Actual ages of these plutons are likely near the low end of the reported range because the plutons also cut rocks coeval with the Sanpoil Volcanics, thought to be no older than 52 Ma.

The alkali-lime index (Pecock, 1931) of the Keller Butte suite is calcic, the Devils Elbow suite calc-alkalic, and the Herron Creek suite alkalic (Holder and Holder, 1988, p. 1979). Comparison of normative compositions of late-stage aplites and leucogranites with experimental data indicates that the Keller Butte and Herron Creek suites completed crystallization at pressures of 500–1,200 bars (2–3 km below the surface) (Holder and Holder, 1988, p. 1974).

On the basis of chemical and strontium isotope data, Carlson and others (1991, p. 13,332) suggest that the Keller Butte suite was derived through small degrees of partial melting of lower crust. Holder and Holder (1988, p. 1978) and Carlson and others (1991) agree that rocks in the Devils Elbow suite include a direct contribution from the upper mantle. Holder and Knaack (1988) suggest that the Herron Creek suite also includes a direct contribution from the upper mantle. Holder (1986) attributes the higher K_2O content and lower alkali-lime index of the suite (compared to those of the Devils Elbow suite) to a smaller degree of partial melting and a greater depth of segregation of magma from the residue.

No plutonic rocks younger than Eocene have been recognized in the Okanogan area (Fig. 2).

Eocene Volcanic Rocks, Dikes, and Associated Pyroclastic and Epiclastic Deposits

Eocene epiclastic, pyroclastic, and volcanic rocks unconformably or structurally overlie the Paleozoic and Mesozoic basement rocks in the province. The Eocene rocks are present as thick stratiform sequences filling structural depressions, including the Toroda Creek, Republic, Keller, and First Thought grabens and the Kettle Falls and Spokane–Enterprise half-grabens. Eocene strata are also present as erosional outliers in the valley of the Okanogan River and cap summit areas in the Okanogan Range to the west (Fig. 2).

Eocene rocks in the Republic and Toroda Creek grabens

The Eocene sequence is divisible into three units: the O'Brien Creek Formation at the base, Sanpoil Volcanics in the middle, and Klondike Mountain Formation at the top (Muessig, 1962; Pearson and Obradovich, 1977) (Figs. 4 and 5). The O'Brien Creek consists chiefly of light-gray, pale greenish-gray and pale brownish-gray sandstone, siltstone, and conglomerate. In addition, the O'Brien Creek includes crystal lithic tuff that, unlike tuffs higher in the section, contains embayed grains of pyroclastic quartz and dark-gray chips of argillite (Pearson and Obradovich, 1977, p. 4). Biotite rhyolite lava forms the basal part of the formation in the southern part of the Republic and Keller grabens (Moye, 1984).

The Sanpoil Volcanics consists chiefly of medium-gray to brown hornblende-pyroxene-biotite lithoidal rhyodacite and quartz latite lava flows. Moye (1982) suggested that the volume of these rocks in the Republic graben exceeds 800 km^3.

The lower part of the Klondike Mountain Formation consists of monolithologic breccia, volcanic breccia, conglomerate, fine tuff, sandstone, siltstone, mudstone, laharic deposits (tuff breccia), and lava flows. In the Republic area, the basal part of this assemblage forms a mappable unit (Tom Thumb Tuff Member) consisting of a coarse fluvial basal conglomerate and overlying interlayered tuffaceous lacustrine siltstone, mudstone, sandstone and several lava flows (Muessig, 1967; Holder and others, 1989, p. 198).

Muessig's (1967) definition of the Tom Thumb Tuff Member and assignment of this unit to the Klondike Mountain Formation rather than the underlying Sanpoil Volcanics has been questioned by Tschauder (1989). Muessig (1967) inferred that the Tom Thumb is bounded both above and below by angular unconformities. However, Tschauder notes (1989, p. 242) that the basal contact as revealed through penetration by numerous drill holes is gradational rather than unconformable and that rocks comprising the Tom Thumb are lithologically similar to rocks elsewhere included in the Sanpoil. Tschauder (1989), therefore, does not use the name "Tom Thumb" for the purposes of his discussion, but rather includes the rocks in that unit in the Sanpoil Volcanics in his report.

Holder and others (1989) note the presence of detritus of the Sanpoil Volcanics in the basal conglomerates of the Klondike Mountain Formation (basal part of the Tom Thumb Member as defined by Muessig, 1967), but they infer that the contact with the underlying Sanpoil is conformable. The invasive basalt flow near the base of the Tom Thumb at Republic (unit Tbi of Muessig, 1967, pl. 1) is paleomagnetically reversed (sample K11, Fox and Beck, 1985, p. 330), perhaps indicating its temporal equivalence with the Klondike Mountain Formation rather than the Sanpoil Volcanics.

The lower part of the Klondike Mountain Formation, composed chiefly of lacustrine sedimentary deposits, is unconformably overlain by the upper part of the formation, composed chiefly of thick, glassy black to light bluish-gray rhyodacite lava flows and thin, brown, columnar-jointed rhyodacite lava flows. Aggregate thickness of the entire Klondike Mountain Formation in the Republic area is estimated to be approximately 1,000 m (D. R. Gaylord and S. M. Price, cited by Holder and others, 1989, p. 198).

Some distinctions among the sections in the Toroda Creek, Republic, and Keller grabens are worthy of note. The basal part of the Sanpoil Volcanics in the Toroda Creek graben consists of pyroxene trachyte flows. These silica-undersaturated rocks thicken to the north in the Greenwood area of British Columbia (Fig. 1), where they include rhomb porphyry and analcite lava (Daly, 1912, p. 80) and sodic trachytes and phonolites (Monger, 1968, p. 14). The basal part of the Klondike Mountain Formation in the Toroda Creek graben contains several thick layers of monolithologic breccia intercalated with lake beds (Pearson, 1967) (Fig. 4). Individual layers of the breccia consist variously of broken or brecciated greenstone, serpentine, and granitic rocks. As noted by Pearson and Obradovich (1977, p. 27), lithologically similar rocks are present at or near the Sanpoil–Klondike Mountain contact in the Republic graben. Monolithologic breccia is present only in the basal part of the Eocene section in the Oroville area (Fig. 3) and is apparently absent from the Keller, First Thought, Kettle Falls, and Spokane–Enterprise structural basins.

The basal 1.3 km of the Sanpoil Volcanics in the Keller graben consists of a dozen or more fairly thin hornblende-biotite rhyodacitic lithoidal lava flows (Fig. 7). The megascopic and microscopic texture of the rhyodacite is conspicuously eutaxitic, indicating that these now devitrified lavas formed as zonally compacted and welded ash flows. Ash flows are rare elsewhere in the region. A 10-m-thick eutaxitic flow is present near the base of the sequence of rhyodacitic lava flows on Whitestone Mountain near Tonasket (Fig. 2). However, extensive development of ash flows appears to be unique to the Keller graben.

Dikes

The O'Brien Creek Formation and Sanpoil Volcanics in the Toroda Creek, Republic, and First Thought grabens are cut by a plethora of rhyodacite and quartz latite dikes (Muessig, 1962). The dikes generally trend north-northeast, parallel to the trend of the grabens. The dikes are locally intimately mixed or gradational with and of similar composition to the Sanpoil; hence they are thought to be its intrusive equivalents (Muessig, 1967, p. 57). They do not cut the overlying Klondike Mountain Formation (Pearson and Obradovich, 1977, p. 16).

The rhyodacite and quartz latite dikes and an older set of biotite rhyolite dikes cut the core complexes and the plutonic and low-grade metamorphic rocks outside the core complexes (Moye and others, 1982, p. 218; Holder and Holder, 1988, p. 1976). The biotite rhyolite dikes are chemically similar to the tuffs in the O'Brien Creek Formation and, like them, contain argillite chips and embayed grains of quartz, thus are likely their intrusive equivalents (Moye,

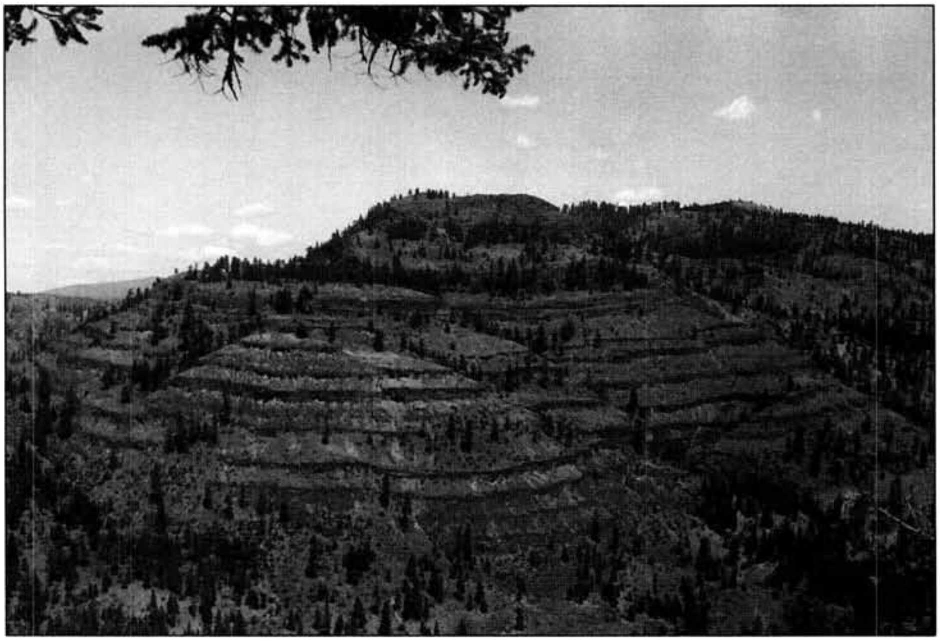


Figure 7. View north showing multiple layers of welded ash-flow tuff in Sanpoil Volcanics at the north end of the Keller graben.

1984). The rhyodacite and biotite rhyolite dikes are particularly abundant in a 5- to 10-km-wide zone adjoining the east side of the Republic graben where they cut plutonic rocks of the Keller Butte suite and high-grade metamorphic rocks of the Kettle dome (shown as "dike complex" in plutonic rocks of the Devils Elbow suite on Fig. 2). In this zone, the dikes commonly constitute more than 50 percent of the bedrock. Moye (1984) estimates that 10 km of extension is needed to accommodate the dikes at the southern end of the Republic graben. The dikes are also abundant south-southwest of and on strike with the Republic graben south of the Columbia River.

Age of Eocene Volcanic and Epiclastic Rocks

K-Ar ages reported by Pearson and Obradovich (1977, p. 40) suggest eruption of the tuffs in the O'Brien Creek Formation at about 54.5 ± 1.5 Ma (one date), the extrusive rocks of the Sanpoil Volcanics at between 53.8 ± 1.8 and 49.6 ± 3.0 Ma (nine mineral separates from six samples), and the Klondike Mountain Formation at between 50.4 ± 1.7 and 47.5 ± 1.7 Ma (two samples). Paleomagnetic data, summarized by Fox and Beck (1985, p. 329), indicate that the Klondike Mountain is reversely polarized (12 sites), the Sanpoil is normally polarized (85 sites), the O'Brien Creek is of mixed polarity (3 sites), and the undivided O'Brien Creek and Sanpoil(?) is normally polarized (two sites). Absence of magnetic reversals in the Sanpoil suggests that it was erupted in its entirety during a single magnetic polarity event. The K-Ar data indicate that this event was most likely anomaly 21, which spanned approximately 1.7 m.y. at between 52 and 50.3 Ma (paleomagnetic time scale of Mankinen and Dalrymple, 1979).

Eocene rocks in the Oroville area

At Oroville (Fig. 3), the Tertiary strata consist chiefly of a conglomerate unit intertonguing with and overlain by sandstone, siltstone, and tuffaceous sandstone. The conglomerate includes a basal part containing cobbles derived from the Anarchist Group and Kobau Formation and an upper part containing cobbles of granitoid rock derived chiefly from the Similkameen composite pluton (Rinehart and Fox, 1972). The lower and upper parts consist of numerous conglomerate layers, each several meters thick, separated by thin layers of sandstone and siltstone. Textures and contact relations of the conglomerate layers indicate their deposition as debris flows. The sandstone and siltstone are locally fossiliferous, containing plant fossils of probable Eocene age (J. A. Wolfe, written commun., cited by Rinehart and Fox, 1972, p. 61). The epiclastic rocks are intruded by rhyodacitic bodies dated (K-Ar) at 52.7 ± 2.6 and 53.4 ± 2.3 Ma (Rinehart and Fox, 1972, p. 62). The conglomeratic basal part of the section could be older than Eocene. It lithologically resembles the Pipestone Canyon Formation of Barksdale (1948) near Winthrop, from which pollen of likely Paleocene age has been recovered. (See Barksdale, 1975, p. 51.)

South of Oroville, the rhyodacite has been extensively fractured and zeolitized. Typical mineral assemblages include laumontite and albite as replacements of primary calcic plagioclase, indicating metamorphism in the laumontite-prehnite-quartz facies (Winkler, 1967, p. 154).

Eocene rocks in the Tonasket area

Near Tonasket (Fig. 2), the Tertiary section consists of very light gray bedded tuff and tuffaceous sandstone at the base, overlain by light greenish-gray crystal lithic tuff, and that, in turn, by light-gray to maroon columnar-jointed lava flows, and, at the top, by conglomerate containing clasts of volcanic rock. Like the rhyodacite south of Oroville, the lava flows are extensively zeolitized. A sample of the relatively unaltered basal lava flow from the northwest side of Whitestone Mountain was dated at 50.2 ± 1.8 Ma (Rinehart and Fox, 1972, p. 62), suggesting its temporal equivalence with the Sanpoil Volcanics. Tertiary volcanic rocks to the south (west of Wagonroad Coulee and near Omak) consist chiefly of altered dacite lava flows and minor amounts of pyroclastic rocks and conglomerate (Rinehart and Fox, 1976; Sims, 1984).

Eocene rocks in the Orient area

The tripartite division of Eocene strata defined in the central part of the map area may also apply to the Tertiary section in the First Thought graben (Fig. 2). The section there consists, in ascending order, of sandstone assigned to the O'Brien Creek Formation, lithoidal biotite-hornblende and biotite-hornblende pyroxene lavas assigned to the Sanpoil Volcanics, and lake beds and thick, black, glassy rhyodacite flows assigned to the Klondike Mountain Formation (Pearson and Obradovich, 1977). The lithoidal lava flows are magnetically normal (sites S36 and S37, Fox and Beck, 1985, p. 331), supporting correlation of these rocks with the Sanpoil.

Eocene rocks in the Kettle Falls area

South of Kettle Falls (Fig. 2), the Tertiary rocks form a homoclinal sequence of conglomerate and sandstone dipping 50–30 degrees south to west-southwest (Mills, 1985). The base of the sequence is not exposed, but the lowest unit is a 210-m-thick conglomerate containing clasts of quartzite, argillite, and chert-pebble conglomerate (Mills, 1985, p. 4), evidently derived through erosion of cratonic lower Paleozoic strata exposed to the south and of Quesnellian upper Paleozoic to Lower Triassic strata exposed to the north. The conglomerate is overlain by 150 m of sandstone, which is intruded or overlain by dacite breccia (Mills, 1985, p. 4). Biotite and hornblende from a sample of the dacite breccia were dated (K-Ar) at 51.7 ± 1.3 and 52.3 ± 1.8 Ma, respectively (sample A7, Pearson and Obradovich, 1977, p. 40). The dacite breccia cuts or is overlain by 275–300 m of bedded volcanic breccia, and that is overlain in turn by 1,500 m of volcanic conglomerate with interbeds of finer grained clastic rocks, including pebbly sandstone, tuffaceous sandstone, and siltstone (Mills, 1985, p. 4).

The lowest conglomerate and overlying sandstone were assigned to the O'Brien Creek Formation, and the dacite breccia and volcanic conglomerate, sandstone, and siltstone to the Sanpoil Volcanics (Pearson and Obradovich, 1977). However, the beds assigned to the O'Brien Creek differ from the O'Brien Creek in the Republic, Toroda Creek, and Keller grabens in that they contain cobbles of cratonic derivation. Rocks near Kettle Falls assigned to the Sanpoil differ from the Sanpoil elsewhere in that they are chiefly sedimentary.

METAMORPHIC CORE COMPLEXES

The Okanogan and Kettle gneiss domes are large bodies of dynamically metamorphosed rock elongated north-south, lying side by side, with the much smaller Vulcan Mountain dome sandwiched between them (Fig. 2). The Okanogan dome is approximately 75 km long and 55 km wide, and the Kettle dome (including the part in British Columbia) is 115 km long and 30 km wide. The eastern flank of the Okanogan dome is exposed in the Coco Mountain window.

The three domes are composed of medium- to high-grade metamorphic rocks and syntectonic granite of the Keller Butte suite. The metamorphic rocks are lineated and foliated throughout. The enveloping and locally infolded granite is a penetratively lineated and foliated orthogneiss in zones several to many kilometers wide adjacent to contacts with metamorphic rocks in the interior of the domes. Foliation and lineation gradually diminish away from these zones as the orthogneiss grades to nearly structureless granite that forms zones several kilometers wide adjacent to contacts with country rocks outside the domes.

The metamorphic rocks include two contrasting assemblages: (1) aluminous metaclastic and metavolcanic rocks referred to as the metamorphic rocks of Tenas Mary Creek (Parker and Calkins, 1964) and (2) layered mafic and felsic gneiss and schist of the Tonasket Gneiss (Snook, 1965; Fox and others, 1976). The Tenas Mary Creek rocks of the Kettle gneiss dome are the southern extension of the Grand Forks Group of Preto (1970). The Tonasket and Tenas Mary Creek units include bodies of hornblende-biotite granodioritic gneiss. The aluminous assemblage crops out in the Kettle and Vulcan Mountain domes and at the eastern margin of the Okanogan gneiss dome. The mafic and felsic gneiss and schist assemblage is restricted to the central and western parts of the Okanogan gneiss dome.

The metamorphic rocks of Tenas Mary Creek consist of biotite-muscovite-sillimanite schist, diopside marble, amphibolite, orthogneiss, and orthoquartzite containing laminae of muscovite-sillimanite schist. Individual layers are typically several meters to 100 m thick; some are continuous for as much as 20 km. The layers are locally folded into recumbent isoclines, along with myriad, much-transposed, near-parallel but cross-cutting dikes and interlayered sill-like bodies of leucocratic muscovite granite gneiss and gneissic pegmatite.

The Tonasket Gneiss consists of amphibolite, amphibolitic gneiss, amphibole (biotite + sillimanite) schist, leucocratic gneiss, and biotite-hornblende granodioritic gneiss. Layering in the unit, defined chiefly by variation in color index, is at two scales. Individual layers that reach 10 m in thickness are themselves thinly laminated to thinly layered.

Metamorphic Grade

Mineral assemblages in the amphibolitic core at the deepest structural levels exposed in the Okanogan gneiss dome include anthophyllite, labradorite-andesine, muscovite, and (or) phlogopite, spinel, and apatite. Felsic laminae a half to 3 cm thick are composed of andesine, orthoclase, and gedrite. Other felsic laminae are composed of quartz, andesine, orthoclase, muscovite, and phlogopite, or quartz, andesine, muscovite, phlogopite, sillimanite, and garnet (Fox and Rinehart, 1988, p. 9). Corundum (sapphire), spinel and, locally, olivine are also present (Snook, 1965), as is the assemblage sillimanite-garnet-cordierite. These assemblages suggest equilibration of core-zone rocks in the granulite facies (Snook, 1965), followed by retrogression to middle amphibolite facies, at pressures between those of the Abukuma and Barrovian types (Fox and Rinehart, 1988, p. 10). The amphibolitic rocks are commonly altered or partially altered. In addition to relict labradorite, they are

Figure 8. View to the north showing homoclinally west-dipping mylonitic gneiss forming the western flank of the Okanogan gneiss dome. The south-flowing Okanogan River cuts through the nose of the west-northwest-plunging antiform in foliation in the center of this view. To the west (left), the low-angle, west-dipping normal (detachment) fault bounding the west side of the gneiss dome is concealed beneath surficial deposits in Wagonroad Coulee, an abandoned meltwater channel of the Okanogan River. The fault surface approximately parallels foliation in the gneiss dome.

composed of actinolitic amphibole, talc, chlorite, clinozoisite, and zeolite minerals (Fox and Rinehart, 1988, p. 9).

Pelitic schists and quartzite in the aluminous sedimentary and metavolcanic rocks of the Tenas Mary Creek unit contain garnet, sillimanite, biotite, and muscovite at the deepest structural levels exposed, estimated to be 3 km by Cheney (1980, p. 473) in the central part of the Kettle dome and 1.3 km by Fox and Wilson (1989) in the southern part of the Kettle dome. Cordierite with sillimanite (or andalusite instead of sillimanite) are present at higher structural levels. Kyanite is present in lieu of sillimanite in the southern part of the dome. Marble typically contains diopside and, locally (east of Togo Mountain), chondrodite. Corundum and spinel are present west of Laurier (Rhodes, 1980). At low structural levels, the rocks of the Vulcan Mountain and Kettle gneiss domes apparently equilibrated in the intermediate part of the Abukuma-type amphibolite facies (as defined by Winkler, 1967, p. 121).

Internal Fabric

The most pervasive elements of the fabric of the rocks in the core complexes are: (1) foliation, (2) lineation, and (3) folds. Foliation is nearly horizontal in the central part of the complexes and steepens to 10–30 degrees near the margins (Fig. 8). The structure of the complexes is thus broadly domal. Compositional layering is grossly parallel to the foliation. Structure in the interior of the Okanogan dome is somewhat more complex—the foliation defines three culminations with intervening saddles (Fox and Rinehart, 1988, p. 3).

The lineation is defined by subparallel arrangement of elongate crystals (amphibole, mica, sillimanite), trains of minerals (feldspar, apatite), and ribbons of quartz. The lineation is nearly unidirectional, generally trending N55–60°W except for an anomalous deviant zone along the long axis of the Kettle dome (Fig. 9). The foliated and lineated rocks are pervasively mylonitized (Fig. 10). Rotation and granulation of crystals and transposition of cross-cutting dikes along foliation planes in the mylonitic rocks indicate that the foliation is a fluxion structure, formed through flowage-like movement and flattening of the then-hot and ductile rocks in the core complexes.

Rocks in the complexes are extensively folded. At least two generations of folds are commonly distinguishable. The earlier includes recumbent isoclines, some of which are intrafolial, with amplitudes ranging from microscopic to several hundred meters (Figs. 11, 12). These folds are ubiquitous in the foliated and lineated parts of the complexes. Their fold axes are typically parallel or subparallel to the accompanying lineation. However, in places the trend of fold axes shows wide scatter (for example, Wilson, 1981, p. 68–72), particularly in the marbles and amphibolites at the eastern margin of the Kettle gneiss dome.

At high structural levels in the gneiss domes, second generation open folds with upright axial planes are superimposed on the earlier recumbent isoclines. Trend of fold axes of the younger set is typically approximately north, but scatter is much greater than that of the earlier set.

Protolith

The protolith of the metamorphic rocks of Tenas Mary Creek unit was evidently a thickly layered sequence of argillite, sandstone, limestone, and basalt flows and hornblende-biotite granitic plutons intrusive into them. The age of this material is not known, but the sedimentary and volcanic constituents are compositionally similar to rocks of the Anarchist, Covada, and Rossland Groups outside the complexes. Detrital zircons from a pure quartzite collected at a locality in the Grand Forks Group in Canada yielded $^{207}Pb/^{206}Pb$ ages ranging from 570 to 3,319 Ma (Ross and Parrish, 1991, p. 1263). As noted by Ross and Parrish (1991), the analyses were markedly discordant, suggesting Pb loss or crystal overgrowth during metamorphism. Nevertheless, they indicate that the sedimentary protolith of the Grand Forks Group (and its correlative, the metamorphic rocks of Tenas Mary Creek) is composed at least in part of Precambrian detritus.

The protolith of metamorphic rocks in the central and western parts of the Okanogan gneiss dome could have been interlayered basalt and tuffaceous (basaltic) sediments cut by hornblende-biotite granodiorite, diorite, and granite. The age of the protolith of paragneiss and orthogneiss in the central part of the Okanogan dome is probably at least in part approximately 87 to 85 Ma, judging from the $^{206}Pb/^{238}U$ ages of zircon (Fox and others, 1976, p. 1223; Potter and others, 1991, p. 90).

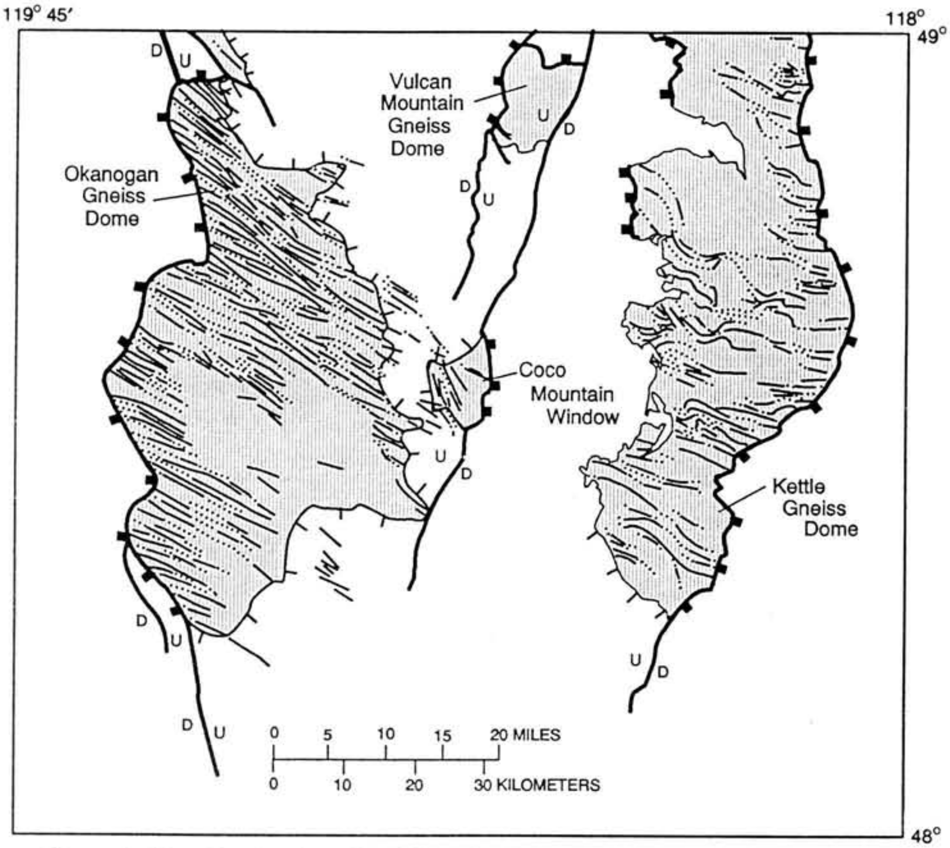


Figure 9. Map showing the strike of lineation in gneiss domes in north-central Washington. Dotted traces are inferred. Gneiss domes are bounded by low-angle normal (detachment) faults (blocks on upper plate), gradational contacts (ticks on upper plate), and contacts with younger plutonic rocks. High-angle normal faults (U, up; D, down) splay away from the detachment faults.

STRUCTURES BOUNDING THE CORE COMPLEXES

Western Contact

The western contact of the Okanogan gneiss dome is a west-dipping detachment fault (here referred to as the Wagonroad Coulee fault) juxtaposing mylonitic rocks of the gneiss dome against an upper plate composed of Mesozoic plutonic rocks, Ordovician to Upper Triassic low-grade metamorphic rocks, and Tertiary epiclastic and volcanic rocks (Fig. 8). Although the fault is largely concealed beneath alluvium and glacial drift, its location is closely bracketed by outcrops. Regionally, the fault cuts across compositional units in the gneiss dome (Fig. 2), but on a local scale the strike of the fault parallels that of foliation in the mylonite. Assuming that the respective dips are also parallel, the fault must dip 5–30 degrees to the west. East of Tonasket the dip shallows, and the fault is physically represented by a zone of mylonitic to ultra-mylonitic orthoclase-bearing limy wacke of the upper plate, of which a 10-m thickness is exposed. Outliers of this material caked on the upper surface of the gneiss dome were mapped as roof remnants by Waters and Krauskopf (1941). The fault undulates from horizontal to 15 degrees west dip and projects below nearby greenstone breccia and Tertiary sedimentary and volcanic strata. The Tertiary rocks dip east 20–30 degrees into the fault.

The Wagonroad Coulee fault curves eastward around both the northwest and southwest flanks of the gneiss dome. At the southwest flank, 3 km northwest of Omak Lake, the fault zone consists of 50 m of chloritic breccia in which clasts of plutonic rock are mixed with high-grade metamorphic rock.

Two faults—Crystal Springs and Goose Flats—splay away from the Wagonroad Coulee detachment into the upper plate along this margin (Fig. 2). The curvilinear trace of the Crystal Springs fault is subparallel to that of the Wagonroad Coulee detachment. The Crystal Springs fault is therefore inferred to be a low-angle normal fault dipping shallowly to the southeast. The Goose Flats fault is a major fault, cleanly separating migmatitic layered gneiss on the east from a complex assemblage of schist, gneiss, and plutonic rocks on the west, but its dip and direction of displacement are unknown.

Along the northwest flank of the Okanogan gneiss dome, the Wagonroad Coulee detachment curves eastward, separating a lower plate consisting of the penetratively foliated, mylonitized, and lineated granite of Mount Hull from middle and upper plates consisting, respectively, of low-grade metamorphic rocks of the Anarchist Group and the superjacent Tertiary basin fill (Fig. 3). A layer of marble belonging to the Anarchist is plastered against the lineated granite, conforming both to the detachment surface and to foliation in the granite.

The upper plate, consisting of monolithologic greenstone breccia, granitic conglomerate, and epiclastic deposits along with slivers of the floor on which it was deposited, is tectonically detached from the subjacent middle plate composed of rocks of the Anarchist Group and Kobau For-

Figure 10. Mylonitic augen gneiss, the characteristic texture of plutonic rocks of the Keller Butte suite of Holder and Holder (1988) in the gneiss domes. The coin is 18 mm in diameter.

mation by an undulating sub-horizontal fault, here referred to as the Similkameen fault (Fig. 3). The slivers consist of mafic alkalic complexes (Fox, 1973, 1977) and weakly metamorphosed mafic lava of the Ellemeham Formation (Rinehart and Fox, 1972).

The Similkameen fault is exposed in roadcuts along Tonasket Creek, 3 km east of Oroville, and in the canyon of the Similkameen River, 3 km west of Oroville. At Tonasket Creek, brecciated alkalic rocks (malignite and nepheline syenite) and detritus derived from the alkalic rocks are faulted over gypsiferous shale and fine-grained marble. The fault is sharp but undulatory. Crumpled black to bluish-gray, incoherent gypsiferous shale of the middle plate abuts 3 cm of reddish-gray gouge, which grades upward into reddish-gray to greenish-gray brecciated malignite of the upper plate. Upper plate rocks extend southward as outliers overlying marble of the middle plate, and one small outlier rests directly on the mylonitic rocks of the gneiss dome.

In the canyon of the Similkameen River, the fault is a planar zone of gouge 15–30 cm thick, dipping 12 degrees west and separating granite conglomerate of the upper plate from fractured green phyllite of the Kobau Formation below (Fig. 13). High-angle fractures and microfaults cutting upper plate rocks curve to shallow westerly dips and merge with the fault in a 15- to 40-cm-thick zone of brecciated granite conglomerate that forms the basal part of the upper plate. The geometry of this intersection indicates that upper plate rocks at this locality likely moved west relative to the footwall.

Figure 11. Isoclinal fold, eastern flank of Kettle gneiss dome. The interval between graduations on the pole is 10 cm.

West of Osoyoos Lake, Tertiary conglomerate in the upper plate dips 30–45 degrees to the east (Fig. 3). Reappearance of the fault and basal upper plate rocks east of the lake indicates the presence of a north-trending, west-dipping normal fault concealed beneath the waters of Osoyoos Lake. This fault cuts the upper plate, displacing the west side down possibly as much as 2 km. This fault probably continues southward in the subsurface to a concealed intersection with the Wagonroad Coulee detachment fault.

Eastern Contact

The penetratively lineated and foliated mylonitic rocks forming the eastern margin of the Kettle dome (Campbell, 1938) are bounded on the east side by a low-angle, east-dipping detachment fault, referred to as the Kettle River fault (Rhodes and Cheney, 1981, p. 366) (Fig. 2). The fault is subparallel to local foliation and compositional layering in the mylonites, but regionally it cuts obliquely across the major compositional units (quartzite, marble, amphibolite, and augen gneiss), forming the eastern margin of the gneiss dome. The down-dip subsurface extension of the mylonites between 6 and 10 km east of the dome was seismically imaged at a depth of 1.5 to 2 km, suggesting that in this area the Kettle River fault dips approximately 11 degrees east (Hurich and others, 1985, Fig. 2).

The fault typically juxtaposes low-grade metamorphic rocks of the upper plate against sillimanite-grade rocks of the lower plate. At exposures 1 km east of the Kettle River and 6 km southeast of Orient, the fault is a zone several meters thick of lineated, greenish to brownish-gray chloritic microbreccia containing folded laminae of brecciated pegmatite. The zone overlies intensely fractured wedges of mylonitic and locally folded white quartzite containing sparse muscovite. Quartzite is also present as boudins in the chlorite microbreccia. Lineation and fold axes in these rocks trend approximately S80°E. The chloritic microbreccia is overlain by upper plate marble, dolomite, greenstone, and crenulated greenschist and metasiltstone. Unmetamorphosed porphyry dikes (lithologically similar to Eocene dikes intrusive into the Sanpoil Volcanics in the First Thought graben) cutting the upper plate terminate at the fault, likely indicating that the fault truncates the dikes.

The fault also truncates the First Thought graben. High-angle faults bounding the graben are likely listric to the Kettle River detachment, as shown by Rhodes and Cheney (1981, Fig. 3). To the south, the detachment apparently intersects the concealed northern extension of the Huckleberry Range fault.

Farther south, at Elbow Lake (Fig. 2), the fault cuts a small granitic pluton intruded into rocks of the upper plate (Fox and Wilson, 1989, p. 209). The fault juxtaposes a 30-m-thick layer of lineated and boudinaged sillimanitic quartzite forming the outer margin of the core complex against the granite, which is increasingly fractured and brecciated toward the fault. Two kilometers to the southwest, the fault juxtaposes a lens of cataclastic or-

Figure 12. Isoclinal fold in interlayered augen gneiss and fine-grained gneiss, Okanogan gneiss dome. Pocketknife for scale.

Figure 13. View south of west-dipping Similkameen fault, exposed on the south bank of the Similkameen River. Massive, well-indurated Tertiary conglomerate composed of granite cobbles and boulders in the hanging wall, phyllite of Kobau Formation in the footwall. The hammer is 33 cm long.

thogneiss armoring the outer margin of the core complex against phyllite and marble of the upper plate. The fabric of the orthogneiss grades from chaotic near the fault to penetrative and homogeneous away from the fault. A half kilometer northwest of the fault the fabric of the orthogneiss is indistinguishable from that of other nearby rocks of the core complex. The orthogneiss is judged to be the displaced and partially assimilated continuation of the granite.

MEDIAL ZONES BETWEEN THE GNEISS DOMES

The Toroda Creek, Republic, and Keller grabens drop Eocene volcanic rocks and underlying low-grade metamorphic rocks down between the Okanogan, Vulcan Mountain, and Kettle gneiss domes. The graben fills are extensively intruded by dikes (the hypabyssal intrusive equivalents of the volcanic rocks) and paralleled by linear belts of epizonal Eocene plutonic rocks of the Devils Elbow and Herron Creek suites (Carlson and Moye, 1990; Holder and Holder, 1988).

Toroda Creek Graben

The Toroda Creek graben is bounded on the east by the north-northeast-trending Bodie Mountain fault (Fig. 4). This is a major normal fault with likely displacement of several kilometers. Three-point solutions indicate that the fault dips 20–30 degrees to the west, and a splay exposed in an adit dips 23 degrees (Pearson, 1967, p. 3). At its north end, the fault merges with the curvilinear trace of the shallowly dipping detachment fault bounding the upper surface of the Vulcan Mountain gneiss dome (Pearson, 1967).

Eocene strata in the graben are in general inclined to the east, causing the oldest Tertiary strata (O'Brien Creek Formation) to crop out at the western margin. In detail, judging from Pearson's (1967) map, attitudes of strata in the graben range from flat to vertical, with westward as well as eastward inclinations. Deformation may reflect folding of the Sanpoil Volcanics and O'Brien Creek Formation prior to deposition of the overlying Klondike Mountain Formation. If so, the O'Brien Creek and Sanpoil are probably detached from basement on an east-dipping low-angle fault (cross section, Fig. 4) analogous to the Similkameen fault in the Oroville area and the Lambert Creek fault in the Republic area.

Republic Graben

Bacon Creek and Scatter Creek faults

The Republic graben (Fig. 5) is bounded on the west by a complex zone of subparallel low- to high-angle east-dipping normal faults. The Bacon Creek fault forms the boundary northwest of Republic. This fault dips 60 degrees east and offsets the base of the Tertiary strata by at least 3,400 m (Muessig, 1967, p. 92). West of Republic, the Bacon Creek fault intersects the Scatter Creek fault and transfers most of this displacement to it. The dip of the Scatter Creek fault diminishes south of this intersection, and west of Copper Mountain a flat shear zone marked by 2.5 m of friable breccia may represent the fault (Muessig, 1967, p. 93).

South of Copper Mountain, the fault continues as a zone of subparallel though anastomosing north-northeast-trending faults. The western fault in the zone is typically a moderately to shallowly east-dipping detachment marking the eastern edge of the Coco Mountain window into the Okanogan gneiss dome. Another fault in the zone, the Long Lake fault, dips 35–50 degrees to the southeast and is marked by a zone of breccia that is locally 45 m thick (Staatz, 1964, p. F56).

Sherman fault

The graben is bounded on the east throughout much of its length by the near-vertical Sherman fault. Displacement on this fault is not known, but it is probably several kilometers. Northeast of Republic, the Sherman fault intersects the St. Peter fault, which locally forms the western boundary of the Kettle gneiss dome. The dip of the St. Peter fault shallows northward, and at its northern extremity, the fault dips 40 degrees west (Parker and Calkins, 1964, p. 71). At one place the trace is marked by a breccia zone 215 m wide (Parker and Calkins, 1964, p. 71).

Lambert Creek fault

The contact of the volcanic fill of the graben with subjacent basement (weakly metamorphosed Permian and Triassic strata and Mesozoic to early Tertiary plutonic rocks) is extensively exposed in the east-central part of the graben, east of Republic (Fig. 5). In this area, the volcanic fill is detached from basement by the undulating Lambert Creek fault. At roadcuts 2 km east of Curlew Lake, the fault zone is 2 m thick. Sharply angular pebble- to cobble-size clasts of greenstone and quartzite compose the lower part of the fault zone. It grades upward into an upper part 10–20 cm thick in which sand-size clasts float in an aphanitic matrix. The upper part of the fault zone is cut by anastomosing films and veinlets as much as a half centimeter thick of bluish-black pseudotachylite. Shears subsidiary to the main fault zone have slickensides trending N88°E. The pseudotachylite locally has a barely perceptible, fine, thready lineation approximately perpendicular to the slickensides.

The Lambert Creek fault as shown here (Fig. 5) includes the part north of 48°38′N latitude originally mapped by Muessig (1967). In addition, it includes a 16-km-long segment to the south, mapped by Muessig (1967) as a system of chord-like linear faults along the non-linear eastern contact between volcanic fill on the west and basement rocks on the east. Three-point solutions indicate the southern part of this segment dips 20 degrees to the west.

The Lambert Creek fault likely forms the basal contact of the Eocene fill of the graben far west of its mapped trace, perhaps as far west as the axial part of the graben. This interpretation differs from that of Muessig (1967, fig. 14), who suggested that the continuation of the Lambert Creek fault west of the Mires Creek fault intersected rocks structurally high in the pile that are now eroded away. As interpreted here (cross section, Fig. 5), the Lambert Creek fault cuts off the Mires Creek fault, imbricating the Tertiary section west of its trace. The Mires Creek fault is likely a former western continuation of the Lambert Creek fault.

Sanpoil syncline

The O'Brien Creek Formation, Sanpoil Volcanics, and the basal part of the Klondike Mountain Formation are folded into a large-scale syncline that forms the structural axis of the Republic graben. The syncline parallels the general trend of dikes cutting rocks in and outside the graben. The fold, named the Sanpoil syncline by Muessig (1967, p. 110), is approximately 45 km long. Amplitude of the fold is probably several kilometers.

Near Republic, the west limb dips 10–20 degrees east, and the east limb dips 70–90 degrees west and locally is overturned. Farther south, the west limb steepens to 40–60 degrees, and the east limb shallows to 50–60 degrees.

The middle and upper parts of the Klondike Mountain Formation are only gently inclined toward the fold axis. Older Tertiary strata were therefore folded chiefly during the interval between deposition of the lower part of the Klondike Mountain (the Tom Thumb Tuff Member as shown by Muessig, 1967, pl. 1) and the unconformably overlying epiclastic deposits of the middle part of the Klondike Mountain. Minor folding continued during or after deposition of the middle part of the Klondike Mountain Formation.

CONTACTS AT THE NORTHERN AND SOUTHERN MARGINS OF THE DOMES

Contacts between lower and upper plates at the northern margin of the Okanogan gneiss dome and the southern margin of Okanogan and Kettle gneiss domes are gradational zones in the syntectonic plutons of the Keller Butte suite. Locally in the central parts of the domes, remnants of low-grade metamorphic rocks of the upper plate grade downward into higher grade rocks of the lower plate. Contact relations at the northern margin of the Okanogan dome are further complicated by the presence of a northwest-trending zone of shearing, referred to as the Ninemile Creek fault zone.

Ninemile Creek Fault Zone

The Wagonroad Coulee fault is cut off by the N50°–65°W-trending Ninemile Creek fault zone, which places the much-attenuated Osoyoos pluton and upper plate rocks against the nonlineated northern margin of the syntectonic Mount Hull pluton of the Keller Butte suite. The zone is marked by a 50- to 1,500-m-wide belt of coarse-grained nonlineated syenite gneiss. Bodies of pyroxenite and serpentine locally cut the upper plate rocks adjacent to the syenite gneiss. The syenite gneiss grades erratically northeastward over 20 to 30 m into the metasedimentary and metavolcanic rocks of the upper plate; hence it likely originated through metasomatic replacement of these rocks (Krauskopf, 1941).

At the 49th parallel, 2 km east of Osoyoos Lake, and at excellent exposures in abandoned railroad cuts 9 km east of Oroville, the zone consists of coarse-grained mafic to felsic syenite cut by pegmatite. At these localities the zone is intensely fractured. At exposures to the east, however, the zone is much less intensely fractured.

At its western extremity and north of Mount Hull (Fig. 3), gneissosity in the zone—and presumably the zone itself—dips 25–70 degrees north-northeast. To the east, the dip shallows to near horizontal, and the zone, as represented by discontinuous bodies of the syenite gneiss, appears to be dismembered. Movement in the zone is reflected only by the weaving, anastomosing gneissosity, which is thought to represent coarsely recrystallized shear structure.

Hornblende from the syenite gneiss yielded K-Ar ages of 59.6 ± 1.7 and 55.3 ± 1.6 Ma, and biotite, 51.3 ± 1.5 Ma (Fox and others, 1976, p. 1220). Epidote and titanite yielded fission-track ages of 63 ± 3 and 66 ± 7 Ma (Naeser and others, 1970). These ages probably reflect movement in the zone and crystallization of the syenite between 65 and 60 Ma, followed by progressive cooling of the syenite through the blocking temperatures of these minerals between approximately 60 and 50 Ma. Brittle fracturing in the western part of the zone and offset of the Similkameen fault suggest local reactivation of the fault after 50 Ma.

Gradational Contacts between the Lower and Upper Plates in Plutons of the Keller Butte Suite

Metamorphic rocks of the interior of the Okanogan dome dip shallowly northward to northeastward beneath mylonitic megacrystic to medium-grained granite gneiss composing the Mount Hull pluton (Fig. 10). At Mount Hull, this contact is extensively invaded by sill-like bodies of tonalite, which are mylonitic but distinctly less so than the granite gneiss. The dynamic metamorphic features of the core complex—mylonitization, shallowly dipping foliation, and north-northwest-trending lineation—extend from the interior of the gneiss dome northward across the contact where they continue with gradually diminishing intensity several kilometers into the granite. The northern boundary of the gneiss dome and the contact between upper and lower plates is thus a gradational zone several kilometers wide.

Contacts of deformed parts of the plutons with the underlying high-grade metamorphic rocks are conformable with foliation and layering, perhaps as a result of extreme transposition during ductile flow. However, the several generations of leucocratic granite and pegmatite-alaskite dikes cutting the high-grade metamorphic rocks have been rotated varying amounts. Although the oldest are conformable with foliation and layering, each succeeding generation is less so, indicating that ductile flow and magma generation were concurrent. Ductile deformation of the Keller Butte plutonic rocks evidently began shortly after or perhaps concurrently with crystallization of the oldest plutons in the suite.

The nonfoliated northern part of the Mount Hull pluton is bounded on the northwest by the Ninemile Creek fault zone and by the northern extension of the Wagonroad Coulee fault. To the east, the pluton cuts the syenite marking the trace of the Ninemile Creek fault and intrudes the low-grade metamorphic rocks of the upper plate to the north (Fox, 1978) (Fig. 2). These rocks are metamorphosed in the greenschist facies at distances greater than a kilometer from the pluton and locally reach staurolite grade locally within a kilometer. Steeply dipping dikes of pegmatite radiate northeastward from the pluton into the metamorphic rocks.

The southern boundary of the Okanogan dome is a gradational contact in a nested series of plutons of the Keller Butte suite. The southernmost of these plutons, the megacrystic Coyote Creek pluton (Fig. 2), intrudes older plutons of the suite, which are everywhere penetratively deformed, and is itself locally deformed along its northern margin. The gradational contact is truncated to the west by the Wagonroad Coulee fault and its southern continuation, the Goose Flats fault.

The southern boundary of the Kettle dome is a gradational contact in the Daisy Trail pluton of the Keller Butte suite (Fig. 2). Dynamic metamorphic features of the core complex—domal foliation and unidirectional lineation—gradually diminish in intensity from north to south in the pluton, and the southern part of the pluton is massive. The gradational contact trends southeast and presumably dips shallowly southwest, parallel to foliation. The contact is truncated by the Kettle River fault, which continues to the west-southwest for at least 5 km and probably 15 km, cutting upper plate rocks. The contact of the pluton with low-grade phyllite of the upper plate to the south is intrusive. The southeastern lobe of the pluton east of the southwestern continuation of the Kettle River fault also invades low-grade metamorphic rocks of the upper plate.

Gradational Contacts with Metamorphic Rocks of the Upper Plate

Low-grade metamorphic rocks overlap the east-central part of the Okanogan dome near Aeneas. The low-grade rocks include phyllite, black quartzite (likely the Covada Group), greenstone, and chert breccia (likely the Anarchist Group). They are extensively cut by relatively undeformed pegmatite dikes and generally lack the penetrative lineation and foliation characteristic of the Okanogan dome. They are therefore considered parts of the upper plate. However, they apparently grade westward into gneiss and schist that are infolded with quartzo-feldspathic gneiss of the dome (Rinehart and Greene, 1988). It thus appears that in this area the upper plate has been infolded with and partially assimilated by the core complex.

Effects of dynamic metamorphism at least locally decrease upward in the Kettle dome, suggesting gradation to upper plate rocks now removed by erosion. For example, near Grizzly Peak (Fig. 2), shallowly dipping domiform foliation (fluxion structure) and lineation are penetrative at lower elevations, but within a hundred meters or less (vertically) of ridge summits, the intervals between discrete, lineated shear surfaces increases to several centimeters, then farther upward, to several meters. The component of the lateral displacement on these surfaces in a direction orthogonal to lineation ranges from several centimeters to 10 m, as revealed by offset of pegmatite dikes (Fig. 14).

SUMMARY AND INTERPRETATION

The Okanogan, Vulcan Mountain, and Kettle gneiss domes form metamorphic core complexes whose internal structures—penetrative lineation, semihorizontal fluxion structure, folded and near-recumbent dikes—indicate large-scale horizontal translation and thinning through simple shear. Metamorphic mineral assemblages indicate that the deepest exposed levels of the core complexes reached pressures and temperatures of the amphibolite facies and, locally, the granulite facies, followed by retrogression to temperatures and pressures of the greenschist and zeolite facies. These features suggest that the core complexes are uplifted and exposed elements of formerly ductile middle crust.

The domes are temporally, spatially, and genetically associated with the Paleocene and Eocene Colville batholith. The batholith comprises three plutonic suites: from oldest to youngest, the Keller Butte, Devils Elbow, and Herron Creek suites (Holder and Holder, 1988). The leucocratic biotite and biotite muscovite granites of the Keller Butte suite form extensive syntectonic bodies both inside and outside the domes. These granites were likely derived through partial melting of crustal rocks and were emplaced prior to and concurrent with ductile flow of the protolith of the high-grade metamorphic rocks (Tonasket Gneiss and metamorphic rocks of Tenas Mary Creek) in the gneiss domes. The mafic hornblende-biotite granodiorite and monzodiorites of the Devils Elbow suite were likely generated in part through melting of mantle and emplaced during the waning stages of ductile flow. The hornblende-biotite granodiorites of the Herron Creek suite, although synextensional, postdate climactic early Eocene volcanism and ductile flow.

The association of these rocks with the domes indicates that the causal tectonic event included ultrametamorphic as well as extensional processes, processes in which rocks at depth were partially melted, mobilized, and either intruded or extruded at higher structural levels.

The extensional and ultrametamorphic episode was approximately concurrent with formation of the Colville batholith, judging from contact relations and relative deformation of its components. The oldest plutons of the batholith, the Keller Butte suite (60–55 Ma), and dikes related to those plutons are progressively more deformed with increasing age in the Kettle and Okanogan domes. These rocks are cut by undeformed plutons of the Herron Creek suite (post-Sanpoil Volcanics) and by hornblende rhyodacite dikes of probable Sanpoil age (≈52–50 Ma). The Kettle dome is also cut by undeformed biotite rhyolite dikes that are probably the same age as the O'Brien Creek Formation (55 Ma?) and by undeformed plutons of the Devils Elbow suite. The ductilely deformed rocks of the Okanogan gneiss dome are cut by the Devils Elbow suite (Swimptkin Creek pluton) which, though largely undeformed, is locally penetratively sheared, lineated, and mylonitized. The Swimptkin Creek pluton evidently intruded the gneiss dome during the waning stage of ductile flow (Singer, 1984). The data summarized above indicate that ultrametamorphism and accompanying ductile flow in the domes was under way shortly after, if not before, 60 Ma (compare Orr and Cheney, 1987, p. 69.) Ductile deformation at levels presently exposed in the Kettle dome apparently had ended by 55 Ma and in the Okanogan dome by approximately 50 Ma.

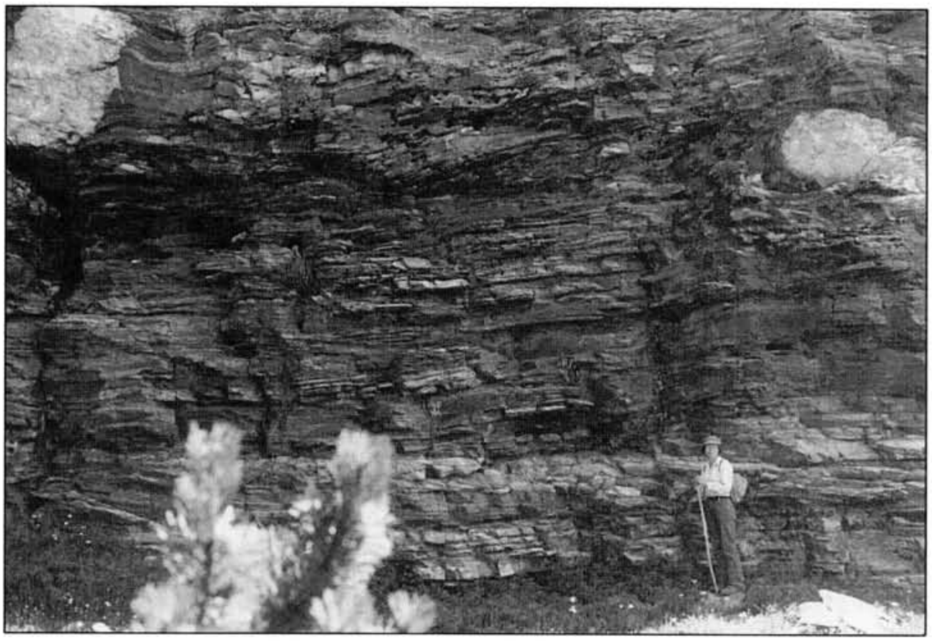


Figure 14. View east (oblique to horizontal lineation trending S65°–80°E) of light-toned bodies of pegmatite cutting horizontally layered dark-toned quartzite and biotite schist at Grizzly Peak, Kettle gneiss dome. The pegmatite bodies are inferred to be parts of an initially steeply dipping, east-striking late-kinematic dike, dismembered by translation perpendicular to lineation on thin shear zones conformable with layering. The pegmatite is little deformed between the shear zones.

The Tonasket Gneiss, previously considered Late Cretaceous in age (Fox and others, 1976), and the metamorphic rocks of Tenas Mary Creek were formed during this ultrametamorphic event and are thus Paleocene and early Eocene in age.

The presence of gradational contacts between the domes and the upper plate and of incompletely assimilated bodies of upper plate rocks in the domes indicates that the domes grew upward through accretion. The ductility of upper plate rocks immediately above the ductile-brittle transition zone evidently increased, probably as a result of warming by conduction from below and from invasive magmas, through internal frictional heating, and through permeation of liquids associated with pegmatitic and granitic magmas. When sufficiently ductile, these rocks then flowed with the main body of the dome, acquiring the distinctive foliation and other deformational features characteristic of the lower plate. Thinning of the brittle crust over the domes thus resulted in part from upward advance of the ductile front and in part from mechanical thinning of the upper plate through faulting.

The domes are ringed by faults on which their ascent from mid-crustal depths to the surface was partly accommodated. Associated with the faults are north-northeast-trending dikes, grabens, and folds, notably the Sanpoil syncline. The trend of dikes and grabens indicates extension of the brittle upper crust in a west-northwest direction.

The north-northeast-trending Sanpoil syncline probably formed synchronously with the Lambert Creek fault (Muessig, 1967, p. 111). The folding seemingly requires compression in the same direction at approximately the same time that rocks in and east of the graben (and in the region) were being extended through faulting and intrusion of dikes. This is possible if the folding was restricted to an upper plate decoupled from the regional stress.

The floor of the Republic graben was evidently tilted through differential movement on the bounding faults, causing an upper plate composed of the graben fill to break away from subjacent basement and slide westward toward the deepest part of the graben. In this formulation, the Sanpoil syncline is a fold in the detached upper plate, formed as it crumpled due to loads resulting from the lateral component of the weight of its extensions at higher elevations. The Lambert Creek fault is accordingly the detachment fault on which the upper plate broke away from autochthonous basement and moved west toward the keel of the graben.

The faults associated with the domes can be grouped into five classes:

(1) Low-angle to vertical normal faults that cut upper crustal rocks. This class includes the inferred fault concealed by Osoyoos Lake and the Crystal Springs, Goose Flats, Bodie Mountain, Bacon Creek, and Sherman faults. Dips range from near-vertical (Sherman fault) to 20 degrees (Bodie Mountain fault). The faults bound the grabens and, along strike, obliquely intersect the low-angle faults bounding the domes, as discussed below. The faults of this class probably originated as major high-angle planar normal faults cutting the entire brittle crust. Except for the Sherman fault, they have been rotated to dips of 20–30 degrees through thinning

of brittle crust. The Sherman fault likely originated late in the extensional process.

(2) Low-angle normal (detachment) faults that place upper crustal rocks against the formerly ductile middle crustal rocks of the gneiss domes (Wagonroad Coulee, Scatter Creek, St. Peter, and Kettle River faults). These faults flank the eastern and western sides of the domes and commonly dip 10–20 degrees outward. The faults are locally marked by zones of chloritic breccia, and in places they bound zones of mylonite half a kilometer or more thick in the lower plate. On strike, these faults transect the gradational zones discussed below and continue into the upper plate as the major normal faults discussed above. In the Eocene the faults evidently extended from the surface or near-surface (where they controlled the location of then-active depocenters) to the brittle-ductile transition zone.

(3) Diffuse or gradational low-angle to sub-horizontal (detachment) zones that locally separate the formerly ductile middle crustal rocks of the domes from upper crustal (brittle) roof rocks. These contacts are gradational zones as much as a kilometer thick in syntectonic granites of the Keller Butte suite. Where the granites are absent, the zones form gradational contacts between brittlely deformed greenschist-facies rocks of the roof and ductilely deformed amphibolite-facies rocks of the domes. The zones likely mark the shallowest crustal level to which the ductile front (the transitional contact between brittle upper crust and ductile middle crust) climbed during the Paleocene and Eocene.

(4) Low-angle to sub-horizontal (detachment) faults that offset Eocene volcanic and epiclastic basin fill from basin floors. These faults include the Similkameen and Lambert Creek faults, and probably the curvilinear faults bounding the Keller graben (Fig. 2). Eocene basin fill and slivers of older rocks forming the upper plate of these faults apparently broke away from their depositional basal contacts and slid laterally to deeper parts of the basin. Sliding was probably driven by tilting of the basins resulting from differential movement on basin-bounding normal faults and from rising of the gneiss domes.

(5) Inferred transform faults cutting upper crustal rocks—rocks that were or became brittle through cooling during extension. These faults include the Ninemile Creek fault and the fault that bounds Eocene rocks 5 km south of Kettle Falls (Fig. 2). They strike west-northwest, parallel to the direction of extension of brittle upper crust; hence they are likely transform faults.

The three classes of low-angle normal faults discussed above are commonly referred to as detachment faults, although they have very different origins and kinematic implications. The faults offsetting Tertiary basin fill from basin floors are shallow crustal structures that were initiated as and have remained gently inclined faults. The rotated major normal faults, on the other hand, formed as steep faults that probably were subsequently rotated to low inclinations. They cut the entire brittle crust and thus necessarily, in their lower parts, placed ductile crust against brittle crust thus forming major parts of the lateral boundaries of the core complexes. The diffuse or gradational zones forming upper contacts of the core complexes are likely attenuated remnants of the ductile-brittle transition zone. They began as and have remained near-horizontal zones through the extensional process.

The low-angle faults (detachments) bounding the eastern and western sides of the core complexes probably started as planar normal faults with dips in the range seismically observed at major historically active normal faults in and peripheral to the Basin and Range Province, that is, 40–75 degrees (for example, Westaway, 1991). If so, rotation to their present dip of 12–30 degrees implies that brittle crust initially overlying the core complexes was extended 30 to 300 percent.

In summary, the onset of ultrametamorphism and extension of the Okanogan region in the early Tertiary probably began at approximately 60 Ma with partial melting in the crust. Intrusion of the resulting magmas formed the granites of the Keller Butte suite. Parts of these plutons extended to shallow crustal levels. Ductile flow of middle crustal metamorphic rocks, if not already under way, began shortly thereafter. Eruptions of silicic pyroclastic rocks of the O'Brien Creek Formation tapped volatile components of these granitic melts and were the first surficial expression of the ultrametamorphic event then in progress at depth. The shallow basin in which the sediments of the O'Brien Creek accumulated was subsequently broken by faults, tilted, and, between 52 and 50.3 Ma, inundated by ash flows and lavas of the Sanpoil Volcanics. At depth, hypabyssal plutonic rocks of the Devils Elbow suite, coeval with the Sanpoil, pooled in areas of lineated and foliated high-grade metamorphic rocks and, concurrently or subsequently, flowed ductilely along with their host as the gneiss domes grew beneath the Eocene basins.

Extension probably climaxed between 50 and 48 Ma, after eruption of the Sanpoil Volcanics and before eruption of the lavas of the Klondike Mountain Formation. Deepening and tilting of the ancestral O'Brien Creek–Sanpoil basin culminated at this time in detachment (at the Lambert Creek fault) and sliding of the basin fill toward the axis of the basin. The Sanpoil was concurrently folded, broken, and decoupled from its feeder dike complex, now in part exposed along the eastern side of the Republic graben. Masses of greenstone and granite stripped from the rising Okanogan and (or) Kettle gneiss domes then poured into the basin, forming the monolithologic breccias that now overlie the deformed rocks of the Sanpoil. To the west, the Tertiary basin fill at Oroville broke away from its floor and slid westward off the bulging Okanogan gneiss dome.

Volcanism then resumed with eruption of the lava flows of the Klondike Mountain Formation, which pooled in the central part of the still-subsiding O'Brien Creek–Sanpoil basin. In a final deformational paroxysm, the basin was wedged apart by a spine of its basement rising up through its floor, exposing the Vulcan Mountain gneiss dome and separating the Toroda Creek graben from the Republic graben. Concurrent movement on the Kettle River fault cut off the base of the First Thought graben and back-tilted its volcanic fill.

The extensional episode ended about 48 Ma, leaving the Okanogan region possibly doubled in width, with extensive tracts of still-steaming lava, crystallizing plutonic rocks, and newly disinterred mid-crustal metamorphic rocks.

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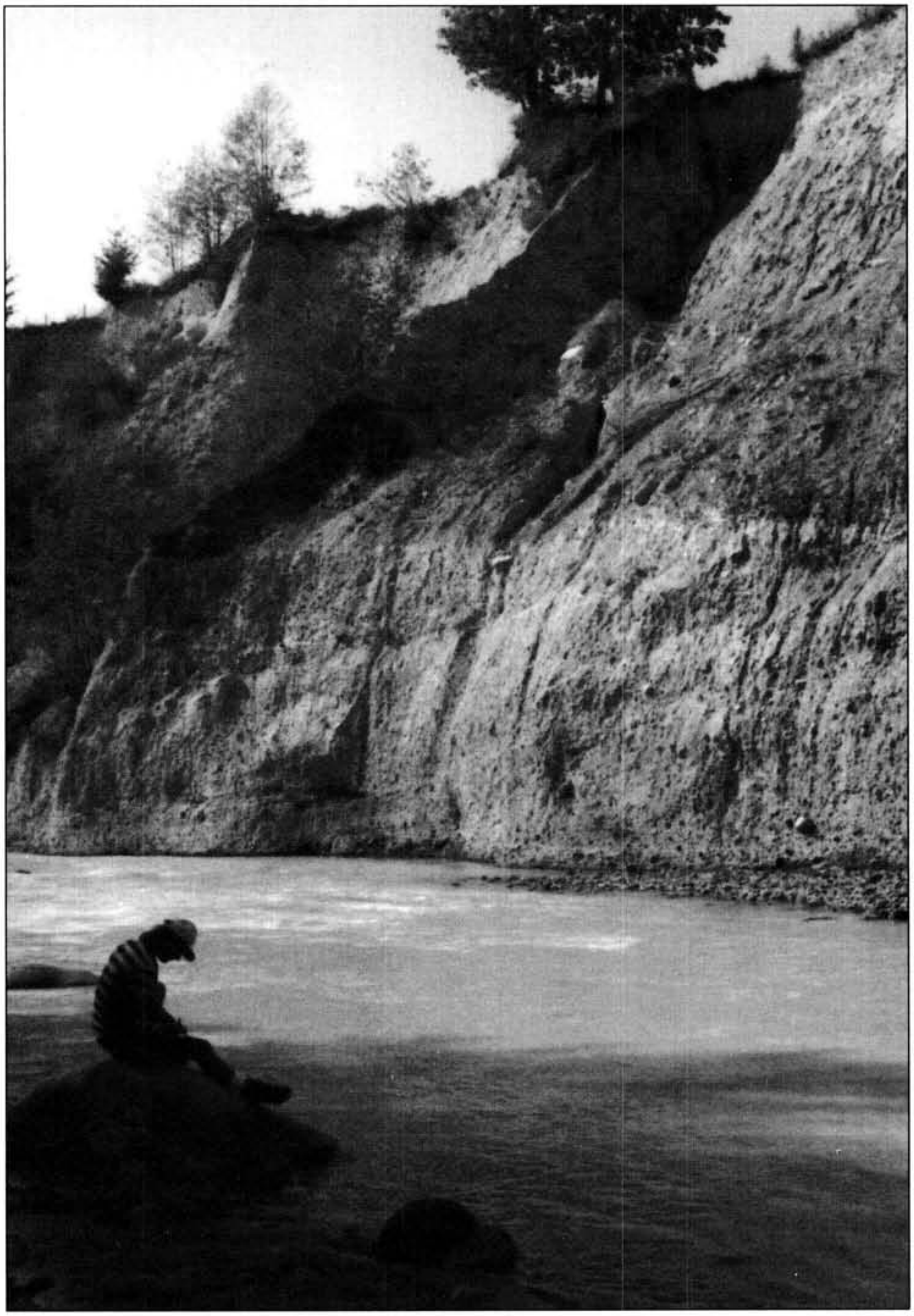

The Osceola Mudflow deposit (3–5 m thick) caps a 30-m-high terrace composed chiefly of glacial deposits of Vashon age. The Osceola Mudflow was a cohesive lahar that was generated about 5,000 yr B.P. when at least 3 km^3 of the former summit of Mount Rainier collapsed, probably during eruptive activity. The lahar flowed down the White River valley and now underlies hundreds of square kilometers of the Puget Lowland. This location is along the White River a few hundred meters downstream (west) of State Route 410 at Buckley. A short distance upstream (to the right in the photo), the Osceola Mudflow becomes a valley fill deposit, and a nested sequence of terraces composed of numerous lahar deposits from Mount Rainier that post-date the Osceola locally overlies it. Photo by Patrick Pringle, 1993.

Major Faults, Stratigraphy, and Identity of Quesnellia in Washington and Adjacent British Columbia

E. S. Cheney, M. G. Rasmussen, and M. G. Miller
Department of Geological Sciences, AJ-20
University of Washington
Seattle, WA 98195

ABSTRACT

The Quesnel terrane in northeastern Washington and adjacent British Columbia has three regional unconformity-bounded sequences. The oldest, a Carboniferous to Permian assemblage, consists of the ophiolitic Knob Hill Group and the predominantly pelitic Attwood Group, the relative ages of which are still uncertain. These are unconformably overlain by the Upper Triassic (Carnian to Norian) Brooklyn Formation of polymict clastic rocks, fossiliferous limestone, and greenstone. In Washington, part of the Cave Mountain succession northwest of Omak is in this sequence. Brooklyn rocks are, in turn, unconformably overlain by pelitic and mafic volcanic rocks of the Jurassic Rossland Group.

These three sequences are cut by the Chesaw thrust, which places the Knob Hill Group over the Attwood Group. The best indicators of the fault are ultramafic rocks, many of which are altered to listwanite. Preliminary kinematic indicators suggest that the direction of transport was top-to-the-east. The age of the thrust is poorly constrained.

On the southwestern margin of Quesnellia, the undated southwestern metamorphic belt includes amphibolite-facies orthogneisses and pelitic gneisses. The most distinctive orthogneiss contains ferrohastingsite and megacrystic K-feldspar. These metamorphic rocks are thrust over the Quesnellian greenschist-facies rocks along the Dunn Mountain fault. Because of this metamorphic inversion and phyllitic zones at least a kilometer thick along the fault, the Dunn Mountain is probably a crustal-scale fault; it is younger than the Chesaw fault.

The Chesaw and Dunn Mountain faults are folded by northwesterly striking and plunging folds. Quesnellian rocks are unconformably overlain by tiny remnants of clastic rocks of the Cretaceous Sophie Mountain Formation and by the more extensive arkosic and volcanic rocks of the Eocene Challis sequence.

The lithologies, stratigraphy, structural history, and regional setting of pre-Cretaceous Quesnellia are similar to those of the Golconda terrane and overlying rocks of northwestern Nevada.

INTRODUCTION

This paper is a preliminary description of the Quesnel terrane, or Quesnellia, in north-central Washington and adjacent British Columbia (Fig. 1). The main purpose is to describe two major low-angle faults, the Chesaw and the Dunn Mountain faults, that cut the Quesnellian rocks. A second objective is to stress that the sequence stratigraphy of Quesnellia is more widespread than commonly realized. Thirdly, we point out the lithotectonic similarity of the Quesnellian rocks to rocks of northwestern Nevada.

Northeastern Washington and adjacent British Columbia constitute a region of complex geology and limited outcrop. Little mapping at scales greater than 1:40,000 has been done, and the geochronological, paleontological, and geochemical databases are small. Thus, our discussion of Quesnellia acknowledges a variety of unresolved questions.

Geologic Setting

Figure 2 (p. 52) shows the regional geology of Quesnellia. The Washington part of this map is modified from Stoffel and others (1991, the 1:625,000 map) by reinterpreting their 1:250,000 map. The part in British Columbia is a compilation of Parrish and others (1988), Tempelman-Kluit (1989), Fyles (1990), and Andrew and others (1991). Table 1 is the explanation of the units shown in Figure 2 and subsequent maps.

Quesnellia is dominated by upper Paleozoic mafic volcanic and pelitic rocks that are unconformably overlain by both Triassic and Jurassic rocks (Monger and others, 1982; Silberling and others, 1987; Monger and Berg, 1987). Nearby terranes in British Columbia also are overlain by these or similar Mesozoic rocks; together with Quesnellia, these terranes form the amalgamated Intermontane superterrane that accreted to North America in the mid-Jurassic (Monger and others, 1982; Monger and Berg, 1987; Lambert, 1989). In the United States, Quesnellia is bounded by an eastern assemblage of predominantly pelitic Ordovician to Carboniferous rocks. On the southwest it is bounded by a belt of amphibolite-facies metamorphic rocks. Quesnellia is unconformably overlain by tiny remnants of coarse clastic rocks of the Upper Cretaceous Sophie Mountain Formation (Little, 1982; Brown and others, 1981, fig. 8) and the more extensive and much thicker volcanic and arkosic rocks of the Eocene Challis sequence (Cheney, 1994, this volume).

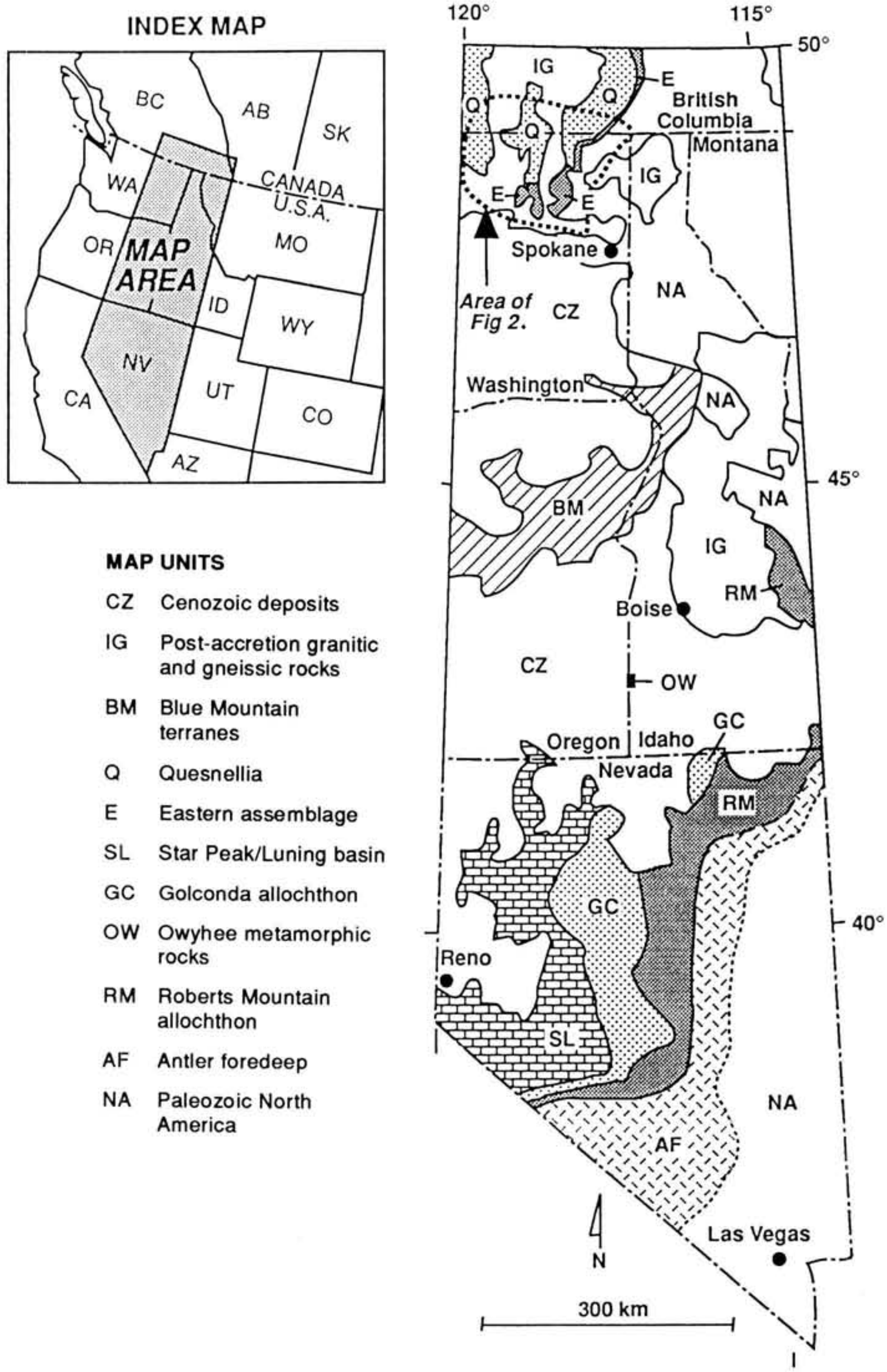

Figure 1. Lithotectonic terrane map of part of the North American cordillera. Sources of data: Monger and Berg (1987), Silberling and others (1987), Whiteford (1990), Murchey (1990), Burchfiel and others (1992), and Figure 2, this paper.

Figure 2 illustrates the discontinuous distribution of the Quesnellian rocks. This is caused by the younger metamorphic core complexes and the structural lows in which the Eocene Challis sequence is preserved (Cheney, 1994, this volume). This discontinuous distribution probably accounts for the previous non-recognition of the regional stratigraphic and structural features we describe.

Background

Most Quesnellian rocks in Washington were last mapped in the 1960s and early 1970s (Parker and Calkins, 1964; Muessig, 1967; Pearson, 1967, 1977; Fox, 1970, 1978; Rinehart and Fox, 1972, 1976), before concepts of plate tectonics and accreted terranes became popular (Monger and others, 1982). Nor were these concepts incorporated into the last major geologic synthesis of northeastern Washington (Stoffel and others, 1991) and the most recent geologic map of the state (Schuster, 1992).

Daly (1912) named Upper Paleozoic greenschist-facies pelites and greenstones east of the Okanagan Valley in British Columbia the Anarchist Series. Waters and Krauskopf (1941) and Rinehart and Fox (1972) believed that the greenstones unconformably overlie the pelitic section. Rinehart and Fox (1972) named the pelitic rocks the Anarchist Group and the greenstones the Kobau Formation.

Fox and Rinehart (1968) noted numerous bodies of fuchsite-bearing magnesite-dolomite-quartz rocks along the greenstone/pelite contact. They believed that these rocks are most likely sedimentary magnesite at the base of the greenstone succession or are dolomitic beds that have been hydrothermally replaced by magnesite. In 1975, Cheney mapped the contact between the pelitic rocks and the overlying greenstones between Molson and Chesaw (east of the Okanogan Valley in Washington) as a thrust (McMillen, 1979). Fox and Rinehart (1968, table 2) had reported that the magnesitic rocks contain 1,500–2,000 ppm Cr and Ni. Cheney confirmed the high Ni content with five additional samples and concluded that the magnesitic rocks resulted from the carbonatization of ultramafic rocks along the thrust (McMillen, 1979, table 1, p. 30). Carbonated ultramafic rocks, now termed listwanites, are commonly regarded as dismembered parts of ophiolites in major fault zones; listwanites can have gold contents that reach economic grades and tonnages (Ash and Arksey, 1990).

McMillen (1979) discovered additional localities of magnesite-bearing rocks in the Chesaw area. Furthermore, he noted that the Triassic Brooklyn Formation occurs structurally below the magnesites and the Kobau greenstones. McMillen named the Chesaw thrust and noted that the contact between greenstones and underlying pelites that Fox (1970, 1978) had mapped as far west as Molson, and which Cheney had mapped as a thrust, is the western continuation of the Chesaw thrust.

On the basis of the distribution of magnesite-bearing rocks in the Okanogan Valley and of serpentinites in the Kettle River valley to the east, Cheney and others (1982) and Orr and Cheney (1987) inferred that the Chesaw thrust is regional in extent. Because the Chesaw fault had not been well documented and, therefore, was underemphasized by Stoffel (1990b) and Stoffel and others (1991), we began mapping in the Okanogan Valley in 1990. The mapping by Fox (1970, 1978) and Rinehart and Fox (1972, 1976) allowed us to plan traverses to test hypotheses.

Despite considerable previous Canadian work in Quesnellia, much of which was reviewed by Peatfield (1978), Fyles (1990) provided the breakthrough. In the Greenwood–Grand Forks area, in which the Chesaw fault had been inferred (Orr and Cheney, 1987, fig. 2), he mapped not one, but four northward dipping thrusts. Fyles' descriptions (1990) of the pelite-dominated Attwood Group and the greenstone-dominated Knob Hill Group appeared to match those for the Anarchist Group and greenstone-dominated Kobau Formation of Rinehart and Fox (1972) in Washington. Fyles also redescribed the lithology, stratigraphy, and paleontologic definition of the Brooklyn Formation, which unconformably overlies both the Attwood Group and the Knob Hill Group. As a result, the Greenwood–Grand Forks area is the best area with which to compare the stratigraphy and structure of the Quesnellian rocks of Washington.

STRATIGRAPHY

In Figure 3 and Table 1, we divide southern Quesnellia into three sequences. Because previous authors have demonstrated that unconformities exist between these successions, we conclude that these are regional unconformity-bounded sequences, rather than tectonostratigraphic units (or subterranes). To correlate radiometric dates with the stages of Mesozoic systems, we use the time scale of Sloss (1988).

We start with a review of the better studied Canadian sections before discussing similar rocks in Washington. We then propose several stratigraphic revisions for British Columbia and Washington; most of these will require confirmation by definitive lithostratigraphic, paleontological, or geochemical studies.

A belt dominated by Ordovician to Carboniferous pelitic rocks lies between Quesnellia and the Proterozoic to Paleozoic unconformity-bounded sequences indigenous to North America on the southeast (Fig. 2). Most of the rocks in this belt are described as the Ordovician Covada Group (Stoffel and others, 1991; Smith, 1991). Megascopically, most of the rocks resemble Anarchist pelites, but whereas the Anarchist is composed mostly of volcanic material, the Covada Group is subarkosic (Fox and Rinehart, 1974; Smith and Gehrels, 1991; Smith, 1991). The Covada Group extends as far west as the southern part of the Republic graben (K. F. Fox, Jr., U.S. Geological Survey, written commun., 1992). Bedded barite deposits occur north of Colville (Moen, 1964; Dutro and Gilmour, 1989); at least one of them is Devonian (Dutro and Gilmour, 1989). We informally refer to these Ordovician to Carboniferous predominantly pelitic rocks as the eastern assemblage.

The relation of Quesnellia to the eastern assemblage is unknown. The eastern assemblage is nowhere overlain by the Triassic and Jurassic sequences characteristic of Quesnellia (Fig. 2). Thus, the eastern assemblage could be a separate terrane, a distal portion of North America, or both, that was thrust back onto the continent prior to the arrival of the amalgamated terrane that contains Quesnellia (Smith and Gehrels, 1991). Accordingly, we omit the eastern assemblage from our discussion of Quesnellian rocks.

Western Assemblage

The recognition of thrust faults within Quesnellia indicates that previous authors were describing a tectonic assemblage, not a stratigraphic succession. Accordingly, we use "western assemblage" for all the Carboniferous to Permian greenschist-facies rocks. In this catchall unit, we also provisionally include (Fig. 2) the Carboniferous to Permian pelitic units east of the Kettle metamorphic core complex—the Mount Roberts and Flagstaff Mountain Formations (Stoffel and others, 1991; Andrew and others, 1991). However, our focus will be on that part of Quesnellia west of the Kettle metamorphic core complex.

To describe the western assemblage, we use the names for rock units currently in use in the Greenwood–Grand Forks area. These are Attwood Group for the pelite-dominated succession and the Knob Hill Group for the greenstone- or chert-dominated succession (Church, 1986; Fyles, 1990). Other names in southern Quesnellia may have precedence, but these two had originated by 1912 (Little, 1983; Church, 1986). Without definitive biostratigraphic, lithostratigraphic, or other evidence, it is not possible to demonstrate whether all pelite-dominated and all greenstone- or chert-dominated units of southern Quesnellia are everywhere the same lithostratigraphic units. However, in the interest of promoting geologic continuity (until proven otherwise), we extend these names to Washington.

Attwood Group

Type area: According to Fyles (1990), the Attwood Group in the Greenwood–Grand Forks area (Fig. 4) is mainly dark-gray to black argillite, siliceous argillite, phyllite, and slate, with minor dark-gray limestone, chert- and argillite-chip conglomerate, and greenstone. Fossils in limestones range in age from Carboniferous to Permian. The rocks are tightly folded. Fyles (1990) was not able to establish a consistent internal stratigraphy in any of the five areas of pelitic rocks north of the international boundary (Fig. 4). He assumed that, on the basis of their general lithologic similarity, all five are Attwood Group.

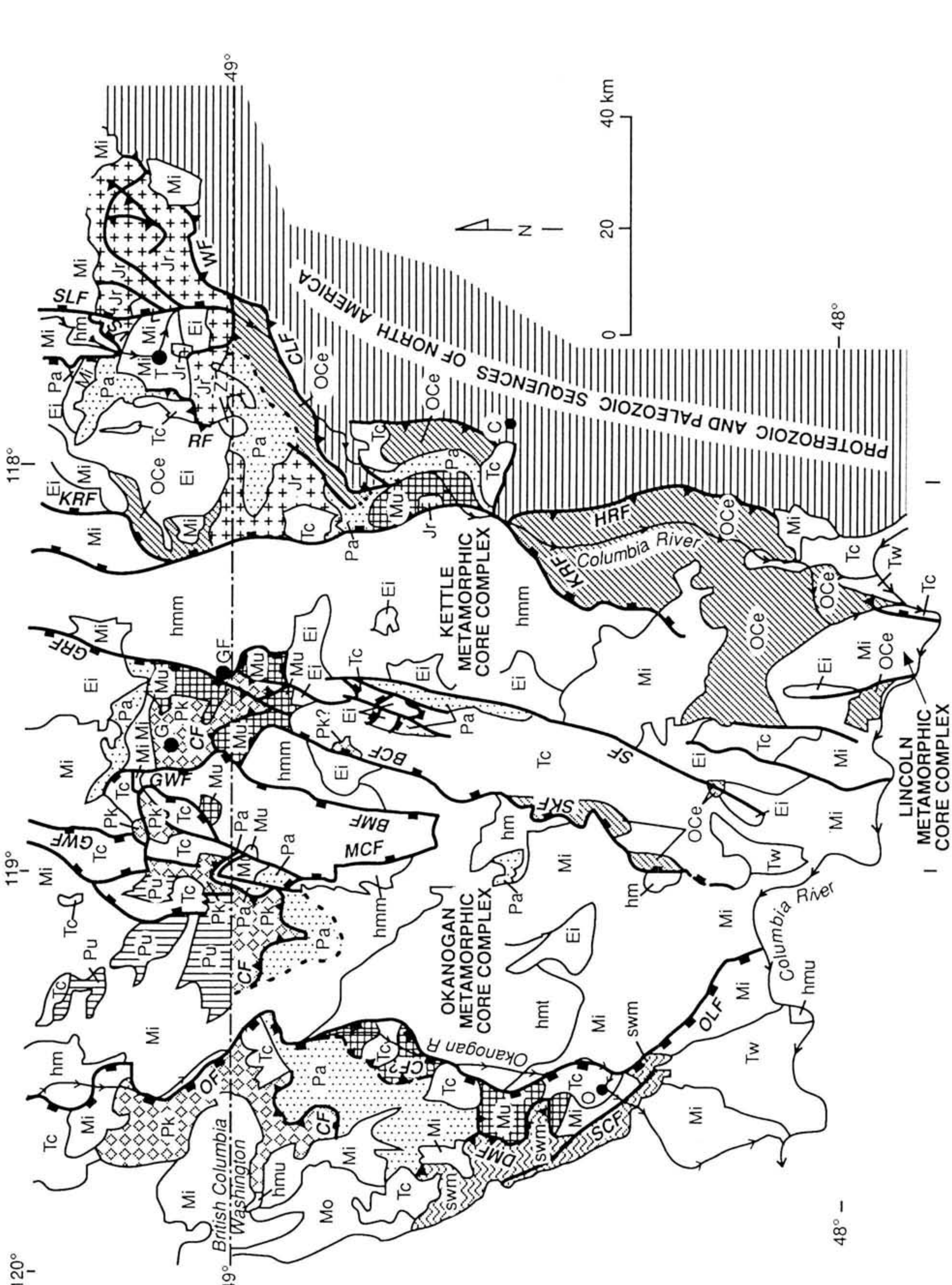

Figure 2. Regional geologic map of the southern part of Quesnellia and bounding units. See Table 1, facing page, for explanation, including abbreviations. Sources of data: Andrew and others, 1991; Fyles, 1990; Parrish and others, 1988; Stoffel and others, 1991; and Tempelman-Kluit, 1989.

Table 1. Explanation for geologic and geographic features of Figures 2, 4, 5, 6, and 7

MAP UNITS

Cover sequences

Tw	Miocene to Pliocene Walpapi Sequence, predominantly Columbia River Basalt Group	
	Tc	Eocene Challis Sequence
	Tcsi	Scatter Creek hypabyssal rocks
	Ts	Sanpoil Volcanics, rhyodacitic
	To	O'Brien Creek Formation, arkosic
Z	Cretaceous Zuni Sequence Sophie Mountain Formation, conglomerate	

Plutons

Ei	Eocene felsic plutons
Ji	Jurassic Shasket Creek alkalic plutons
Mi	Mesozoic felsic plutons
Mo	Mesozoic felsic orthogneisses

Quesnellia

Mu	Mesozoic rocks, undivided		
Jr	Jurassic Rossland Group		
	Jrh	Hall Formation, argillite	
	Jre	Elise Formation, mafic volcanic rocks	
	Jref	felsite	
TRb	Triassic Brooklyn Formation, argillite		
Pu	Carboniferous to Permian, western assemblage, Knob Hill and Attwood Groups		
	Pa	Attwood Group, undivided	
		Pap	pelitic phyllite
		Paa	argillite
		Paf	felsic volcanic rocks
		Pac	carbonate rocks
	Pk	Knob Hill Group, undivided	
		Pkg	Palmer Mountain greenstone
um	ultramafic rocks and listwanite		

Southwestern Metamorphic Belt

Mom	Mesozoic(?) megacrystic orthogneiss
hmg	Mesozoic(?) pelitic gneiss
hmp	Mesozoic(?) pelitic phyllite
swm	Mesozoic(?) paragneiss and orthogneiss

Eastern Assemblage

OCe	Ordovician to Carboniferous, undivided, mostly pelitic rocks, including Ordovician Covada Group

Metamorphic Core Complexes

Ei	Eocene felsic plutons
Mi	Mesozoic felsic plutons
hmt	Mesozoic Tonasket orthogneiss
hm	metamorphic rocks, undivided
hmm	Tenas Mary Creek assemblage, Cretaceous orthogneiss and Proterozoic and Paleozoic paragneiss

Metamorphic rocks

hmu	rocks of unknown age and origin

FAULTS

high-angle fault—dashed where approximate

detachment fault—blocks on upper plate

thrust fault—sawteeth on upper plate

FOLDS

upright antiform

overturned antiform

TOWNS

C, Colville	G, Greenwood	GF, Grand Forks
O, Omak	R, Republic	T, Trail

ABBREVIATIONS OF FAULT NAMES

BCF	Bacon Creek fault
BMF	Bodie Mountain fault
CMF	Cayuse Mountain fault
CLF	Columbia fault
CF	Chesaw fault
DMF	Dunn Mountain fault
EMF	Eagle Mountain fault
GDF	Gold Drop fault
GRF	Granby River fault
GWF	Greenwood fault
HRF	Huckleberry Ridge fault
KRF	Kettle River fault
LCF	Lind Creek fault
LMF	Lambert Creek fault
MAF	Mount Attwood fault
MCF	Myers Creek fault
MWF	Mount Wright fault
No. 7F	Number 7 fault
OF	Okanogan fault
OLF	Omak Lake fault
RF	Rossland fault
SCF	Salmon Creek fault
SF	Sherman fault
SKF	Scatter Creek fault
SLF	Slocan Lake fault
SNF	Snowshoe fault
TMF	Thimble Mountains fault
VS	Valkyr shear zone
WF	Waneta fault
WMF	White Mountain fault

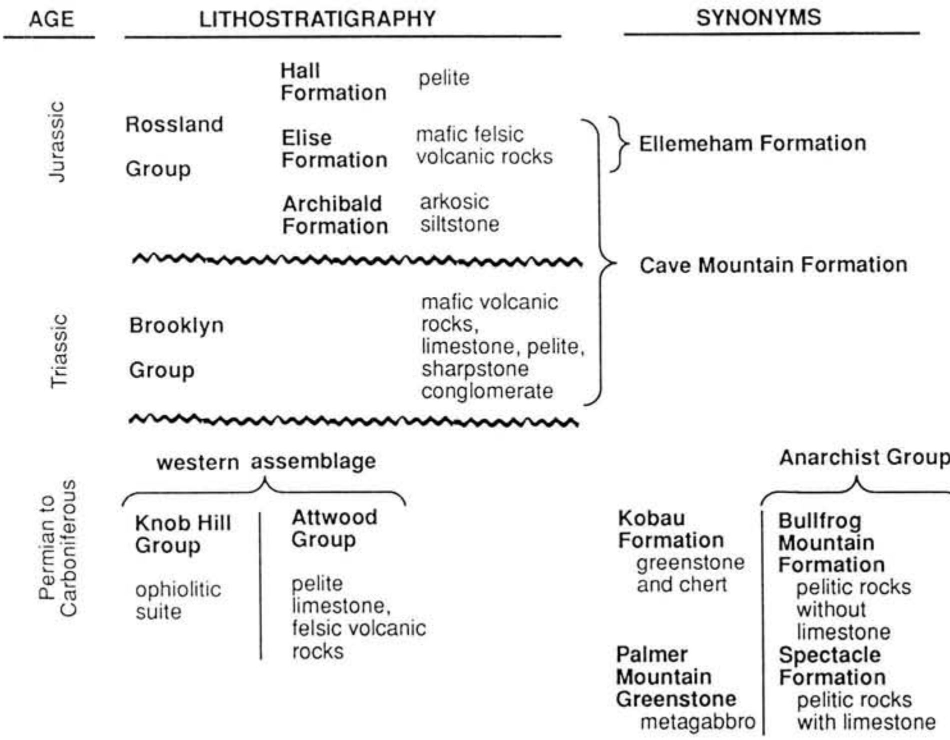

Figure 3. Regional unconformity-bounded sequences of southern Quesnellia. The horizontal wavy lines represent regional unconformities. The vertical lines emphasize that the Permian to Carboniferous units are largely coeval.

Proposed correlatives: Rinehart and Fox (1972) subdivided the pelitic Anarchist Group into a lower Spectacle Formation consisting of sharpstone conglomerate, limestone, graywacke, siltstone, and black slate and an upper Bullfrog Mountain Formation composed of similar rocks but without limestone. Hereafter, we refer to all the pelite-dominated Upper Paleozoic rocks in Washington as Attwood Group (Fig. 3) or, synonymously, pelitic Anarchist rocks.

Felsic metavolcanic and metavolcaniclastic rocks are under-reported in the Attwood Group, probably because most of the previous mapping was conducted before the geology of volcanic-hosted massive sulfide deposits was widely understood. Table 2 lists localities of such felsic rocks we have found in Washington; additional localities probably exist. East of Republic (Fig. 5), commercial gold deposits are associated with tabular bodies of magnetite and massive sulfide adjacent to felsic metavolcaniclastic rocks at the Lamefoot and Overlook mines. Both pelitic and felsic volcaniclastic rocks at Overlook contain graded beds; these provide some of the best evidence that part of the section at Overlook is overturned (Fig. 5, cross section). The largest belt of felsic rocks we have mapped to date is 5 km long, from Hicks Canyon to Wannacut Lake (loc. 2 on Fig. 6). Rinehart and Fox (1972) mapped the felsic rocks of this belt as both Spectacle Formation and Bullfrog Mountain Formation.

Knob Hill Group

Type area: The type area for the Knob Hill Group is Knob Hill (Little, 1983), about a kilometer southwest of the Phoenix mine (loc. 1 on Fig. 4). Although Little (1983) suggested that serpentinites associated with the Knob Hill Group are parts of a disrupted ophiolite, Fyles (1990) was the first to suggest that the entire Knob Hill Group is a disrupted ophiolitic suite. According to Fyles (1990), the Knob Hill Group has two distinct facies of deformation; the less deformed part is north of the Lind Creek fault (Fig. 4). Fyles (1990) noted that the Carboniferous to Permian age for the Knob Hill is based on a single fossil locality described by Little (1983).

Although the Knob Hill Group consists of six major lithologies, only serpentinite is shown separately (as unit um) in Figure 4. The following description of the six lithologies is from Fyles (1990), except where otherwise noted.

(a) Serpentinites are sheet-like bodies along faults. Some serpentinites have relict pyroxene; others are altered to listwanite.

(b) Old Diorite (Church, 1986) is coarse-grained, texturally varied hornblende diorite. The hornblende is commonly partly chloritized, and the rock is cut by numerous crisscrossing veinlets of felsic minerals, calcite, and epidote (Church, 1986). Bodies of Old Diorite as much as 4 km long are associated with serpentinite and less deformed parts of the Knob Hill Group in the hanging wall of the Lind Creek fault. Smaller bodies are associated with more deformed rocks along the Mount Wright and Mount Attwood faults. Church (1986) reported a K-Ar age of 258 ± 10 Ma for one sample of Old Diorite.

(c) Greenstones are aphanitic, basaltic to andesitic pillowed lavas and derivative fragmental rocks. These are part of the less deformed portion of the Knob Hill Group.

(d) Gray to buff chert, generally without ribbon-like bedding, commonly is highly fractured. Associated rocks are black argillite and rare limestone.

(e) Buff to gray sharpstone and pebble conglomerates are mostly composed of chert fragments. These rocks and chert-dominated, fine-grained sedimentary rocks are associated with the less deformed greenstones.

(f) The more deformed part consists of foliated rocks otherwise similar to Knob Hill lithologies.

Proposed correlatives: A suite of greenstone and chert very similar to the Knob Hill Formation occurs in Washington in the Okanogan Valley (Fig. 6). Rinehart and Fox (1972) believed that the greenstones were unconformable upon the Anarchist pelitic rocks and were, therefore, Triassic. Rinehart and Fox (1972) named these rocks above the Permian Anarchist pelitic rocks the Kobau Formation because west of Osoyoos Lake they are continuous with the Kobau Group of Bostock (1940) in British Columbia. Yet Okulitch (1973) maintained that the Kobau Group is Paleozoic, and he noted that near Palmer Lake (Fig. 6) the Kobau Formation is as complexly deformed as the Kobau Group of British Columbia and more complexly deformed than nearby parts of the Anarchist assemblage. The Permian–Triassic deformation in this region would (logically) result in any greenstones unconformably above the Anarchist Group being less complexly deformed than the Anarchist Group (Read and Okulitch, 1977). Recognition of the contact between the pelites and greenstones as the Chesaw thrust resolves the problem and implies that the Kobau, or Knob Hill Group, could be older than the pelitic rocks.

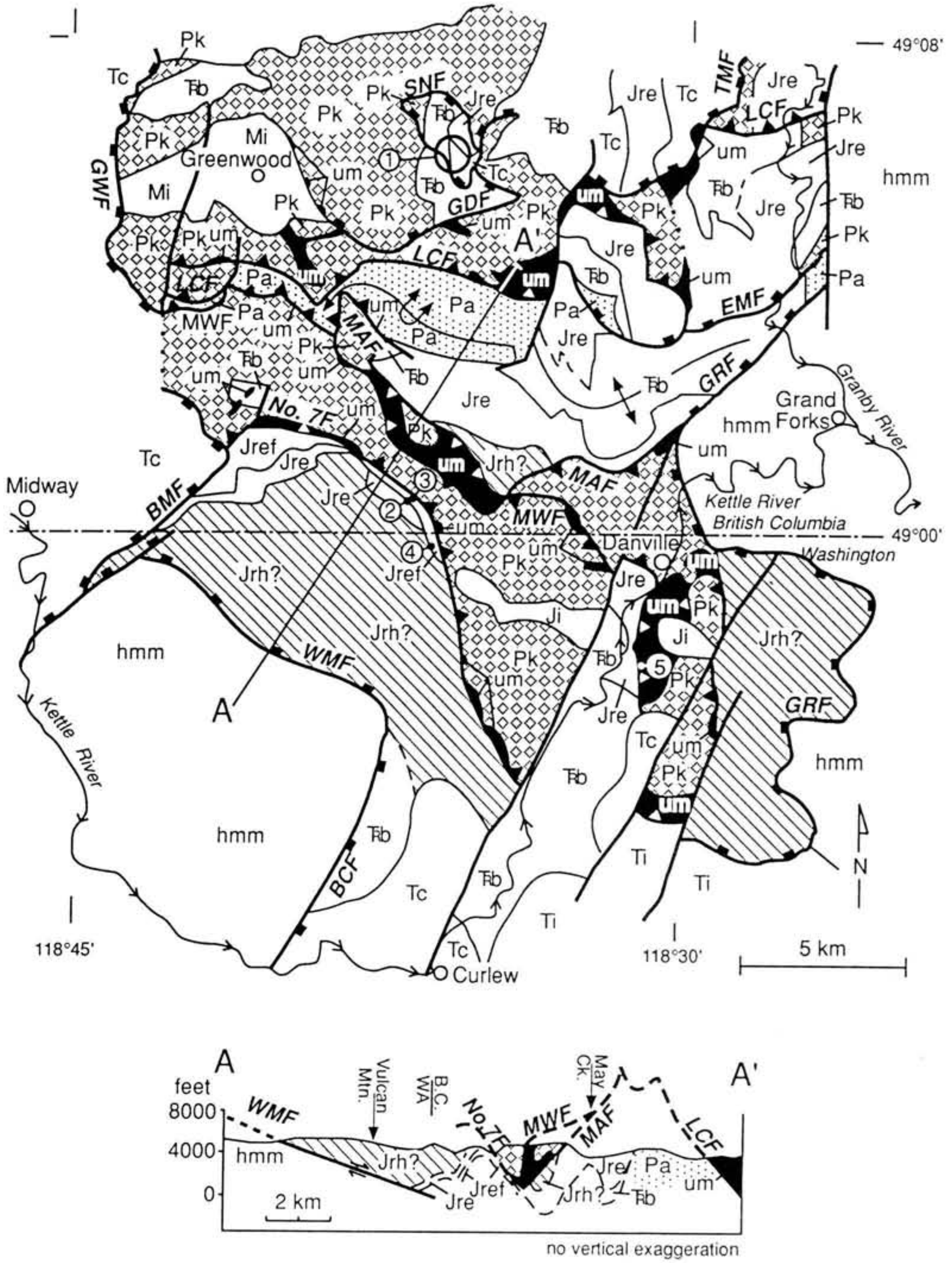

Figure 4. Geologic map of the Greenwood–Curlew area. See Table 1 for explanation of most units; Ti, Tertiary intrusives. Sources for this compilation for British Columbia are Little (1983), Church (1986), and Fyles (1990); sources for Washington are Parker and Calkins (1964), Pearson (1977), and Herdrick and Bunning (1984). The datum for the cross section is mean sea level. The cross section is published with the permission of Britannia Gold Corp.

The greenstones in Washington also appear to be an ophiolitic suite. The most abundant rock type is variously phyllitic, aphanitic greenstone with interbedded quartzite (metachert) and rare argillite, the Kobau Formation of Rinehart and Fox (1972). Serpentinites, and especially listwanites, are discontinuous bodies (Fox and Rinehart, 1968; Rinehart and Fox, 1972) at or near the base of the greenstones. Localities of some of these are shown as unit um in Figures 6 and 7.

The Palmer Mountain greenstone (Rinehart and Fox, 1972), like the phyllitic greenstone, is structurally above the pelitic Anarchist rocks (Rinehart and Fox, 1972). The greenstone ranges from poorly foliated and aphanitic to

Table 2. Localities of Quesnellian felsic volcanic rocks

Name Area	Location Figure, number	Stratigraphic unit	Mineralization	Reference
Lone Star mine Danville	Sec. 2, T40N, R33E Fig. 4, #4	Elise Formation	Disseminated pyrite and chalcopyrite	Parker and Calkins, 1964, p. 86
Morning Star mine Danville	sec. 16, T40N, R34E Fig. 4, #5	Elise Formation	rare massive pyrite	Parker and Calkins, 1962, p. 88
City of Paris mine Boundary	1 km NW of Lone Star Fig. 4, #3	Elise Formation	disseminated pyrite and chalcopyrite	Church, 1986, p. 33
Lexington mine Boundary	1 km NW of Lone Star Fig. 4, #2	Elise Formation	disseminated pyrite and chalcopyrite	Church, 1986, p. 33
Overlook mine Cooke Mountain	sec. 18, T37N, R34E Fig. 5, #1	Attwood Group	massive pyrrhotite and magnetite	Huntting, 1956, p. 123, 195
Lamefoot mine Curlew Lake	sec 9, T37N, R33E Fig. 5, #2	Attwood Group	massive pyrite and magnetite	Huntting, 1956, p. 195
Copper Mountain Republic	sec. 15, T36N, R32E none	Attwood Group	massive pyrite and magnetite	none
Hot Lake Oroville	sec. 7, T40N, R27E Fig. 6, #1	Attwood Group	massive pyrite and pyrrhotite	none
Hicks Canyon Wannacut Lake	sec. 35, T40N, R26E Fig. 6, #2	Attwood Group	disseminated pyrite	none
Silver Mountain mine Horse Springs Coulee	sec. 34, T37N, R26E Fig. 6, #3	Attwood Group	disseminated pyrite	Huntting, 1956, p. 310
Lemansky Lake	sec. 13, T37N, R25E Fig. 6, #4	Attwood Group	none; possibly a dike	none
Buckhorn Mountain Chesaw	SE part T40N, R29E Fig. 7	Attwood Group	disseminated pyrite	McMillen, 1979, p. 50

medium-grained, green (chlorite and epidote) metadiorite and metagabbro. Dunites (Rinehart and Fox, 1972) and chromitite bands (Huntting, 1956) occur west of Palmer Lake. We suggest that the Palmer Mountain greenstone is lithologically similar to the Old Diorite of the Greenwood–Grand Forks area.

Brooklyn Formation

Type area: In Quesnellia and adjacent terranes, Triassic rocks commonly overlie an unconformity (Read and Okulitch, 1977). In the Greenwood–Grand Forks area (Fig. 4), the Triassic Brooklyn Formation unconformably overlies both the Attwood Group and the Knob Hill Group (Fyles, 1990). Thus, if the Attwood Group and Knob Hill Group were once separate terranes, they were amalgamated in pre-Brooklyn time. However, because segments of the Chesaw fault cut the Brooklyn Formation, this amalgamation must pre-date the Chesaw fault. The following description of the Brooklyn Formation is from Fyles (1990) for the type area near Greenwood.

The Brooklyn Formation may be as much as 2 km thick. The main lithologies, beginning with basal sharpstone conglomerate and grading laterally and upward through finer grained clastic rocks into limestone and (or) greenstones, are repeated one or more times (column B, Fig. 8). The greenstones indicate that the Brooklyn Formation is in greenschist facies. The basal sharpstone conglomerate contains rounded to angular fragments of gray, buff, green, or maroon chert, white and purple quartz, greenstone, limestone, chlorite schist, and quartz-biotite gneiss. Clasts of several Knob Hill lithologies are present (Church, 1986; J. T. Fyles, B.C. Geol. Survey Branch, written commun., 1992). Minor rhyolitic tuff and green tuffaceous sandstone lenses occur in the sharpstone conglomerate. Fine-grained rocks consist of green to brown sandstone, and massive gray, green, maroon or black siltstone with disseminated pyrite or pyrrhotite. Locally a conglomerate of limestone clasts in a matrix of maroon tuffaceous siltstone and sandstone disconformably overlies the basal sharpstone conglomerate; the clasts contain Carboniferous or Permian conodonts.

Limestones vary from dark-gray, flaggy, and argillaceous to massive gray and white to heterogeneous limestone conglomerates. Poorly preserved megafossils in Brooklyn limestones appear to be mostly middle Triassic (Ladinian) (Little, 1983), but Little's unit uTsv (which contains Norian corals) is now included by Church (1986) and Fyles (1990) in the Brooklyn Group. However, Fyles (1990; written commun., 1992) reported that all the conodonts are Middle Triassic (Ladinian).

Proposed correlatives: Outliers of the Brooklyn Formation are known in Washington. Parker and Calkins (1964) described Triassic limestones north of Curlew (Fig. 4), which Little (1983) correlated with his unit uTsv. Pearson (1967), Read and Okulitch (1977), and McMillen (1979) recognized Brooklyn sharpstone conglomerate and (or) limestone conglomerate northeast of Buckhorn Mountain near Chesaw (Fig. 7). A kilometer southeast of Chesaw

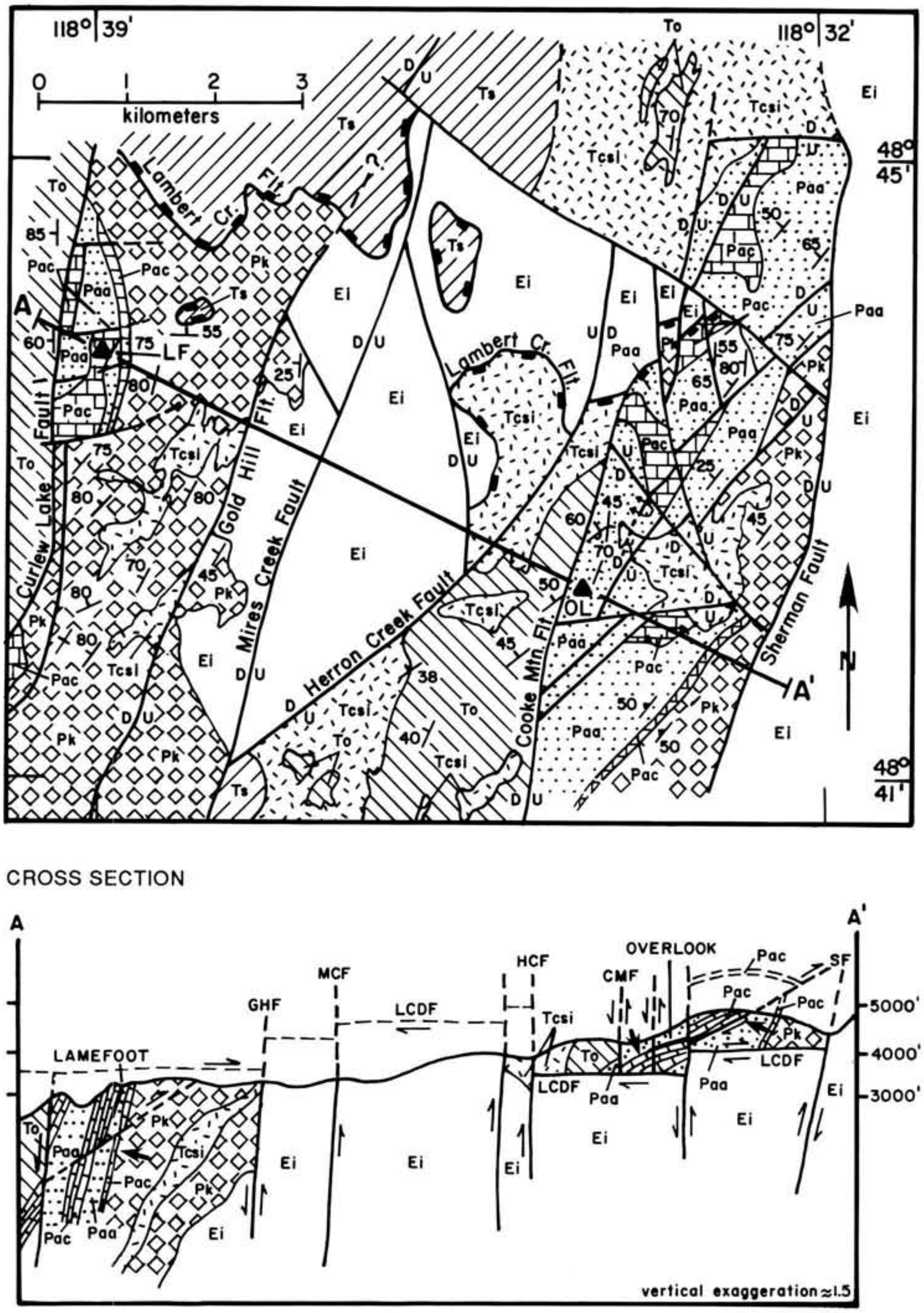

Figure 5. Geology of the Overlook–Lamefoot area. See Table 1 for explanation of most units. Sources of data: Muessig (1967) and unpublished mapping by M. G. Rasmussen, 1990–1992. On map, mineral deposits are: OL, Overlook; LF, Lamefoot. Map units not given in Table 1 are To, O'Brien Creek Formation; Ts, Sanpoil Volcanics. Geologic cross section A–A′ includes faults not listed in Table 1: GHF, Gold Hill fault; MCF, Mires Creek fault; LCDF, Lambert Creek detachment fault; HCF, Herron Creek fault; and CMF, Cooke Mountain fault. Bold arrows indicate stratigraphic facing directions of graded beds.

(Fig. 7), McMillen mapped sharpstone conglomerate (1979, pl. 9C) and a stretched conglomerate of limestone (1979, pl. 9B), both of which he assigned to the Brooklyn Formation on his geologic map. A conglomerate of unknown age with clasts of limestone and greenstone in an aphanitic matrix (Muessig, 1967) occurs a kilometer north of the Lamefoot prospect of Figure 5.

Rossland Group

In the Rossland area east of the Kettle metamorphic core complex, the Jurassic Rossland Group consists of three parts (Little, 1982; Höy and Andrew, 1989): a lower Archibald Formation, a middle Elise Formation, consisting predominantly of greenstone, and an upper, weakly deformed succession of clastic rocks, the Hall Formation. Sinemurian fossils (206–200 Ma) occur in the Archibald Formation, whereas fossils as young as middle Bajocian (>175 Ma) occur in the Hall Formation (Little, 1982). Heretofore, neither the Archibald Formation nor the Hall Formation have been recognized west of the longitude at which the Columbia River crosses the international boundary.

Elise Formation

Canadian occurrences: Rocks of the Elise Formation are andesitic to basaltic (Little, 1982). Although the rocks are unfoliated, chlorite and epidote are abundant. Fragmental textures are common. Flows are feldspar and augite or hornblende porphyries. The basal part of the formation contains ellipsoidal fragments of limestone from the underlying Mount Roberts Formation. Bedded tuffs and pelitic rocks are interbedded with the massive and fragmental rocks. Sill-like intrusive bodies are widespread and are characterized by stout prisms of pyroxene or hornblende as long as 6 mm.

In the Greenwood–Grand Forks area (column B, Fig. 8), massive to fragmental greenstones and pyroclastic rocks unconformably overlie the clastic and carbonate rocks of the Brooklyn Formation (Fyles, 1990, fig. 3). Little (1983) and Tempelman-Kluit (1989) included these greenstones in the Rossland Group, but several greenstones have lenses of limestone with Ladinian conodonts (Fyles, 1990;

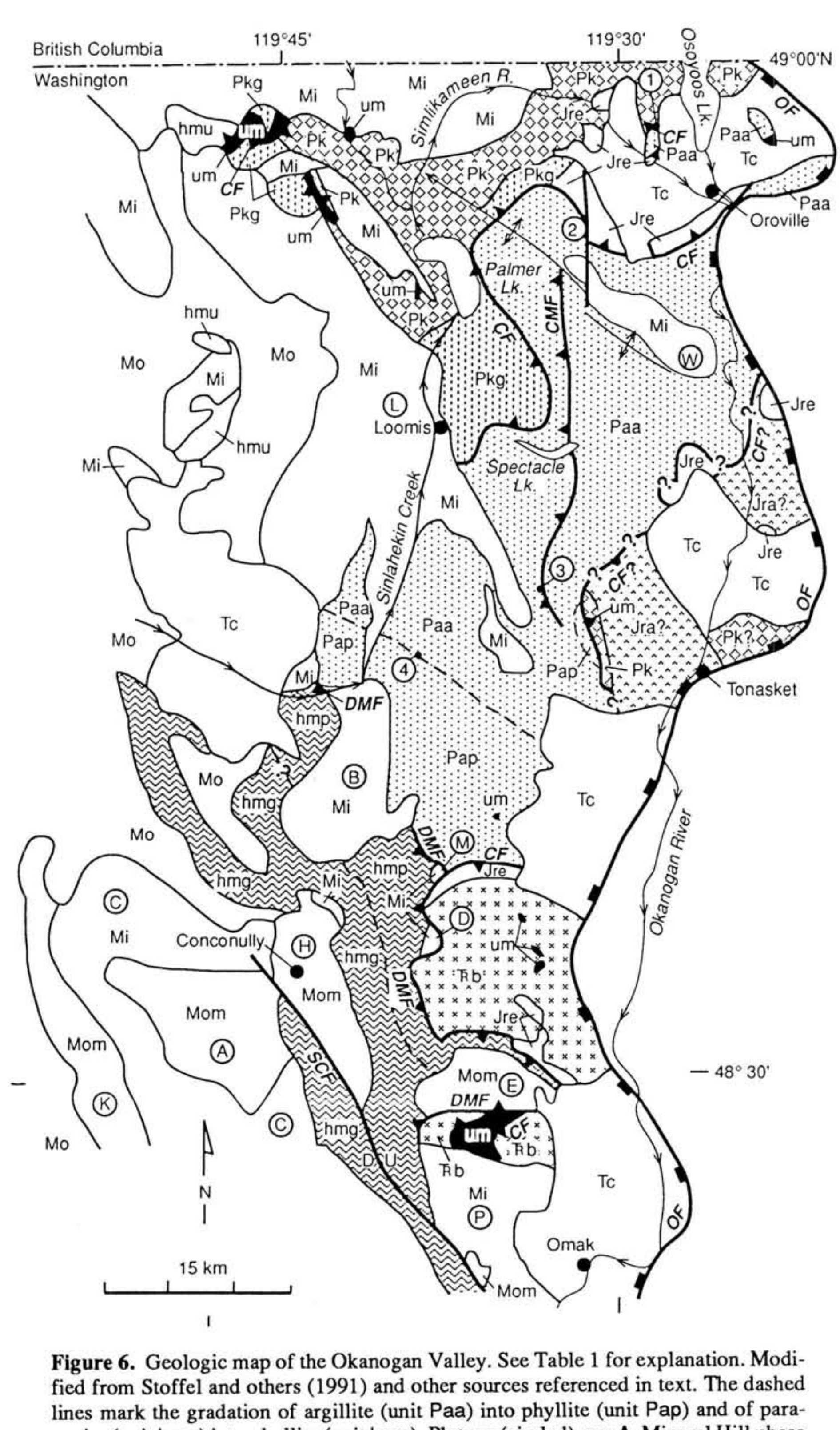

Figure 6. Geologic map of the Okanogan Valley. See Table 1 for explanation. Modified from Stoffel and others (1991) and other sources referenced in text. The dashed lines mark the gradation of argillite (unit Paa) into phyllite (unit Pap) and of paragneiss (unit hmg) into phyllite (unit hmp). Plutons (circled) are: A, Mineral Hill phase of the Conconully pluton; B, Blue Goat; C, Conconully; D, Dunn Mountain; E, Evans Lake; H, Happy Hill; K, Leader Mountain; L, Loomis; M, Mud Lake; P, Pogue Mountain; and W, Whiskey Mountain. Localities (circled) with felsic metavolcanic rocks are, 1, Hot Lake; 2, Hicks Canyon; 3, Silver Mountain mine; and 4, Lemansky.

written commun., 1992). Thus, unless the conodonts are reworked, some of the greenstones are, as Fyles (1990) concluded, part of the Brooklyn Formation.

We cannot yet distinguish Brooklyn fragmental greenstones from Rossland fragmental (Elise Formation) ones. In Figures 4, 6, and 7 and Table 1, we have labeled all greenstones with meta-igneous or fragmental textures "Jre" with the realization that some may be Brooklyn Formation. Our point is that most of the greenstones with original textures, whether Brooklyn or Elise, occur above the clastic and most of the carbonate units of the Brooklyn Formation (Fyles, 1990, fig. 3) and thereby provide a third unit to use in the regional interpretation of maps.

The Elise Formation may contain felsic metavolcanic rock. The variously foliated Lexington quartz porphyry in the footwall of the No. 7 fault (unit Jref in Fig. 4) has been considered a felsic intrusion (Little, 1983; Church, 1986, 1992; Fyles, 1990). In its western portion, the porphyry is "coarse-grained, irregular, and transgressive and clearly is an intrusion" (J. T. Fyles, written commun., 1992). A U-Pb age on zircons from the Lexington is 199.4 ± 1.4 Ma (Church, 1992), or Rossland (not Brooklyn) in age. Parker and Calkins (1964) and Peatfield (1978) suggested that the quartz porphyry might be a felsic metavolcanic rock; its age, apparently concordant map pattern with the overlying greenstone, and mineralization (Table 2) at the Lexington deposit of Britannia Gold Corp., City of Paris, and Lone Star mines support this suggestion. Perhaps the Lexington porphyry is intrusive to the west and extrusive to the east.

According to Lambert (1989), the mafic Elise Formation has the overall chemistry and isotopic composition of oceanic rocks without significant continental contamination. Thus Lambert included the Rossland Group in Quesnellia. The granodioritic Nelson plutons intrude both the Rossland Group and North American sequences (Andrew and others, 1991) and yield zircon ages of 164 ± 1 Ma and 169 ± 2 Ma (Lambert, 1989).

Possible correlative: In the Okanogan valley from Tonasket northward, Rinehart and Fox (1972) described several localities of unfoliated, fragmental greenstones resting unconformably on the western assemblage. They named these the Ellemeham Formation. Read and Okulitch (1977) and Stoffel and others (1991) correlated the Ellemeham Formation with the Elise Formation. However, the Ellemeham Formation is undated paleontologically or radiometrically. It is shown as unit Jre in Figure 6.

Stratigraphic Revisions

Brooklyn Formation

We infer that the Cave Mountain Formation described by Rinehart and Fox (1976) northwest of Omak (Fig. 6) is part of the Brooklyn Formation and, probably, the Rossland Group. Column A of Figure 8 compares the partially dolomitized section at Cave Mountain with a section (column B) of Brooklyn Formation described by Fyles (1990, fig. 3). Rinehart and Fox (1976, pl. 1, sections D-D′, E-E′, F-F′) inferred that two or three different units of the Cave Mountain succession overlie the sharpstone conglomerate. This only can be shown schematically in column A of Figure 8 because the exact stratigraphic and facies relations of the three units are unknown. If two or three units do overlie the sharpstone conglomerate, facies changes in the Cave Mountain succession may be as abrupt as Fyles (1990, fig. 3) illustrated for the Brooklyn Formation.

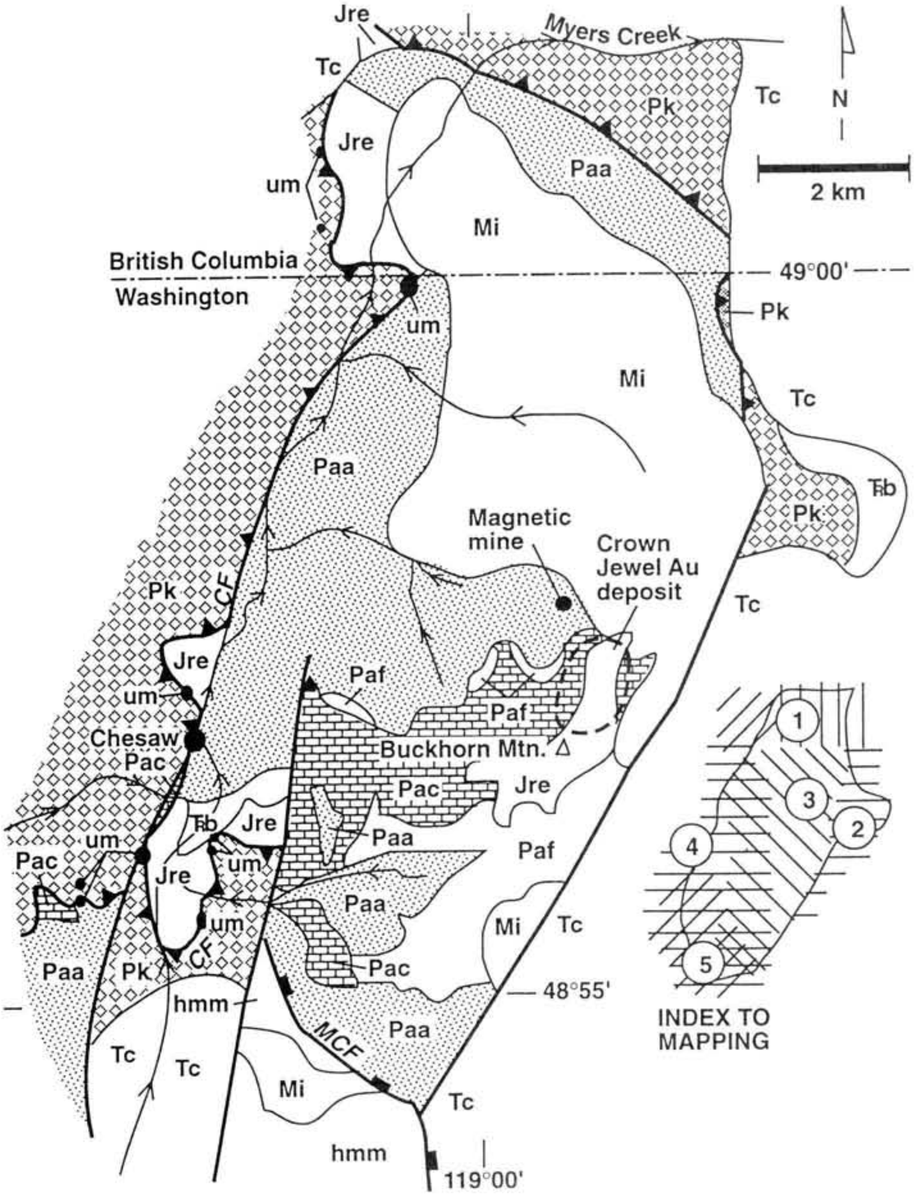

Figure 7. Geologic map of the Chesaw area. See Table 1 for explanation. Sources of data are: 1, Fyles, (1990); 2, Pearson (1967); 3, McMillen (1979); 4, Fox (1978); and 5, unpublished mapping by E. S. Cheney (1991, 1992).

The basal unit of the Brooklyn Formation is a sharpstone conglomerate dominated by angular clasts of chert (Fyles, 1990). A similar conglomerate occurs beneath the carbonate rocks of the Cave Mountain succession (Rinehart and Fox, 1976). Although Waters and Krauskopf (1941) implied that the conglomerate near Omak is conformable with the overlying carbonate rocks, Rinehart and Fox (1976) regarded the contact as an unconformity, perhaps because they found clasts of the conglomerate in the lowermost carbonate rocks. Thus, Rinehart and Fox (1976) assigned the conglomerate to the Anarchist Group.

Megafossils in the uppermost carbonate unit of the Cave Mountain succession are Carnian to Norian (Misch, 1966; Rinehart and Fox, 1976), as are fossils in the Brooklyn Formation (Church, 1986; Fyles, 1990). Thus, even if the sharpstone conglomerate at the base of the Cave Mountain succession is Anarchist, the overlying carbonates correlate with the Brooklyn Formation.

The upper basaltic member of the Cave Mountain succession overlies three units of the Cave Mountain succession (Rinehart and Fox, 1976, pl. 1); thus, the contact is either an unconformity or a fault. The basalt has equant megacrysts of hornblende (Rinehart and Fox, 1976), as do some of the greenstones in the Greenwood–Grand Forks area (Little, 1983).

Column C of Figure 8 shows the interpretations offered here. The lithostratigraphic succession (sharpstone con-

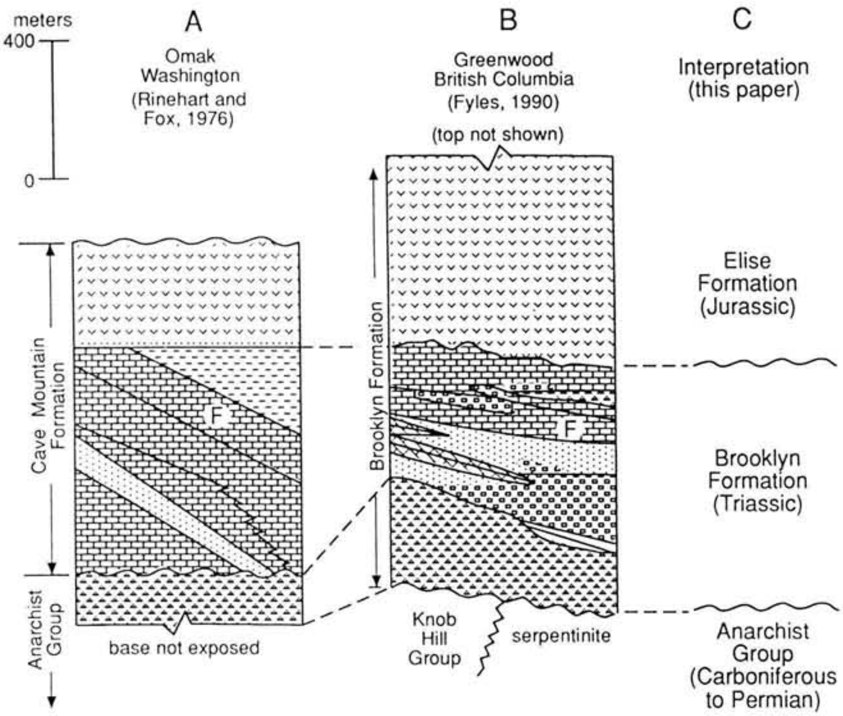

Figure 8. Comparison of the Brooklyn Formation with the Cave Mountain succession. As explained in the text, column A is a schematic representation of the Cave Mountain succession near Omak, WA, from the descriptions of Rinehart and Fox (1976). Rinehart and Fox (1976) placed the sharpstone conglomerate in the Anarchist Group unconformably below the carbonate units. They did not recognize an unconformity below the metabasalt. Column B is from Fyles (1990, column 6 of fig. 3). v's, metabasalt; triangles, sharpstone conglomerate dominated by fragments of chert; bricks, limestone; tilted bricks, dolomite; dots, sandstone and siltstone; boxes, conglomerate with limestone clasts; dashes, argillite; parallel lines, thin-bedded or massive siltstone. F designates the location of Carnian or Norian megafossils as mapped by Rinehart and Fox (1976) and Fyles (1990).

glomerate, carbonate rocks, and metabasalt) and the fossils in the carbonate rocks at Omak and Greenwood correlate. If an unconformity does exist above the sharpstone conglomerate near Omak, it may be an intraformational (or intrasequence) disconformity, such as those described by Fyles (1990) in the Brooklyn Formation. The metabasaltic rocks could be either Brooklyn or Rossland but are shown in column C and in Figures 4 and 6 as the Elise Formation of the Rossland Group.

Brooklyn or Elise greenstones

At Chesaw (Fig. 7), green Knob Hill phyllite structurally overlies unfoliated greenstones, some of which contain stubby mafic phenocrysts. Thus these unfoliated greenstones are shown as unit Jre in Figure 7, but they were mapped as Kobau Formation (western assemblage) by Fox (1978). This interval of unfoliated greenstone accounts for the otherwise curious fact that listwanites, which mark the Chesaw thrust, occur 35–50 m above the base of the greenstones. Detailed mapping might show that Brooklyn or Elise rocks occur structurally beneath Knob Hill rocks elsewhere between Chesaw and the Okanogan valley.

Unfoliated greenstone on Buckhorn Mountain (Fig. 7) was thought to be a Kobau or Knob Hill equivalent (Fox, 1978; McMillen, 1979; Hickey, 1992) in thrust contact with Attwood rocks (McMillen, 1979; Orr and Cheney, 1987; Stoffel and others, 1991). However, for the following reasons, these greenstones most likely are Brooklyn or Elise Formation unconformable upon the Attwood Formation:

(a) The greenstones are not phyllitic. Instead, they contain pillows, volcaniclastic textures, dacitic to rhyolite tuffs, and pelitic interbeds (McMillen, 1979; Hickey, 1992).

(b) Stubby hornblende or pyroxene phenocrysts are common in non-fragmental rocks (David Jones, Battle Mountain Gold Corp., oral commun., 1991).

(c) Geologists of the Battle Mountain Gold Corp. have shown us that in drill core no fault zone exists between the greenstones and underlying marbles.

(d) The greenstones also occur above a pelitic unit containing chert-pebble and sharpstone conglomerate (Hickey, 1992).

Archibald and Hall Formations

To test whether the Archibald and Hall Formations exist west of the Kettle metamorphic core complex, the following descriptions from Little (1982) and Höy and Andrew (1989) are germane. The Archibald Formation is hard, brittle, dark-gray to black argillaceous siltstone and arenaceous argillite with minor quartzite and graywacke. Beds are laminated, and graded bedding is common. Clasts in siltstone consist of plagioclase, calcite, quartz, K-feldspar, chloritized mafic minerals, magnetite, and hematite. Because in the Rossland area, the Archibald Formation overlies the Pennsylvanian Mount Roberts Formation (Little, 1982), this contact is an unconformity. In places, the Archibald Formation grades up into the Elise Formation, but locally the Elise is also unconformable on the Mount Roberts Formation (Andrew and others, 1990).

The Hall Formation is at least 1,400 m thick. It can normally be distinguished from the Archibald Formation by its softness and fissility. The Hall Formation is predominantly black, carbonaceous shale and buff to brown argillaceous sandstone. Some siltstone and minor graywacke are present. The basal conglomerate has pebbles of volcanic rocks similar to those in the Elise Formation, and the graywackes have 10–50 percent quartz, feldspar, mafic minerals, and fragments of volcanic and sedimentary rocks (Mulligan, 1952). The middle 300 m of the formation consists of coarse sandstone, grit, and pebble conglomerate (Höy and Andrew, 1989).

Rinehart and Fox (1976) included pelitic rocks 0–10 km west of Tonasket (Fig. 6) in the Anarchist group. Stoffel (1990b) and Stoffel and others (1991) suggested they might be Covada Group (eastern assemblage), an assignment that K. F. Fox, Jr. (written commun., 1993) favors. The following suggest that these rocks may be Archibald Formation:

(a) They occur structurally above well-foliated greenstone (phyllite) with fuchsite-bearing magnesite at its base (Rinehart and Fox, 1976, pl. 1). These greenstones and magnesite probably are Knob Hill Group.

(b) The pelitic rocks are spatially associated with volcaniclastic rocks of the Ellemeham Formation (which may be part of the Rossland Group).

(c) These pelites do not contain sharpstone conglomerate.

(d) They do contain metawacke, meta-arkose (with angular clasts of quartzite and granite), and phyllites with angular grains of quartz, albite, microcline, and muscovite (Rinehart and Fox, 1976; Stoffel, 1990b). This mineralogy could fit the Covada Group, but not the Anarchist Group (Fox and Rinehart, 1974).

(e) They are in the same structural low (adjacent to the western margin of the Okanogan metamorphic core complex) that preserves other comparatively young rocks (the Cave Mountain succession, Ellemeham Formation, and Challis sequence).

A possible area of Hall Formation occurs between the No. 7 fault and the White Mountain fault along the international boundary south of Greenwood, BC (unit Jrh of Fig. 4). Orr (1985) thought that these rocks were part of the Okanogan metamorphic core complex, whereas Fyles (1990) assumed that they are Attwood Group. They could be Hall Formation because:

(a) They occur structurally above unfoliated greenstone (Little, 1983; Orr, 1985; Church, 1986; Fyles, 1990), probably Elise Formation (or a Rossland sill), that is associated with the 199 Ma Lexington porphyry.

(b) Both Little (1983) and Orr (1985) commented on the abundance of quartz grains (and lack of feldspar) in these pelitic rocks.

(c) These pelitic rocks are adjacent to a detachment fault (just as young formations in Figure 6 are adjacent to the Okanogan detachment fault).

Southwestern Metamorphic Belt

Amphibolite-facies paragneisses and orthogneisses bound the Quesnellian rocks on the southwest (Figs. 2 and 6). The age of these rocks is unknown; they are intruded by the Conconully pluton, which, as noted below, is most likely about 82 Ma.

The paragneisses southwest of the Salmon Creek fault (Fig. 6) were originally named the Salmon Creek Schists and Gneisses (Menzer, 1964, 1983). Rinehart and Fox (1976) included amphibolite-facies paragneisses on both sides of the Salmon Creek fault in their informally named metamorphic complex of Conconully; they excluded the orthogneisses from this unit. Sims (1984) mapped metasedimentary rocks northeast of the fault as Salmon Creek Schist and Gneiss. Stoffel (1990b) used "granodioritic gneiss of Salmon Creek" for some of the megacrystic orthogneisses and left the metasedimentary rocks unnamed. Unfortunately, the name "Salmon Creek schists and gneisses" now seems to be pre-empted for rocks having a sedimentary protolith (Stoffel and others, 1991). Rather than propose a new name, we informally refer to all of these amphibolite-facies rocks as the southwestern metamorphic belt.

Following Menzer (1983), we exclude from this designation the trondhjemitic gneisses and unfoliated plutons that underlie most of the Okanogan Range west of Conconully and Salmon Creek. Our usage differs from the "metamorphic complex of the Conconully" of Rinehart and Fox (1976) in two aspects. First, we include the orthogneisses in the southwestern metamorphic belt. Second, Rinehart and Fox (1976) believed that the metamorphic complex of Conconully grades from schist and gneiss into greenschist-facies rocks similar to those in the Anarchist Group. We show in the following section of this paper that the amphibolite-facies metamorphic rocks are separated from the greenschist-facies rocks of the western assemblage by the Dunn Mountain fault. Thus the unit Pap in Figure 6, which consists predominantly of pelitic rocks of greenschist facies and which Rinehart and Fox (1976) included in the metamorphic complex of Conconully, we now assign to the western assemblage.

The most distinctive rock in the southwestern metamorphic belt is the megacrystic K-feldspar Leader Mountain orthogneiss of Menzer (1964, 1983). It is labeled Mom in Figure 6. This orthogneiss was mapped by Rinehart and Fox (1976) and by Sims (1984) as the porphyritic phase of the Evans Lake pluton (unit Mom of pluton E of Fig. 6). Figure 6 shows that this gneiss also has been mapped as a phase of the Happy Hill pluton (Rinehart and Fox, 1976), the Mineral Hill phase of the Conconully pluton (Goldsmith, 1952; Menzer, 1964, 1983), a small body adjacent to the southwest margin of the Pogue Mountain pluton (Sims, 1984), the Leader Mountain orthogneiss along the western contact of the Conconully pluton (Menzer, 1964; 1983), and bodies in the pelitic gneisses northwest of the Conconully pluton (Goldsmith, 1952). The descriptions given by the above authors indicate that at each of these localities the orthogneiss has at least five of the following characteristics: it is weakly to well foliated; the K-feldspar megacrysts are as much as 5 cm long; microscopically, these megacrysts are poikiloblastic (sieved) and have irregular margins with the finer grained matrix; biotite is more abundant than hornblende; the hornblende is ferrohastingsite; myrmekitic intergrowths are common; and allanite occurs in more than trace amounts.

Like the pelitic gneisses, bodies of this megacrystic orthogneiss occur on both sides of the Salmon Creek fault. Compositionally similar but non-megacrystic orthogneisses also occur southwest of the Salmon Creek fault (Goldsmith, 1952; Menzer, 1964, 1983) and northeast of the fault in the Evans Lake and Happy Hill plutons. None of these orthogneisses, and certainly not the distinctive megacrystic orthogneiss, intrudes the western assemblage.

STRUCTURE

Recumbent folds are known in all units. Faults are less well known. We found that mappable zones of slaty to phyllitic rocks in the western assemblage mark the larger faults. In addition, the Chesaw fault in the Quesnellian rocks is locally marked by ultramafic rocks. The Dunn Mountain fault places amphibolite-facies rocks against greenschist-facies rocks.

Folding

Map-scale refolded recumbent folds are known in the Kobau Group (Okulitch, 1973). Recumbent folds have been mapped in the Brooklyn Formation (Church, 1986) and in the Cave Mountain succession (Rinehart and Fox, 1976). Rinehart and Fox (1972, figs. 27 and 28) illustrated outcrop-scale recumbent folds in Anarchist pelitic rocks. Outcrop-scale recumbent folds also occur in paragneiss of the southwestern metamorphic belt southwest of the Dunn Mountain pluton.

Figure 5 shows the geology east of Republic. Graded turbidites, graded felsic volcaniclastic rocks, sharpstone conglomerates, and limestone in underground exposures and drill cores in the Overlook mine (Fig. 5, cross section) define a west-dipping fold in the western assemblage. Southeast of the mine, greenstone is overlain, in apparent conformity, by a succession dominated by pelitic rocks. This is the only locality presently known in the region that establishes the relative ages of the Knob Hill and Attwood Groups.

Chesaw Fault

Chesaw area

The Chesaw thrust places the ophiolitic Knob Hill Group over the pelite-dominated Attwood Group. Less than a kilometer southeast of Chesaw, stretched pelitic conglomerate and stretched limestone conglomerate (shown as Trb in Fig. 7) dip southerly to southwesterly, are well foliated (McMillen, 1979), have southeasterly trending lineations, and occur structurally below unfoliated greenstones (unit Jre). Phyllitic clasts in the pelitic conglomerate and clasts in siliceous limestone conglomerate are elongated more than 2:1 parallel to the southeastward (N125°) lineation. These rocks (and included clasts) are so flattened that McMillen (1979) placed the Chesaw thrust below them (not above them and the unfoliated greenstones at the level of the listwanites as in Fig. 7). The simplest explanation for the deformation of these rocks is that a local splay of the Chesaw fault floors them (and has exploited the unconformity between them and the Anarchist pelitic rocks). The interpretation of Fox (1978) is that the limestone conglomerate and unfoliated greenstone are the basal part of the Kobau Formation unconformable upon the pelites of the Anarchist Group.

Preliminary evidence for the easterly direction of transport of the Chesaw fault comes from a listwanite 9.5 km west of Chesaw and 1.5 km south of Molson Hill (loc. I of Fox and Rinehart, 1968). The listwanite is crudely foliated (N295°, dip 25 degrees northeast) and cut by quartz veinlets, most of which are 1–10 mm in width. One set of quartz veinlets strikes N025° and dips 70 degrees northwest; a second set is perpendicular to the first set and subparallel to the foliation. These two sets of veinlets show mutually cross-cutting relations, with the sense of offset of the high-angle veinlets being top to the east. Quartz fibers are perpendicular to the walls of the high-angle veinlets; these fibers are subparallel to stretching lineations that plunge N110° at 21 degrees in the low-angle veinlets.

In thin section this listwanite displays a prominent pressure-solution foliation. Crudely concordant quartz segregations show abundant evidence for dynamic recrystallization, including the development of subgrains and undulatory extinction (Fig. 9A). Combined with the foliation, the low-angle quartz veinlets impart a gross fabric to the rock suggestive of S-C fabric in quartzo-feldspathic mylonites (Simpson, 1986). This fabric suggests transport of the top to the east-southeast. Additionally, a quartz vein-free part of the listwanite has a mesoscopic east-vergent fold.

Okanogan Valley

Before the recognition of ophiolite complexes, the Palmer Mountain Greenstone (unit Pk of Fig. 6) was thought to be a thick pile of mafic volcanic rocks unconformable on the Anarchist pelites (Rinehart and Fox, 1972). We believe that this contact is the Chesaw fault. Significantly, north of Spectacle Lake, the basal 60–80 m of Palmer Mountain greenstone is aphanitic and cut by numerous closely spaced faults. In outcrop, these faults range from less than 10 cm in length to greater than 2 m, are moderately nonplanar, and display slickensided surfaces. Faults with lengths greater than 20 cm in two 3 x 10-m outcrops display a noticeably preferred dip to the northwest (Fig. 10). Where cut by these faults the greenstone is locally foliated. Rinehart and Fox (1972, p. 12) evidently regarded these foliations as "wispy thin laminae" associated with pyroclastic rocks. Instead, these foliations, in fine-grained and strongly chloritized rocks, are probably features related to faulting. In some places, fractures cut obliquely across the foliation. These fractures typically contain fibrous calcite which gives an overall top-to-the-east sense of shear.

North and south of Hicks Canyon (loc. 2 on Fig. 6) the footwall of the Chesaw thrust is marked by slaty rocks. North of Hicks Canyon, the footwall rocks consist of a 300-m-thick zone of slaty felsic volcaniclastic rocks (gray) and pelitic rocks (black). The hanging wall is green phyllite derived from Palmer Mountain Greenstone (but was mapped as mafic intrusive rock and part of the Bullfrog Mountain Formation by Rinehart and Fox, 1972). Nonetheless, Rinehart and Fox (1972, p. 71) mapped part of this contact as a fault that dips 25–40 degrees to the north.

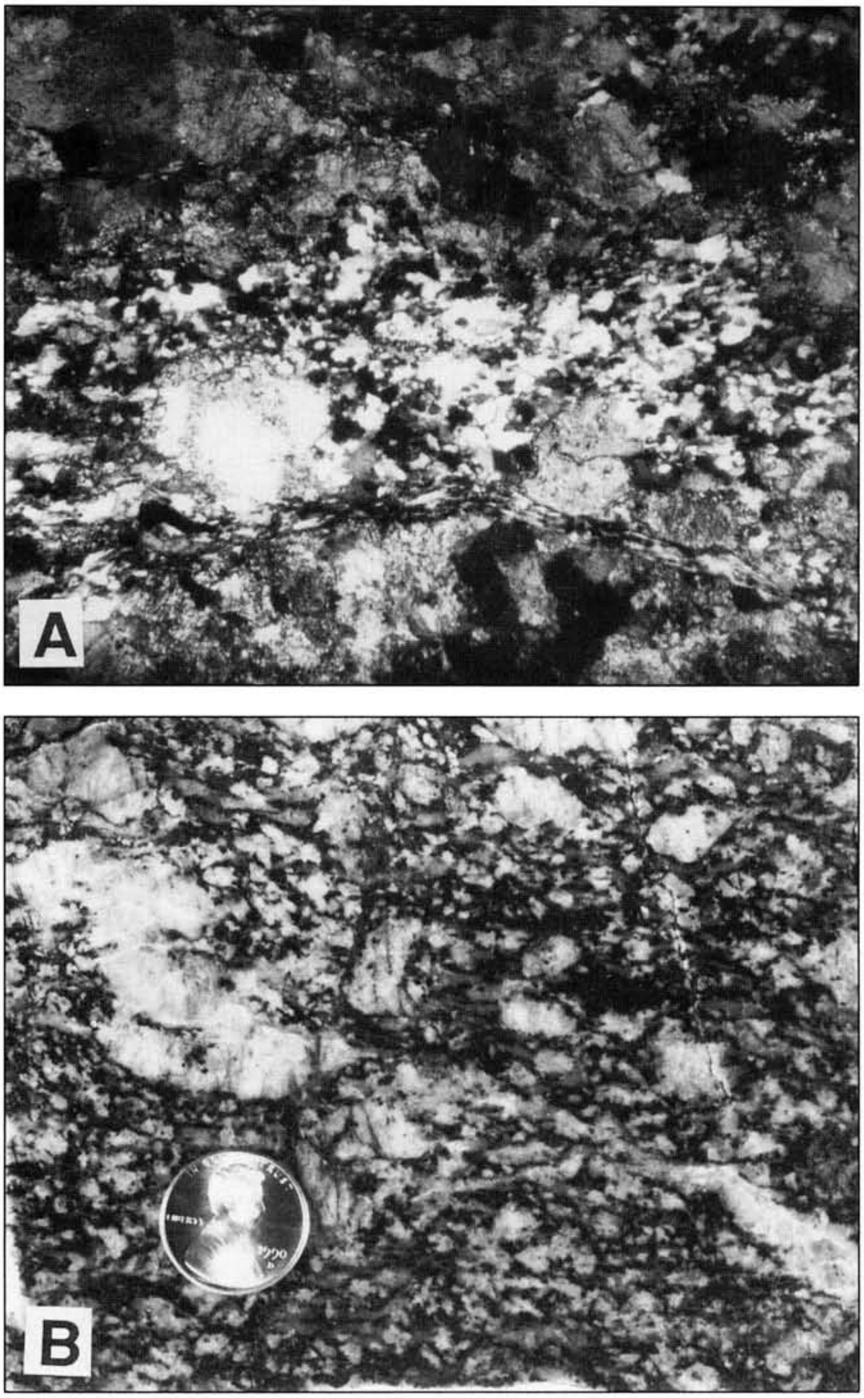

Figure 9. Rocks associated with the Chesaw and Dunn Mountain faults. **A.** Foliated listwanite from the Chesaw fault on the south side of Molson Hill, locality I of Fox and Rinehart (1968); field of view is about 2 mm. White and black grains are quartz; gray is magnesite. **B.** Strongly foliated megacrystic phase of the Evans Lake pluton from the Dunn Mountain fault on the southwestern contact of the pluton. Note the fractured K-feldspar megacrysts. Coin is 19 mm in diameter.

A segment of the Chesaw fault about 2 km long is exposed east of the northerly fault that borders locality 2 of Figure 6. The western end of this segment is along the north end of Wannacut Lake. The footwall is marked by a 200- to 600-m-thick zone of calcite-bearing, green slaty rock (slaty greenstone), black slaty rock (metapelite), and limestone veined by quartz. Within the pale-green slaty rocks are outcrops that have relict medium-grained igneous textures; these are the Permian or Triassic mafic intrusive rocks of Rinehart and Fox (1972). Rinehart and Fox mapped the pale-green slaty rocks as limestone and the conglomerate-bearing pelitic member of the Spectacle Formation (units Psl and Psc, respectively) of the Anarchist Group. The hanging wall is the conglomerate-bearing member of the Ellemeham Formation of Rinehart and Fox (1972).

Rinehart and Fox (1972, 1976) mapped the Cayuse Mountain thrust fault in the Anarchist Group east of the trace of the Chesaw Fault (Fig. 6). They reported that the Cayuse Mountain fault dips 11–25 degrees to the west. Perhaps this fault is satellitic to the Chesaw fault.

Listwanites west of Osoyoos Lake (Fox and Rinehart, 1968, pl. 1) and pyroxenite or gabbro east of the lake (Fox, 1970) occur on the contact of greenstones with felsic volcanic rocks or pelitic phyllites. These are at locality 1 and the um of Figure 6, respectively. We infer that these rocks mark windows in the Chesaw thrust. Perhaps the serpentinites northwest of Palmer Lake also are near or on the Chesaw fault zone.

Between the Evans Lake and Pogue Mountain plutons (E and P, respectively, of Fig. 6) the Cave Mountain succession is bounded below by serpentinite, the bottom of which is not exposed. The serpentinite contains podiform chromitite (Huntting, 1956). Because the Chesaw thrust places ophiolitic rocks over the Anarchist assemblage, we infer that this variously talcose serpentinite marks the Chesaw thrust. Adjacent to the Pogue Mountain pluton, the serpentinite is contact metamorphosed and was described as a calc-silicate granofels by Sims (1984).

The ultramafic rocks in the sharpstone conglomerate in the eastern part of the Cave Mountain succession may represent a thrust barely penetrated by erosion. Waters and Krauskopf (1941) and Rinehart and Fox (1976) regarded this contact as an unconformity. We regard it as the continuation of the fault zone exposed south of the Evans Lake pluton.

We also infer that east of the Mud Lake pluton (M of Fig. 6) the northern contact of the Cave Mountain succession is the Chesaw thrust. Here metabasalt of the Cave

Mountain succession (unit Jre of Fig. 6) overlies phyllitic pelites. Rinehart and Fox (1976) believed that this contact represents the depositional interfingering of their metamorphic complex of Conconully with the Cave Mountain succession. Our mapping shows that areas of metabasalt are synformal, whereas areas of phyllitic pelite are antiformal. Thus, no interfingering exists. Furthermore, the phyllites are western assemblage (unit Pap), not phyllitic portions of the gneisses of the southwestern metamorphic belt.

Designating the fault marked by serpentinites below the Cave Mountain succession as a thrust may be a problem: if the serpentinites are ignored, the fault places younger rocks (the Cave Mountain succession) over phyllitic rocks (unit Pap) that are thought to be older (part of the western assemblage). Although Rinehart and Fox (1976) believed the contact is an unconformity, Fox (written commun., 1993) suggested that if it is a fault, the younger-over-older relation implies a detachment fault. If so, it presumably is related to the nearby Okanogan metamorphic core complex. Another possibility, for which no other known evidence exists, is based on the observation that regionally the phyllitic pelites (unit Pap) are structurally above the Anarchist Group (Rinehart and Fox, 1976, cross section B-B′ of pl. 1). Thus, possibly, the phyllitic rocks are not part of the Anarchist Group but are significantly younger, perhaps the Hall Formation. If so, the fault is a thrust.

West of Tonasket, eastward dipping phyllitic greenstones (unit Pk of Fig. 6) overlain by possible Archibald Formation may represent another klippe of the Chesaw thrust. Magnesitic rocks do occur along the basal contact of these phyllitic greenstones (Rinehart and Fox, 1976).

Greenwood–Curlew area

Figure 4 shows thrusts in the Greenwood–Curlew area that we regard as equivalent to the Chesaw thrust. Fyles (1990) mapped the area north of the international boundary. The U.S. part of Figure 4 is our interpretation of Parker and Calkins (1964), Pearson (1977), Herdrick and Bunning (1984), and our own limited reconnaissance of the Lone Star and Morning Star prospects. For clarity, the map omits the numerous Tertiary hypabyssal intrusions. We have adopted Fyles' adage (1990) that ultramafic rocks commonly (but not always) mark the traces of major thrust faults. Although we place the detachment faults bounding the metamorphic core complexes at different locations than previous authors (Orr and Cheney, 1987; Stoffel, 1990a; Stoffel and others, 1991), that is not the main topic here.

The major difference between the geology of the Greenwood–Curlew area (Fig. 4) and other areas to the west, in which the Chesaw fault occurs (Figs. 6 and 7), is that Fyles (1990) mapped four northward dipping thrusts (No. 7, Mount Wright, Mount Attwood, and Lind Creek faults), instead of one. The absence of multiple faults elsewhere could be due to the lack of detailed mapping. An alternative is that four splays, or duplexes, are restricted to the Greenwood–Curlew area. We examine below a third possibility that the Chesaw fault has been repeated by northwesterly striking folds.

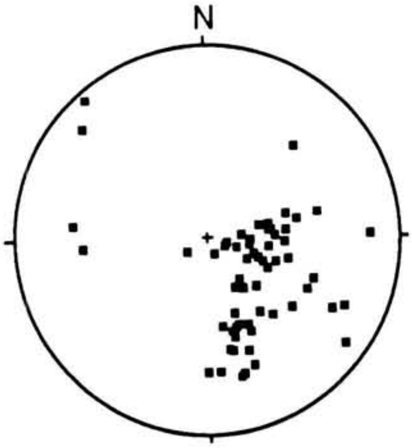

Figure 10. Poles to fault surfaces greater than 20 cm long in two 3 x 10-m outcrops of basal Palmer Mountain greenstone north of Spectacle Lake. N = 60.

None of the contacts mapped by Fyles (1990) are changed in Figure 4. As noted before, Brooklyn or Rossland greenstones are shown as a single unit, Jre, to enhance the map pattern. As previously discussed, we infer that some of the pelitic units are Archibald and Hall Formations.

We reinterpret the location of the Mount Attwood fault east of its northward salient near the center of the map. East of this salient, Fyles (1990) continued the fault along the northern contact of the unit shown as Jre. However, Fyles showed no ultramafic rocks along this contact; we suspect this contact is an unconformity. Therefore, southeast of the northern salient we show the unnamed fault mapped by Fyles north of the Mount Wright fault as the Mount Attwood fault. The southern side of the fault is bounded by the Knob Hill Group (unit Pk of Fig. 6) and ultramafic rocks; the northern side of the fault is bounded by units Jrh? and Jre. This interpretation is consistent with Fyles' criteria for faults and his observation (1990, p. 15) that the Mount Attwood fault is a splay of the Mount Wright fault.

Presumably, the Mount Wright/Mount Attwood thrust zone dips in the same direction as most of the bedding and foliation in its footwall and hanging-wall rocks. Thus, a southwestward dip of the fault zone between 118°33′W and 118°37′W is indicated by the majority of the strike and dips shown by Little (1983), Church (1986), and Fyles (1990). Furthermore, north of this portion of the fault zone, the map pattern is antiformal: the Brooklyn Formation is flanked on the north and south by greenstones (unit Jre). West of 118°37′W, the Mount Wright and Mount Attwood faults and the rocks on the south limb of the antiform all dip to the north (Fyles, 1990), indicating that the fold is overturned (Fig. 4).

The northwesterly trending belt of Attwood Group, Brooklyn Formation, and greenstones (unit Jre) south of the Lind Creek fault appears to be a northwesterly plunging antiformal window in the Chesaw thrust. In this interpretation, the No. 7, Mount Wright/Mount Attwood, and Lind

Creek faults are folded repetitions of a single fault or fault zone (cross section A–A′ of Fig. 4). By way of confirmation, the synformal dips of the Mount Wright and Mount Attwood faults with respect to the No. 7 fault in the central part of cross section A–A′ are similar to the relations shown by Little (1983, cross section A–B) about 2 km to the northwest.

Weaknesses in the antiformal interpretation are that the maps of Little (1983), Church (1986), and Fyles (1990) show few strikes and dips near the trace of the axis we infer in Figure 4 and few strata dip to the south near and east of cross section A–A′ where we infer the fold to be upright. One could argue that bedding beneath the folded unconformities (below the Brooklyn Formation and below unit Jre) and beneath the folded fault (Mount Attwood segment) need not dip southward. In any event, on cross section A–A′ and for 2 km southeast of it, the axial trace shown in Figure 4 coincides with the crudely v-shaped map pattern shown by Fyles (1990) of a limestone in the Attwood Formation and of a sharpstone conglomerate in the Brooklyn Formation. This trace also coincides with a zone of anomalously northward striking bedding in these rocks shown by Little (1983) and with two southwesterly dips in the limestone shown by Little; perhaps these rocks are on or near the crest of the northwesterly striking antiform.

A potential flaw in the antiformal interpretation is the southerly dipping pelitic rocks north of the Mount Attwood fault about 10 km west of Grand Forks (unit Jrh? in Fig. 4). Little (1983, p. 10) reported that the northern outcrops of these rocks contained poorly preserved brachiopods and bryozoa that are "probably Paleozoic, possibly early Triassic." Consequently, both Little (1983) and Fyles (1990) assigned these rocks to the Attwood Group. Because these pelitic rocks structurally overlie the Brooklyn/Rossland greenstones (unit Jre of Fig. 4), Fyles (1990) inferred that they are overturned. However, if the determination of the age of the poorly preserved fossils is incorrect and the rocks should prove to be no older than early Triassic, the antiformal hypothesis would be confirmed.

A northwesterly striking antiform that folds the Chesaw thrust in a manner similar to that suggested here was mapped by Rinehart and Fox (1972, p. 66) near the Whiskey Mountain pluton (P in Fig. 6). The southwest limb of this Whiskey Mountain anticline dips 10–75 degrees southwest, whereas on the northeastern limb, dips are as steep as 90 degrees (Rinehart and Fox, 1972).

Dunn Mountain Fault

Northwest of Omak (Fig. 6), the Dunn Mountain fault separates greenschist-facies Quesnellian rocks from the amphibolite-facies rocks of the southwestern metamorphic belt. Rinehart and Fox (1976) believed that the contact is depositional but realized that it might be a fault. We suspected the existence of a fault on the basis of the metamorphic discontinuity, the lack of significant contact metamorphism of the Cave Mountain succession by the Evans Lake and Dunn Mountain plutons (E and D, respectively, of Fig. 6), and an undescribed thrust fault along the eastern margin of the Dunn Mountain pluton shown by Rinehart and Fox (1976).

Southwestern metamorphic belt

Rinehart and Fox (1976) realized that for several kilometers east of Conconully, metapelites have a higher metamorphic grade than pelites of the western assemblage, which are typically in greenschist facies; however, they could not draw a precise boundary between the two. They noted that the metamorphic rocks near Conconully are biotitic, are intruded by a variety of alaskites and pegmatites, and locally contain garnet, andalusite, and sillimanite near plutons and some chloritoid and kyanite elsewhere. Our reconnaissance mapping and petrography show that most of the metapelites near Conconully are fine-grained biotite-muscovite-microcline-quartz gneisses, have millimeter- to centimeter-scale banding, and are cut by millimeter- and centimeter-wide quartz veinlets. We include these metasedimentary rocks in the southwestern metamorphic belt.

The reason that the contact is difficult to recognize is that the amphibolite-facies pelitic rocks and the pelitic rocks of the western assemblage undergo a textural convergence: each becomes phyllitic in a zone 4–9 km wide adjacent to the Dunn Mountain fault. In Figure 6, the phyllitic part of the southwestern metamorphic belt is shown as unit hmp, and the phyllitic portion of the western assemblage is labeled Pap. Foliations and cleavages in the phyllitic zones generally strike northerly to northwesterly and dip both westerly and easterly at 60 ± 20 degrees. These reversals in dip and the lack of marker units make determinations of the thicknesses of the phyllitic zones unreliable, but they probably exceed 1 km.

The contact between units hmp and Pap can be megascopically distinguished only with difficulty (and in places only within a few hundred meters) by searching for rocks with fine-grained biotite and quartz or disrupted quartz veinlets indicative of gneiss-derived phyllites (unit hmp). Rare outcrops of biotite gneiss or of rocks with relict 1–10-cm-scale bedding, indicative of units hmp and Pap, respectively, also occur in the phyllitic zones. Additionally, hmp tends to have northwesterly trending lineations caused by minute kink-like crenulations or the intersection of cleavage and foliation. These phyllites also contain minor microscopic garnet, staurolite, and sillimanite (fibrolite), whereas adjacent pelitic rocks in the Cave Mountain succession and the western assemblage do not.

We name the boundary between the two different phyllites the Dunn Mountain fault. The name is derived from Dunn Mountain where Rinehart and Fox (1976) illustrated that their metamorphic rocks of Conconully (here mostly hmp) and the Dunn Mountain pluton are thrust over the Cave Mountain succession.

Evans Lake pluton

The Dunn Mountain fault is exposed along the south side of the Evans Lake pluton (E in Fig. 6), where foliations generally dip 20–50 degrees to the northeast. Here the succession of rocks upward is serpentinite, unrecrystallized (non-contact metamorphosed) limestone 0–40 m thick, and a strongly foliated part of the Leader Mountain orthogneiss, the porphyritic phase of the Evans Lake pluton (unit Mom of Fig. 6). These relations indicate the presence of two faults. The upper fault (Dunn Mountain fault) separates Mom of the Evans Lake pluton from the limestones of the Cave Mountain succession, whereas the serpentinite is interpreted as the top of the northeast-dipping Chesaw fault zone below the Cave Mountain succession.

In addition to unit Mom, the Evans Lake pluton has a discontinuous border phase of fine- to medium-grained granodiorite to diorite (Rinehart and Fox, 1976; Sims, 1984), which is too small to show in Figure 6. In the interior of the pluton, Mom is only weakly foliated, but both phases are weakly cataclastic (Rinehart and Fox, 1976; Sims, 1984). As Sims (1984) noted, along the southwest margin of the pluton, Mom is strongly foliated. Biotite is fine grained; some of the feldspar megacrysts are boudinaged, and most of the megacrysts are fractured nearly perpendicular to the foliation (Fig. 9B). Sims (1984) described this foliation as mylonitic and believed that it was a protoclastic magmatic flow foliation. We regard it an indication of the Dunn Mountain fault.

Our preliminary work along the southwestern margin of the Evans Lake pluton gives conflicting directions of tectonic transport. The ultramafic rocks are cut by numerous outcrop-scale, brittle faults; those faults with visible separations of talcose zones indicate overall top-to-the-east transport. Northeast of the highway along Johnson Creek, carbonate rocks of the Cave Mountain succession display south-vergent, overturned folds; possibly these are older structures carried passively by the Chesaw thrust. We are not yet able to discern a consistent direction of transport in the well-foliated Mom of the Evans Lake pluton in the hanging wall the Dunn Mountain Fault.

Cave Mountain succession

We concur with Rinehart and Fox (1976) that the northwesterly striking, southwesterly dipping strip of phyllitic pelitic rocks between the Evans Lake pluton and the Cave Mountain succession (shown as hmp in Fig. 6) is part of their metamorphic rocks of Conconully (that is, the southwestern metamorphic belt). We place the Dunn Mountain fault at the base of these phyllitic rocks, thereby explaining the presence of contact metamorphic andalusite in them and the lack of significant contact metamorphism in the underlying Cave Mountain rocks. The presence of 2 m of limestone above 0.6 m of phyllitic pelitic rocks at the contact in the gully southwest of Evans Lake supports this hypothesis. Placing the Dunn Mountain fault at the base of the strip of phyllite (hmp) also explains why this contact truncates all the underlying metabasaltic rock and half of the next underlying unit of slate and metalimestone of the Cave Mountain succession. Rinehart and Fox (1976) believed that the contact is depositional and that its discordance is due to depositional thinning of the Cave Mountain units.

West of the strip of phyllite, the trace of the fault is marked by limestones that Rinehart and Fox (1976) questioned as being part of the Cave Mountain succession. These rocks contain leucocratic biotitic gneissic interlayers, alaskites, and granitic dikes. In Figure 6, they are included as part of hmp. Farther north, the fault is marked by the paucity of contact metamorphism in the Cave Mountain succession adjacent to the Dunn Mountain pluton and by the lack of contact metamorphism in Pap bordering the Mud Lake pluton. The mafic minerals in the Dunn Mountain and Mud Lake plutons are pervasively altered to microscopic hydrothermal biotite, but the plutons are not cataclastic. Our reconnaissance mapping shows that bedding and cleavage in Pap wrap concordantly around the northeastern part of the Mud Lake pluton and that the metabasalt of the Cave Mountain succession is well foliated east of the pluton.

Near the Mud Lake pluton, pelite below unit Jre of the Cave Mountain succession does have microscopic biotite porphyroblasts subparallel to the foliation of kink bands and has more randomly oriented porphyroblasts of chlorite, some of which replace the biotite porphyroblasts. Some (but not most) of unit Pap northeast of the Mud Lake pluton also has chlorite porphyroblasts. Given the general lack of contact metamorphism around these plutons, these porphyroblasts are perplexing; perhaps they are caused by nearby dikes or by strain heating (Pavlis, 1986) associated with the very thick Dunn Mountain fault zone.

Reconnaissance mapping suggests that various rocks are tectonically interleaved in the Dunn Mountain fault along the eastern contacts of the Dunn Mountain and Mud Lake plutons. In a crescent as much as 300 m wide between the Dunn Mountain pluton and the carbonate rocks of the Cave Mountain succession are outcrops of variously foliated metabasalt, biotitic gneiss, and biotitic phyllite. A body of feldspar porphyroblastic biotitic gneiss >10 m by 1 m occurs in the phyllites less than 100 m northeast of the Mud Lake pluton.

Summary

Misch (1949, p. 687) noted that the Evans Lake pluton is in the core of a syncline. This fold is shown by the synformal folding of the Dunn Mountain fault (Fig. 6). Thus, the Evans Lake pluton is rootless. The sinuous trace of the Dunn Mountain fault (like the Chesaw fault elsewhere) is due to later folding. Because the Dunn Mountain fault is bounded by wide zones of phyllitic rock and places amphibolite-facies gneisses and phyllites over greenschist-facies rocks, it may be a crustal-scale thrust.

Age of the Thrusts

The exact ages of the Dunn Mountain and Chesaw faults remain enigmatic. Near Loomis, the map pattern of Figure 6 implies that the Chesaw fault is truncated by the Loomis pluton. The one dated sample of this large pluton yielded a discordant K-Ar age on hornblende of 194 ± 6 Ma (Rinehart and Fox, 1972; Stoffel, 1990b). Local cataclastic textures and structures in the pluton (Hibbard, 1971) are of unknown significance.

If the Chesaw fault is older than 194 Ma, some of the Rossland Group might unconformably overlie the fault, but no such examples are known. Near Hicks Canyon (loc. 2 of Fig. 6), the Ellemeham Formation is cut by the Chesaw thrust. The 199.4 ± 1.4 Ma Lexington quartz porphyry (unit Jref of Fig. 4) is cut by the No. 7 fault. The quartz-sericite-pyrite-chalcopyrite schist at the Lone Star deposit to the southeast (Fig. 4) most likely is sheared Lexington quartz porphyry. At Lone Star this schist has a ^{40}Ar-^{39}Ar age of 104–103 Ma, with a later thermal disturbance between 60 and 50 Ma (Berger and others, 1991). The alkalic plutons of Shasket Creek (unit Ji of Fig. 4) dated at 163 ± 0.4 by ^{40}Ar-^{39}Ar (Berger and others, 1991) do not appear to cut the No. 7 fault but do seem to intrude serpentinite associated with it (Fig. 4).

Thus the bulk of the evidence suggests that the Chesaw fault may pre-date accretion of Quesnellia, which occurred before intrusion of the 164–169 Ma Nelson plutons (Lambert, 1989). Obviously, if the various faults which we believe are segments of the Chesaw fault prove to have significantly different ages, they could not be part of a single fault system.

The age of the Dunn Mountain fault also is unresolved. The predominantly directionless Blue Goat pluton (B of Fig. 6) seems to cut the Dunn Mountain fault. Rinehart and Fox (1976) reported that the rocks of the pluton commonly are granulated and recrystallized, but the distribution and significance of these textures remains unknown. K-Ar ages on a single sample of this pluton are 141.6 ± 8.2 Ma for hornblende and 98.9 ± 3.0 Ma for biotite (Rinehart and Fox, 1976).

The Dunn Mountain fault cuts the Leader Mountain orthogneiss (Mom) of the southwestern metamorphic belt, but the dating of Mom is unsatisfactory. Three localities have K-Ar ages on hornblende or biotite ranging from 99 to 63 Ma (Stoffel, 1990b, table 1); in particular, the megacrystic gneiss of the Evans Lake pluton has a K-Ar age on biotite of 88.8 ± 2.8 Ma (Rinehart and Fox, 1976). At Leader Mountain (K of Fig. 6) the orthogneiss has a two-point Rb-Sr age of 129 Ma (Menzer, 1970), a K-Ar age on hornblende of 98.5 ± 3.0 (Stoffel, 1990b, table 1), and intrudes a trondjemitic orthogneiss to the west (Rinehart, 1981) that has a K-Ar age on hornblende of 93.5 ± 2.8 Ma (Stoffel, 1990b, table 1). The megacrystic gneiss is in turn intruded by the Conconully pluton (Menzer, 1983), which has a five-point Rb-Sr isochron of 81.9 ± 0.8 Ma (Menzer, 1970; Stoffel, 1990b, table 1) and a 81.2 ± 2.4 Ma K-Ar age on hornblende (Stoffel, 1990b, table 1).

Clearly structural mapping and precise U-Pb ages on zircons are needed. To the west in the Cascade Range of Washington and British Columbia, contractional deformation of the Intermontane and Insular superterranes is mid-Cretaceous (McGroder, 1990). It will be interesting to see if the Dunn Mountain fault is of similar age.

Structures Younger than the Thrusts

Figure 6 shows that the Salmon Creek fault cuts the undated Pogue Mountain pluton (Menzer, 1983; Sims, 1984) and the megacrystic orthogneiss (Sims, 1984). The fault appears to be intruded by the Conconully pluton, the age of which probably is about 82 Ma. Diagonal-slip slickensides plunging to the south-southeast are exposed in subsidiary shears, but the sense of displacement and amount of offset are unknown (Rinehart and Fox, 1976). Because rocks of the southwestern metamorphic belt occur on both sides of the Salmon Creek fault, the fault is not a major one. The major fault (the one that marks the southwestern boundary of the Quesnellian rocks) is the Dunn Mountain fault.

Figures 4, 5, 6, and 7 show that the low-angle detachment faults bordering the metamorphic core complexes cut, and therefore floor, the Quesnellian and Challis rocks. This is confirmed by underground drilling in the Overlook mine (Fig. 5), which shows that the western assemblage and overlying Tertiary rocks are cut off at depth by a chloritic low-angle fault. On the surface to the west (Fig. 5) Muessig (1967) mapped this as the Lambert Creek thrust fault. We interpret this fault as the southern segment of the westward dipping Granby River–St. Peter detachment fault that bounds the western side of the Kettle metamorphic core complex. On the southwestern margin of the Okanogan metamorphic core complex, the Omak Lake detachment fault cuts the southwestern metamorphic belt (unit swm of Fig. 2).

Figures 2 and 6 show that the Mesozoic and Challis rocks are preferentially preserved adjacent to the detachment faults. This pattern may be caused by roll-over anticlines on the faults as described by Harms and Price (1992) on the Newport detachment fault along the Washington–Idaho boundary.

Northwesterly trending folds deform the Tonasket gneiss (unit hmt of Fig. 2) in the post-Challis Okanogan metamorphic core complex (Orr and Cheney, 1987, fig. 2; Fox and Rinehart, 1988). Such folds must, therefore, contribute to the sinuous trace of the Okanogan detachment fault shown in Figure 6 and the more sinuous trace of the Chesaw fault.

LITHOTECTONIC COMPARISON WITH NEVADA

In this section we compare lithologies in Washington with those in Nevada (Fig. 1). Cowan (1992) pointed out that some Cordilleran terranes of great lateral persistence appear to be missing in Washington. One of these is the Upper Paleozoic Golconda allochthon of northwestern Nevada. However, Snook and others (1981) realized that, in a style reminiscent of Nevada, the Ordovician eugeoclinal Covada Group along the Columbia River probably is in thrust contact with broadly coeval North American miogeoclinal rocks on the east. This boundary is shown as the Huckleberry Ridge fault and Columbia back thrust in Figure 2.

The Covada Group of the eastern assemblage appears to be a distal portion of North America thrust back onto the continent (Gehrels and Smith, 1987; Smith and Gehrels, 1991). A counterpart of the eastern assemblage may be the Roberts Mountain allochthon of Nevada (Gehrels and Smith, 1987; Rubin and others, 1990; Smith and Gehrels, 1991; Burchfiel and others, 1992) and the rocks of the Antler foredeep. The Roberts Mountain allochthon consists predominantly of oceanic Ordovician to Lower Mississippian strata, including Devonian bedded barite (Rubin and others, 1990). The allochthon was emplaced eastward against the shelf sequences of North America during the Mississippian Antler orogeny (Elison and others, 1990; Rubin and others, 1990). This allochthon was subsequently overthrust from the west in earliest Triassic time by the Golconda allochthon (Elison and others, 1990).

The Golconda allochthon (Fig. 1) is an assemblage of Paleozoic sedimentary and volcanic rocks. Predominantly sedimentary successions range in age from Devonian to Permian and include turbidites of continental provenance, hemipelagic and pelagic deposits, carbonate beds, and chert (Russell, 1984; Rubin and others, 1990; Murchey, 1990; Jones, 1991). Subterranes dominated by basaltic to andesitic greenstones and chert are Mississippian and Mississippian to Permian (Whiteford, 1990; Jones, 1991).

West of the Golconda allochthon, the Star Peak/Luning basin (Fig. 1) consists of Mesozoic rocks unconformable upon Paleozoic rocks that are coeval but unlike those in the Golconda allochthon (Burchfiel and others, 1992). In northwestern Nevada, younger volcanogenic clastic rocks contain cherts, Carnian to Norian carbonate rocks, and clastic carbonate rocks (the Boulder Creek subunit of Russell (1984), Star Peak Group (Silberling and others, 1987), and unnamed units described by Wyld (1990)). Interestingly, some of these units were originally thought to be terranes (Silberling and others, 1987) but now are recognized as being bounded by unconformities (Wyld, 1990). The Triassic rocks are locally overlain by Upper Triassic to Jurassic andesitic flows and coarse volcanic rocks (the Happy Creek complex of Russell, 1984). At least locally, Golconda-age units are thrust over the Mesozoic rocks (Russell, 1984). All these rocks were multiply deformed between the Middle Jurassic and Early Cretaceous (Russell, 1984; Elison and others, 1990).

The southwestern metamorphic belt also could have counterparts in Nevada and Idaho. Amphibolite-facies pelitic rocks occur west of the Mesozoic rocks in the Star Peak/Luning basin near the Oregon border. However, these metamorphic rocks are believed to be Paleozoic and are not thrust over the Golconda and other rocks (Wyld, 1990). It is tempting to speculate that the amphibolite-facies Owyhee metamorphic rocks in southwestern Idaho (Fig. 1) may be part of the same metamorphic belt.

Counterparts of the Golconda allochthon, the Triassic strata, and the Jurassic volcanic rocks of the Star Peak/Luning basin could be the western assemblage of Quesnellia, Brooklyn Group, and Rossland Group, respectively. The timing of tectonic events in Nevada and Quesnellia also seems to be similar, but the ages of rocks and timing of events need not be exactly the same for these to be parts of the same or related terranes.

All the pre-Cenozoic rocks in Nevada plunge northward below Tertiary rocks (Fig. 1); Quesnellia and the eastern assemblage do the same southward in Washington. However, these belts cannot be continuous in the subsurface below the Tertiary rocks because the lithologically different Blue Mountain terranes intervene (Fig. 1).

CONCLUSIONS

Although the geology and timing of events in southern Quesnellia are still poorly known, we have recognized the major unconformity-bounded sequences (western assemblage, Brooklyn Formation, and Rossland Group). We also have identified two of the major structures (the Chesaw and Dunn Mountain faults). Quesnellia and its bounding units greatly resemble the rocks of northwestern Nevada, for which paleogeographic and plate tectonic models (Jones, 1991) are more advanced than for Quesnellia. The relation to, or interaction of, southern Quesnellia with terranes to the east (eastern assemblage) and to the west of the southwestern metamorphic belt remain poorly known.

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A log protrudes from a rockslide-avalanche deposit along the north bank of Glacier Creek in Whatcom County. The rockslide-avalanche originated at Church Mountain (partially cloud covered in the background) about 6.5 km to the northeast across the North Fork Nooksack River valley. Radiocarbon dates suggest that it took place about 2,500 radiocarbon years ago and may have been triggered by an earthquake. Cary, Easterbrook, and Carpenter first described the lithologic characteristics of the slide in their GSA abstract[1]. The landslide chutes visible on the south face of Church Mountain are part of a larger area lacking well-developed talus deposits that is inferred to be the source area for the rockslide-avalanche. In a detailed study of the deposits, Carpenter[2] noted that the deposit extends for about 9 km and has an estimated volume of 2.8×10^3 m^3. Photo by Patrick Pringle, 1992.

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Tectonostratigraphic Framework of the Northeastern Cascades

Robert B. Miller[1], Ralph A. Haugerud[2], Fred Murphy[1], and Lynda S. Nicholson[1]

[1]Department of Geology
San Jose State University
San Jose, CA 95192-0102

[2]U.S. Geological Survey
c/o Department of Geological Sciences
University of Washington
Seattle, WA 98195

ABSTRACT

The tectonostratigraphic framework of the northeastern Cascades is complex and in part controversial, largely because of the obscuring effects of Cretaceous and Paleogene (~96–45 Ma) plutonism and dynamothermal metamorphism in the crystalline core of the North Cascades (Cascades core). We describe a broad transect through the northeastern Cascades that begins in the Methow basin. The weakly metamorphosed strata of the Methow basin at least in part overlie basement of Lower Triassic ocean ridge basalt and are divisible into Albian and older clastic and volcanic rocks, Albian and Cenomanian(?) arkosic marine turbidites, and Upper Cretaceous shallow marine and continental clastic and andesitic volcanic rocks of the Pasayten Group.

On the southwest, the Methow basin is separated from the Cascades core by the broad, dominantly dextral, Cretaceous to Eocene Ross Lake Fault Zone. A horse more than 60 km long within the fault zone contains lithologies similar to those of the Upper Cretaceous Pasayten Group, but it records metamorphic conditions intermediate between those to the northeast and southwest. Along strike to the southeast are amphibolite-facies metaclastic rocks and metabasites with island-arc tholeiite affinities that may comprise an exotic slice within the fault zone. Farther southwest across strike within the Ross Lake zone is an oceanic assemblage of metamorphosed chert, basalt, limestone, clastic rock, and ultramafite. These rocks are part of the Napeequa unit of the Chelan Mountains terrane of the Cascades core and probably correlate with the low-grade Mississippian to Jurassic Bridge River–Hozameen terrane. These correlations suggest that a terrane boundary lies within the Ross Lake zone.

Southwest of the fault zone, hornblende-biotite schist, amphibolite, calc-silicate rock, and biotite paragneiss occur as rafts within Cretaceous and Paleogene orthogneiss of the Skagit Gneiss Complex. These rafts are similar to more intact belts of metasupracrustal rocks of the Holden assemblage southwest of Lake Chelan. The Holden assemblage consists dominantly of hornblende gneiss, amphibolite, and hornblende-biotite schist and gneiss, plus widespread biotite schist, biotite gneiss, leucogneiss, and calc-silicate rock, and rare pelitic schist and metaconglomerate. These metamorphosed intermediate and mafic flows and tuffs, clastic rocks, and calcareous sediments represent a volcanic arc assemblage that has been intruded by Late Triassic plutons. The assemblage probably correlates with the Cascade River unit, which is at least in part Late Triassic and has previously been assigned to the Chelan Mountains terrane. These correlations and other observations lead us to infer that the Napeequa unit and Cascade River unit may represent separate terranes.

INTRODUCTION

The North Cascades of Washington and British Columbia are commonly treated in Cordilleran-scale tectonic syntheses (for example, Coney and others, 1980; Monger and others, 1982) as a composite terrane consisting of numerous "miniterranes". This lack of detailed terrane analysis in part results from the small sizes of the terranes and the complex structural and thermal events that postdate terrane amalgamation. These effects are particularly evident in the crystalline core of the North Cascades (Cascades core), where widespread Cretaceous and Paleogene (~96–45 Ma) plutonism and coeval, generally medium- to high-grade metamorphism have obscured original relations of protoliths (for example, Tabor and others, 1987a, 1989).

Tabor and others (1987a, 1989) subdivided the Cascades core into three major terranes (Fig. 1): the pre-Late Cretaceous Swakane terrane, which consists mostly of biotite gneiss that has a Precambrian isotopic signature; the dominantly psammitic and pelitic, pre-Late Cretaceous Nason terrane; and the Chelan Mountains terrane. The Chelan Mountains terrane consists of two major supracrustal units, the volcanic arc-derived metaclastic and metavolcanic Cascade River unit and the ocean floor-related, metabasalt- and metachert-rich Napeequa unit (Tabor and others, 1988, 1989; Brown and others, 1993).

Northeast of the Cascades core lie Mesozoic, weakly metamorphosed, dominantly clastic marine and nonmarine strata of the Methow basin. Separating the core from the Methow basin is the Ross Lake Fault Zone. As defined by Misch (1966), the Ross Lake zone is actually a system of faults and fault zones (Fig. 2) of Late Cretaceous to Eocene age that records dominantly dextral strike slip, with components of Paleocene reverse slip and Eocene normal slip (Misch, 1966; Haugerud, 1985; Miller and Bowring, 1990; Miller, 1994). The fault zone forms both a metamorphic and lithologic discontinuity (Miller, 1994), although a

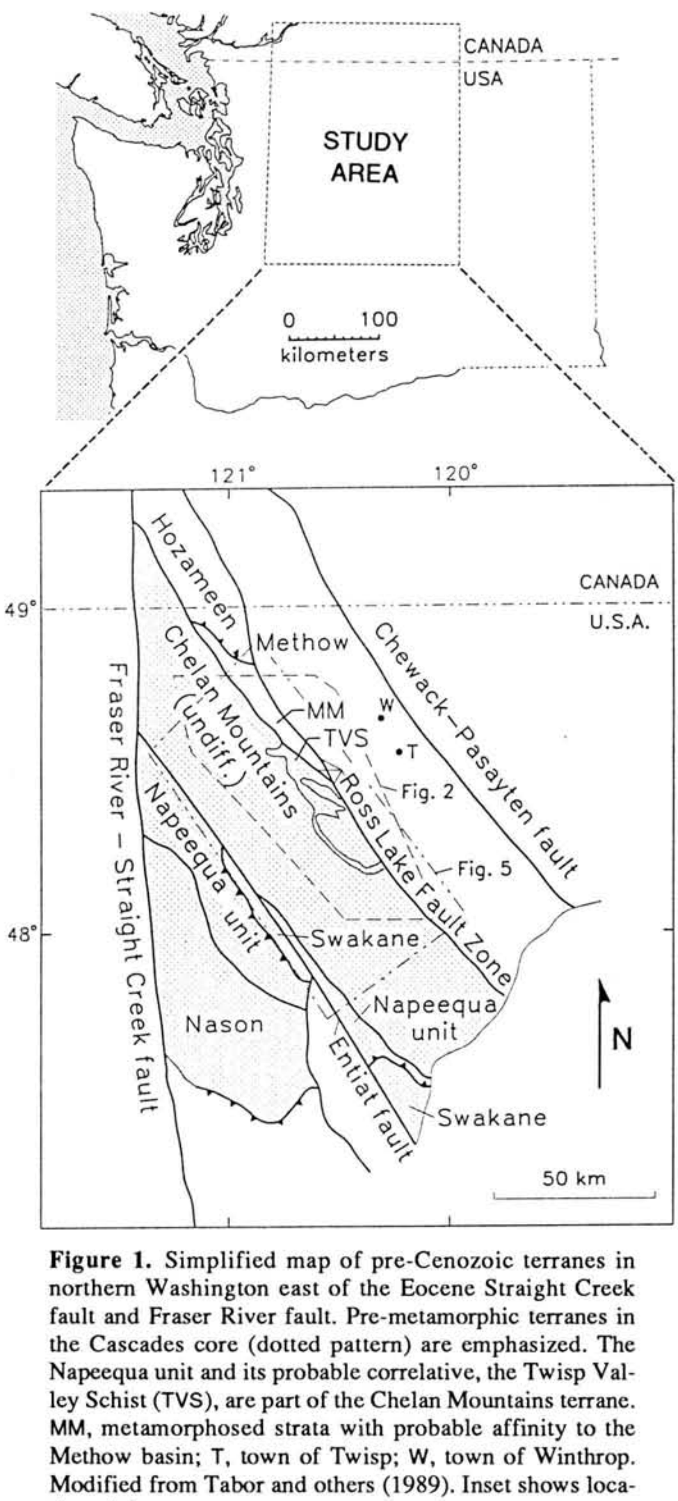

Figure 1. Simplified map of pre-Cenozoic terranes in northern Washington east of the Eocene Straight Creek fault and Fraser River fault. Pre-metamorphic terranes in the Cascades core (dotted pattern) are emphasized. The Napeequa unit and its probable correlative, the Twisp Valley Schist (TVS), are part of the Chelan Mountains terrane. MM, metamorphosed strata with probable affinity to the Methow basin; T, town of Twisp; W, town of Winthrop. Modified from Tabor and others (1989). Inset shows location of the map.

horse more than 60 km long that contains Methow-like lithologies (Fig. 1) records metamorphic conditions that are generally intermediate between those to the southwest and northeast. This horse is bounded on the northeast by the Jack Mountain thrust (not labeled on Fig. 2, but shown on Fig. 5), Hozameen fault, and North Creek fault, and on the southwest by the Twisp River fault and Ross Lake fault, which in the study area has been obliterated by intrusions (Fig. 2) (Misch, 1966; Tabor and others, 1989).

Two markedly different interpretations have been advanced for the tectonic significance of the Ross Lake Fault Zone. In one hypothesis, the Ross Lake zone represents a terrane boundary, separating oceanic rocks of the Napeequa unit of the Chelan Mountains terrane from clastic and volcanic strata of the Methow terrane (for example, Tabor and others, 1989; Miller and Bowring, 1990). In a second model, supracrustal rocks on the west side of the Ross Lake zone are mainly metamorphosed strata of the Methow basin and any terrane boundary lies farther west in the Cascades core (Kriens and Wernicke, 1990). The Ross Lake fault is interpreted by Kriens and Wernicke (1990) to mark an intrusive contact between mid-Cretaceous orthogneiss and Methow rocks, which has been slightly modified by brittle faults and mylonite zones of negligible displacement, and the Ross Lake zone is thus in this view a regionally insignificant structure.

One of the major goals of this paper is to evaluate the contrasting hypotheses for the distribution of units across the Ross Lake Fault Zone. A second objective is to describe the original relations within the Chelan Mountains terrane between the Napeequa unit and Cascade River unit and to examine the hypothesis that these units represent separate terranes. We treat these problems within the framework of a broad northeast-southwest transect extending for more than 40 km through the northeastern North Cascades. The level of mapping along this transect ranges from detailed to reconnaissance. We emphasize supracrustal tectonostratigraphic units (Table 1) and their protoliths that form the framework to the northeastern part of the Cascades core, treating only briefly metamorphism, plutonism, and structures.

METHOW BASIN

Strata of the Methow basin south of 49°N are divisible into three packages (Fig. 3). (1) Lower Albian (uppermost Lower Cretaceous) and older strata are dominantly volcanic-lithic; they are overlain by (2) Albian arkosic marine turbidites of the Harts Pass Formation and strata of the overlying Three Fools unit (informally named in this paper). These older packages lack clear relations with surrounding tectonostratigraphic elements, and we call them the "Methow terrane". Strata of the Methow terrane are unconformably overlain by (3) shallow-marine and continental sedimentary and andesitic volcanic rocks of the Upper Cretaceous Pasayten Group. West-derived chert-pebble conglomerates of the Pasayten Group link the Methow basin to a chert-bearing source terrane to the west in early Late Cretaceous time, as first noted by Tennyson and Cole (1978). The base of the Methow succession is not exposed in Washington. Farther north, Methow strata overlie Early Triassic or older mid-ocean ridge tholeiites of the Spider Peak Formation (Ray, 1986).

Most rocks of the Methow basin are little metamorphosed. Argillites are commonly cleaved and some detrital plagioclase is albitized (Coates, 1974), but depositional features are otherwise well preserved. Greenschist-facies alteration is common in volcanic strata low in the section, and contact-metamorphic andalusite- and cordierite-bearing assemblages are developed around several plutons.

The following stratigraphic summary of the Methow basin draws heavily from unpublished mapping by the U.S. Geological Survey (USGS).

Lower Albian and Older Strata

This package includes rocks that have been mapped as belonging to the Panther Creek, Buck Mountain, Twisp, and Newby units of Barksdale (1975) (Fig. 3). Youngest are massive argillite, sandstone, and thick, lensoidal, granitoid cobble-rich conglomerate of the Panther Creek Formation and similar conglomerates associated with volcanic-lithic sandstone and argillite in the Buck Mountain Formation. The Panther Creek Formation and Buck Mountain Formation in their type areas are both Early Cretaceous in age, on the basis of contained marine fossils (Barksdale, 1975; McGroder and others, 1990; W. P. Elder, USGS, written commun., 1993). The Panther Creek Formation is about 600 m thick in its type area and is truncated at its base by a fault. The base of the Buck Mountain Formation in its type area is a fault, and, barring significant internal faulting, the unit is more than 2.9 km thick (Barksdale, 1975). Fossils collected from Buck Mountain-like lithic sandstone that underlies the Harts Pass Formation in the Pasayten River drainage near 49°N are Albian (Staatz and others, 1971, p. 24; W. P. Elder, USGS, written commun., 1993, on the basis of trigoniid age constraints published by Poulton, 1977).

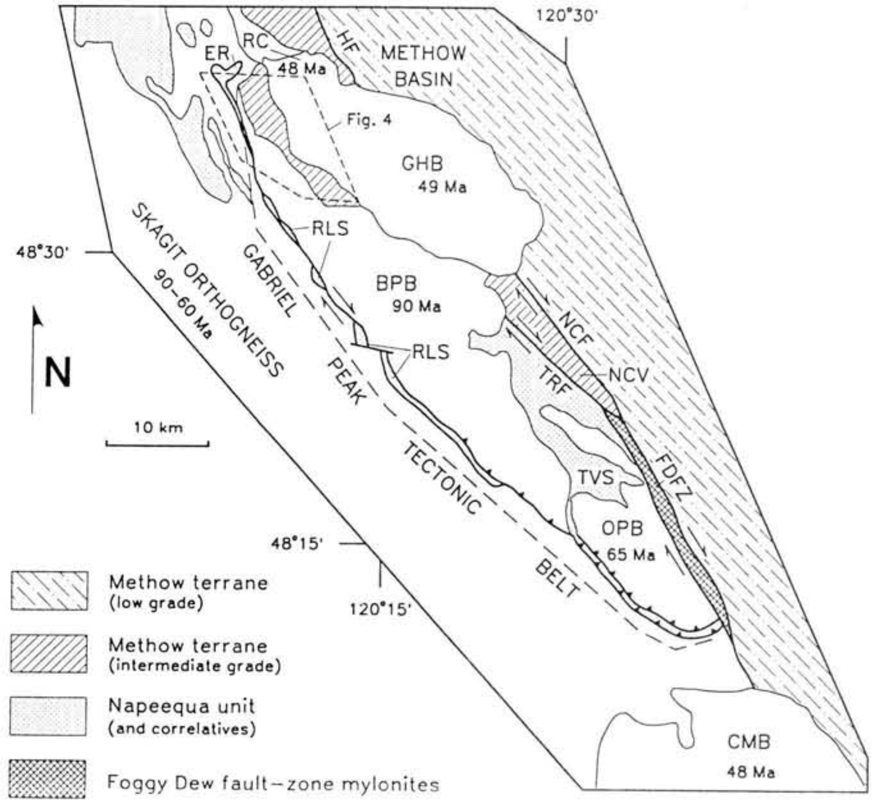

Figure 2. Map emphasizing units within and adjacent to the Ross Lake Fault Zone. The Ross Lake zone is actually a system of faults and fault zones that include the Ross Lake fault proper (north of study area), Hozameen fault (HF), North Creek fault (NCF), Foggy Dew Fault Zone (FDFZ), Twisp River fault (TRF), and Gabriel Peak tectonic belt. The long dashes show the western limit of the Gabriel Peak belt, which is a high strain zone that approaches 5 km in width. BPB, Black Peak batholith; CMB, Cooper Mountain batholith; ER, Elijah Ridge; GHB, Golden Horn batholith; NCV, North Creek Volcanics; OPB, Oval Peak batholith; RC, Ruby Creek heterogeneous plutonic belt of Misch (1966); RLS, Rainbow Lake Schist; TVS, Twisp Valley Schist. Modified from Miller and Bowring (1990) and Miller (1994).

The Twisp Formation, comprising thin-bedded argillite, lithic sandstone, minor limestone, and local tuff and tuff-breccia, has not yielded diagnostic fossils but is probably older than the Buck Mountain and Panther Creek units. Many workers have proposed that the Twisp Formation is correlative with the Lower and Middle Jurassic Ladner Group (Coates, 1974; O'Brien, 1986) exposed in British Columbia. No estimate of thickness is available for the strongly folded Twisp Formation.

Andesitic breccia, tuff, and flows that crop out west of Twisp were assigned by Barksdale (1975) to his undifferentiated Newby Group. These rocks are intruded by the earliest Cretaceous or older Alder Creek stock (Bunning, 1990). Altered andesitic breccias, flows, and tuffs that crop out north of Winthrop are intruded by the latest Jurassic (Stoffel and McGroder, 1990) Button Creek stock and may be correlative with the andesitic rocks west of Twisp. The thickness of these poorly stratified rocks is unknown.

Sparse paleocurrent data from these Albian and older strata suggest an eastern source terrain (Tennyson and Cole, 1978).

Harts Pass Formation and Three Fools Unit

Thick-bedded arkose, less abundant thin-bedded siltstone, shale, and fine sandstone, and local granitoid-cobble conglomerate of the Harts Pass Formation overlie massive argillite of the Panther Creek Formation. The Harts Pass is characterized by one- to several-meter-thick planar sheets of medium- to coarse-grained lithic arkose, commonly with load casts at the base, interior portions that are featureless except for local rip-up clasts and (or) pebbly zones, and ripple-cross-laminated and (or) convolute-bedded, fine-grained tops. Intervening layers of shale, siltstone, and

Table 1. Summary of supracrustal units of the northeastern Cascades

Unit name	Holden assemblage	Skagit Gneiss Complex	Twisp Valley Schist	Rainbow Lake Schist
Rock types	hornblende and biotite-hornblende schist and gneiss, biotite schist, leucocratic biotite gneiss, and clinopyroxene-biotite schist; minor siliceous marble and marble; rare metapelite, quartzite, and conglomerate	biotite schist, calc-silicate schist, amphibolite, and hornblende-biotite schist; rare quartzitic schist and pure marble; injected by a wide variety of variously metamorphosed, commonly leucocratic, intrusive rocks	siliceous biotite schist, impure quartzite, amphibolite, greenschist, calc-silicate rock, marble, metaperidotite, and pelitic schist	quartz-rich biotite schist, amphibolite, and hornblende-biotite schist; rare metapelite, metaperidotite, calc-silicate rock, and marble
Volcanic geochemistry	not available	not available	MORB and oceanic island basalt	basaltic
Protoliths	basalt and andesite, dacite(?), calcareous mudstone, argillaceous limestone, graywacke, limestone, quartz-rich sandstone	sandstone (graywacke?), argillaceous sediments, calcareous mudstone and (or) argillaceous limestone, and basalt and (or) andesite; rare limestone	chert, siliceous mudstone, basalt, and limestone; minor serpentinite and sandstone	chert, siliceous mudstone, basalt, volcanic graywacke, and (or) andesite; rare ultramafite (serpentinite?) and limestone
Inferred tectonic setting	volcanic arc and adjacent basin	island arc and adjacent basin(?)	ocean basin and oceanic island	deep marine basin, possibly fairly close to an active volcanic arc
Provenance of clastic rocks	volcanic arc	volcanic arc	?	volcanic arc?
Age controls	intruded by the Triassic Dumbell plutons, ~225 Ma	intruded by orthogneiss with inferred igneous crystallization ages as old as 87 Ma	intruded by ~90 Ma Black Peak batholith	intruded by protolith of the Skagit orthogneiss; pre-87 Ma(?)
Metamorphic grade	epidote amphibolite to middle amphibolite facies	middle and upper amphibolite facies	greenschist to middle amphibolite facies	middle to upper amphibolite facies

fine-grained sandstone are typically well bedded, with bed thicknesses of a few millimeters to a few centimeters. Sand layers are commonly amalgamated, producing featureless expanses of sandstone. Layers of pebble to cobble conglomerate occur locally near the base of the Harts Pass Formation. These conglomerates are similar to those in the underlying Panther Creek unit and do not appear to constitute a useful stratigraphic marker. Paleocurrents indicate an eastern source for the Harts Pass Formation (Tennyson and Cole, 1978). Marine fossils from the Harts Pass Formation were interpreted by McGroder and others (1990) to indicate an early to middle Albian age. Near the Canadian border, where a complete section of the Harts Pass Formation is exposed, the unit is about 3 km thick.

Finer grained, thinner bedded sandstone and associated siltstone conformably overlie Harts Pass Formation in the headwaters of Three Fools Creek in the western part of the basin. Tennyson (1974) and McGroder and others (1990) divided these strata amongst the Harts Pass and Virginian Ridge Formations, but they are not as sand rich, coarse grained, and arkosic as the Harts Pass and lack the distinctive chert clast-rich beds of the Virginian Ridge. This Three Fools unit overlies Albian Harts Pass strata and underlies the Virginian Ridge Formation that may be Turonian (see below), and thus is of later Albian or Cenomanian age. It is about 1 km thick.

Pasayten Group

The Virginian Ridge Formation, Winthrop Sandstone, and Midnight Peak Formation of Barksdale (1975) are here correlated with the Pasayten Group as defined in Manning Park, British Columbia (Coates, 1974). Lithologies are diverse, and the package contains several unconformities, some of which appear to be of only local extent. The Pasayten Group is separated from older rocks by a significant angular unconformity: at the eastern and southern

Napeequa Unit in the Gabriel Peak–Elijah Ridge area	Metamorphosed Methow(?) strata of the Gabriel Peak–Elijah Ridge area	Foggy Dew Fault Zone	North Creek Volcanics	Methow Basin
biotite ± hornblende ± garnet schist, amphibolite, quartzose schist, marble, ultramafite, and metagabbro	metamorphosed quartzose-chert conglomerate, quartzose sandstone, pelite heterolithic conglomerate, hornblende porphyry, and gabbro/diorite	amphibolite, greenschist, biotite schist, and mylonitic orthogneiss	hornblende-phyric andesite flows, sills, breccias and tuffs, conglomerate, arkose, siltstone, shale	arkose, argillite, lithic sandstone, chert-pebble conglomerate, heterolithic conglomerate, and andesitic tuff, breccia, and flows
not available	not available	island-arc tholeiitic basalt	not available	not available
tuff/volcanic sandstone/ volcanic argillite, basalt, chert, limestone, serpentinite, gabbro	chert-pebble conglomerate, sandstone, argillite, heterolithic conglomerate, hornblende-bearing shallow intrusions, and volcanic breccias and flows(?)	basalt, gabbro, andesite, sandstone, and siltstone	same as rock types	same as rock types
deep marine basin, possibly fairly close to an active volcanic arc	?	island arc	volcanic arc and adjacent basin	fore-arc basin, successor basin(?)
volcanic arc?	oceanic terrane in part (quartzose-clast conglomerate)	?	arc	volcanic arc (J?, lower K), eastern plutonic terrane (Harts Pass Fm., Winthrop Ss.), western oceanic terrane (Virginian Ridge Fm.), intra-basinal (Midnight Peak Fm.)
intruded by 90 Ma plutons. (Miss.?–) Perm.–Triassic (J?) by correlation with Hozameen Group, Cache Creek Complex, and Elbow Lake unit	intruded by ~90 Ma Black Peak batholith; Upper Cretaceous (in part)	amphibolites intruded by 65 Ma Oval Peak batholith	intruded by ~90 Ma Black Peak batholith	Cretaceous and probable Jurassic
high-P (≤9 kb) amphibolite facies	high-P (kyanite) and low-P (andalusite) amphibolite facies	amphibolite facies	greenschist facies (locally upgraded to hornblende hornfels facies next to Black Peak batholith)	unmetamorphosed to andalusite-cordierite hornfels facies

margins of the Methow block the Harts Pass Formation and overlying Three Fools unit were completely eroded prior to deposition of the Pasayten Group.

The Virginian Ridge Formation comprises argillite, thin- to thick-bedded sandstone, and conglomerate. Sandstone and conglomerate are commonly rich in chert clasts. Argillites locally bear limy layers and concretions. Paleocurrent measurements indicate a western source, presumably the Permian (or older) to Jurassic Hozameen Group (Tennyson and Cole, 1978; Trexler, 1985). Trexler (1985) described and named the Slate Peak Member, composed mostly of argillite with minor sandstone and chert-pebble conglomerate, which he inferred to be shallow-marine and deltaic deposits. He also named the overlying Devils Pass Member, with abundant chert-pebble conglomerate, which he inferred to be alluvial fan and delta deposits. In most places the Virginian Ridge Formation overlies the Harts Pass Formation or the Three Fools unit. Thicknesses range from a reported 4.2 km at the southwestern margin of the Methow block (McGroder, 1989) to nil at the eastern margin.

Winthrop Sandstone is dominantly thick- to thin-bedded fluvial sandstone that is commonly cross-bedded and has interlayered siltstone and argillite. Most of this sandstone is arkosic, but volcanic-lithic sandstone is extensive in some areas. The Winthrop lithofacies dominates the Pasayten Group in Manning Park. The Winthrop Formation was in different places deposited on Virginian Ridge Formation, Harts Pass Formation, and sub-Harts Pass strata, and in places it appears to grade laterally into the Virginian Ridge Formation. Estimated thicknesses of the Winthrop range from 50 m on the southwest margin of the Methow block to 4.1 km on the northeast (McGroder, 1989; Barksdale, 1975).

Red beds and andesitic flows, tuff, and breccia of the Midnight Peak Formation overlie, both conformably and unconformably, the Winthrop Sandstone and Virginian Ridge Formation. The unit appears to be largely terrestrial.

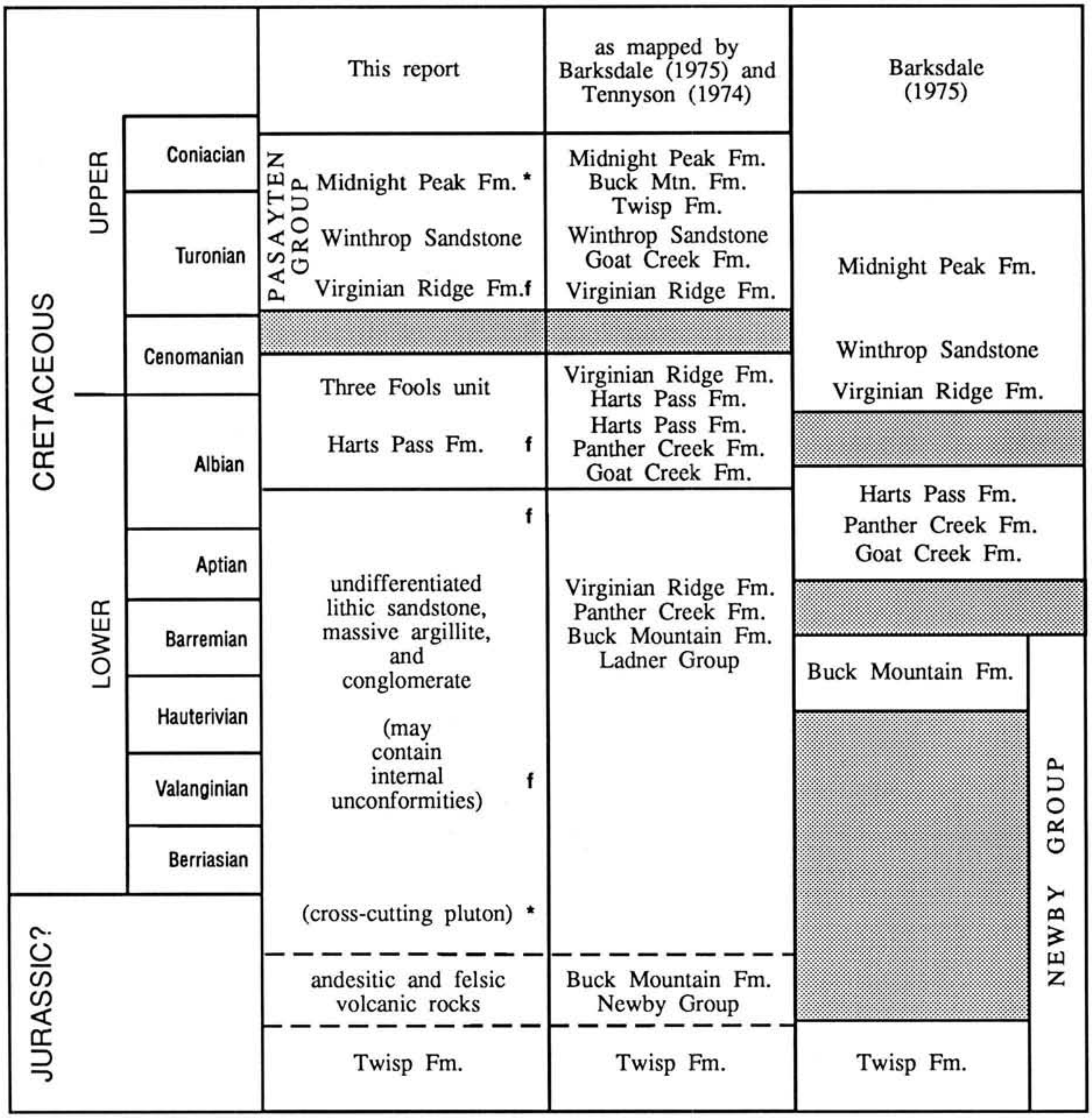

Figure 3. Stratigraphy of the Methow basin as determined by U.S. Geological Survey (USGS) mapping through April 1993, compared to that of Barksdale (1975). Formational identifications of the same rocks as those mapped by earlier workers are shown in the middle column. New age constraints are denoted by * (radiometric ages, L. W. Snee, USGS, written commun., 1992; Stoffel and McGroder, 1990) and **f** (fossil ages, W. P. Elder, USGS, written commun., 1993). Shaded areas represent section missing because of non-deposition or erosion.

Aggregate thickness of the Midnight Peak Formation is as great as 1.5 km (McGroder, 1989).

The Patterson Lake conglomerate of Maurer (1958) and McGroder and others (1990) is restricted to a small area near Winthrop. It comprises sandstone, argillite, and conglomerate rich in clasts of argillite and sandstone. It was considered by Barksdale (1975) and Trexler (1985) to be the basal member of the Virginian Ridge Formation. McGroder and others (1990) suggested that the Patterson Lake unit might pre-date the Pasayten Group. Our unpublished mapping indicates that the conglomerate-bearing unit at Patterson Lake, southwest of Winthrop, lies unconformably on silicic breccias of the Twisp Formation and is either faulted against or unconformably overlies rocks of the Slate Peak Member of the Virginian Ridge Formation. Much of this unit is maroon, and in many respects it is similar to parts of the Ventura Member of the Midnight Peak Formation. Rocks northwest of Winthrop that have been called Patterson Lake conglomerate appear to be a different unit which may be part of the Albian and older (sub-Harts Pass) package.

The age of the Pasayten Group has been unclear. A single horizon in the Slate Peak Member of the Virginian Ridge Formation yields a snail fauna suggested to be

Cenomanian to Turonian (early Late Cretaceous) (Barksdale, 1975). Leaves from the Winthrop have been assigned to the latest Albian–earliest Cenomanian (Crabtree, 1987). The Patterson Lake unit has been called Albian (McGroder and others, 1990) on the basis of fossils collected from rocks northwest of Winthrop that are probably not correlative with the conglomeratic strata around Patterson Lake. A Late Cretaceous (Cenomanian–Coniacian) age for all of the Pasayten Group seems most likely, on the basis of (1) the time needed to develop a basin-wide unconformity that separates Pasayten Group rocks from underlying strata that are middle Albian and younger and (2) a probable Turonian fauna (W. P. Elder, USGS, written commun., 1993) collected from the Pasayten Group west of the Pasayten River where it crosses the International Border. The Midnight Peak Formation is at least in part older than about 88 Ma (Coniacian on time scale of Harland and others, 1990), as it is intruded by the Pasayten and Fawn Peak plutons of that age (Tabor and others, 1968; Riedell, 1979; Stoffel and McGroder, 1990). Elsewhere the unit is as young as 87 Ma (K-Ar age on hornblende separated from andesite, L. W. Snee, USGS, written commun., 1992).

SUPRACRUSTAL UNITS OF POSSIBLE AFFINITY TO THE METHOW TERRANE IN THE ROSS LAKE FAULT ZONE

Clastic and volcanic rocks that show many lithologic similarities to strata of the Methow basin, but that are at higher metamorphic grade, occur within the Ross Lake Fault Zone west of the Hozameen and North Creek faults (Fig. 2). The structural complexity of these rocks increases from southeast to northwest. Numerous intrusive bodies invade the strata on the northeast side of the Black Peak batholith. We describe these rocks from southeast to northwest.

North Creek Volcanics of Misch (1966)

The North Creek Volcanics (Misch, 1966) consist of volcanic and clastic sedimentary rocks that are generally metamorphosed to the greenschist facies. They form a wedge-shaped outcrop belt that is bounded on the southwest by the Twisp River fault and on the northeast by the North Creek fault and is intruded on the northwest by the Eocene (~50 Ma) Golden Horn and Late Cretaceous (~90 Ma) Black Peak batholiths (Fig. 2).

The North Creek Volcanics are strongly deformed in the Twisp River Fault Zone, particularly near the intersection with the North Creek fault. Much of the unit away from these structures appears to form an upright, northeast-dipping, homoclinal section (DiLeonardo, 1991), although the widespread development of cleavage in fine-grained rocks hints at a more complex structure.

The clastic rocks range from shale to conglomerate. Siltstone and fine-grained sandstone are particularly widespread, are commonly thinly interbedded, and display cross bedding in places. The sandstones are mostly arkosic (Misch, 1966) but also include lithic varieties (McGroder and others, 1990). Slate and conglomerate are common near the intersection of the Twisp River fault and North Creek fault. Clast types in the conglomerates are varied. Some conglomerates contain abundant shale fragments and lesser amounts of vein quartz, whereas others are dominated by chert or volcanic fragments.

Volcanic and volcaniclastic rocks of intermediate composition (probably andesite) are also common in the unit. They include breccias, massive flows, and tuffs, some of which have been reworked. Hornblende phenocrysts are present in many breccias and flows, and plagioclase phenocrysts occur in some rocks. The volcanic and volcaniclastic rocks are interbedded with siltstone and shale in some locations, whereas a thick section of massive volcanic rocks is present in the northwestern part of the belt, and thick sections of tuffaceous rocks occur elsewhere.

Deposition of the sedimentary and volcanic strata in the North Creek unit most likely occurred on the flanks of a volcanic arc and in an adjacent subaerial to possibly shallow marine basin. We base this interpretation largely on the intermediate composition of the volcanic rocks, the abundance of breccias and tuffs, and the lack of evidence, such as turbidite structures, for deep-water deposition.

The age and correlation of the North Creek Volcanics are problematic. Misch (1966) considered the unit to be Late Paleozoic or pre-Late Jurassic, but did not specify his evidence. In contrast, Barksdale (1975) and Kriens and Wernicke (1990) correlated the unit with the Jurassic–Lower Cretaceous Newby Group. We concur with Misch (1966) and McGroder and others (1990) that the North Creek Volcanics are not correlative with the Newby Group. Rocks assigned by Barksdale (1975) to the Newby Group include at least three different sets of strata: the Buck Mountain, Twisp, and undifferentiated Newby units mentioned previously in the section "Methow basin". Hornblende-phyric volcanic rocks of the North Creek unit are unlike the volcanic rocks of the undifferentiated Newby unit west of Twisp in which hornblende is generally absent. Furthermore, arkosic sandstone and chert pebble-bearing conglomerate of the North Creek Volcanics lack counterparts in the Buck Mountain and Twisp units. The North Creek Volcanics show some similarities to the Upper Cretaceous Midnight Peak Formation of the Pasayten Group (C. G. DiLeonardo, 1990, unpub. report to Washington Division of Geology and Earth Resources). Most flows in the Midnight Peak Formation, however, contain pyroxene rather than hornblende as their dominant mafic phase, and the red beds of the Ventura Member of the Midnight Peak Formation have not been recognized in the North Creek Volcanics. In summary, the association of arkose and volcanic rock of intermediate composition in the North Creek Volcanics is broadly similar to sequences in the Methow basin, but the North Creek cannot be conclusively correlated with any specific unit. Thus, the possibility that the unit is a far-traveled slice along the Ross Lake Fault Zone cannot be ruled out.

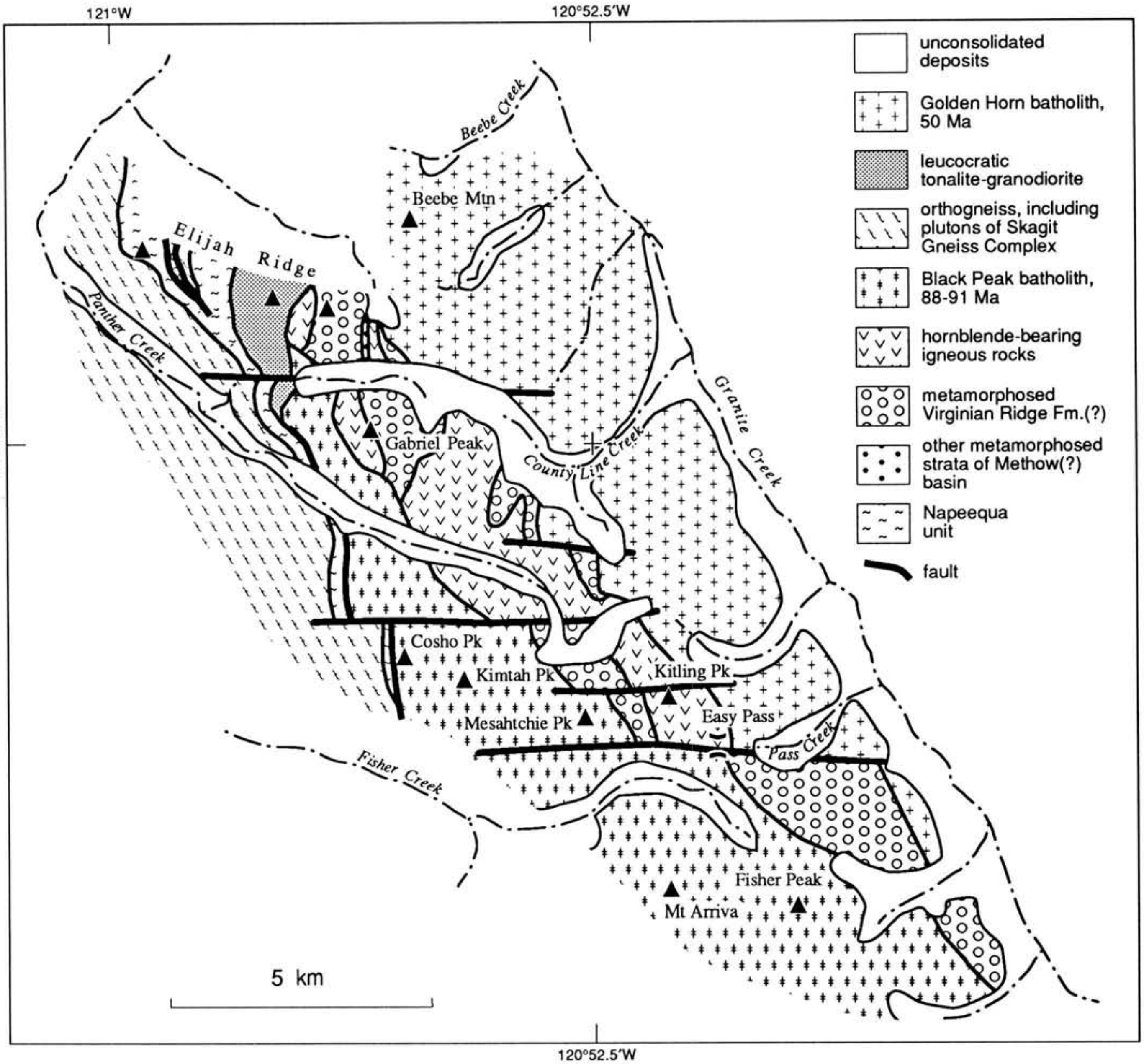

Figure 4. Sketch map of bedrock geology of the Gabriel Peak–Elijah Ridge area.

Rocks of the Gabriel Peak–Elijah Ridge Area

Metasedimentary and meta-igneous rocks exposed in the region between the Black Peak batholith and the Golden Horn batholith (Figs. 2 and 4), extending from Easy Pass northwest to Panther Creek, can be assigned to several lithologic packages, some probably correlative with Methow basin strata to the east. Misch (1966) mapped these rocks as his North Creek Volcanics and Elijah Ridge Hornfelsic Schist. Kriens and Wernicke (1990) assigned them to their heterogeneous orthogneiss unit, their amphibole porphyry unit, their Iandi porphyry unit, and the Newby Group of Barksdale (1975).

Metamorphosed Virginian Ridge Formation(?)

Distinctive metaconglomerate with quartzose clasts and associated metasandstone and minor metapelite crop out sporadically from eastern Elijah Ridge south to a few kilometers southeast of Easy Pass (Fig. 4). Most of the rocks are deformed to L-S tectonites and recrystallized to biotite-bearing quartzite or cordierite-spotted hornfels. Recrystallization is too extensive to confidently identify the clasts as chert pebbles, but observed bed thicknesses of one to several meters, a high proportion of metaconglomerate, and association with metapelite strongly suggest that these rocks are metamorphosed Devils Pass Member of the Virginian Ridge Formation.

A layer of garnet-staurolite schist (meta-argillite) with lenses of garnet-hornblende-plagioclase granofels (metamorphosed limy concretions) and minor metaconglomerate with quartzose clasts on western Elijah Ridge probably also correlates with the Virginian Ridge Formation. Proportions of various lithologies suggest that this is the Slate Peak Member. These schists are presently overlain by hornblende-bearing schist, ultramafite, and metagabbro we assign to the Napeequa unit.

Metamorphosed heterolithic conglomerate and associated metapelite

Several hundred meters of the crest of eastern Elijah Ridge are underlain by metaconglomerates with diverse clast lithologies and abundant metapelites. Clasts include granite, felsic and intermediate volcanic rock, and minor quartzite/metachert. Locally these rocks bear staurolite, garnet, cordierite, and andalusite. Their close association with metamorphosed Virginian Ridge Formation(?) suggests that these rocks are also metamorphosed Methow basin strata. Possible protoliths include the Panther Creek Formation or parts of the Harts Pass Formation.

Hornblende-bearing igneous rocks

The metasedimentary rocks are intruded by a texturally diverse suite of hornblende-bearing rocks. Most common are hornblende porphyries with a greenish matrix; where strongly deformed, these have been transformed into centimeter- to meter-thick layers of mafic schist. Elsewhere the metasedimentary rocks are intruded by equigranular hornblende gabbros. The hornblende porphyries are intruded by, and hence are older than, the approximately 90 Ma Black Peak batholith, yet must be younger than Cenomanian or Turonian if the metasedimentary rocks are indeed the Virginian Ridge Formation. The presence of phyric hornblende is reminiscent of the North Creek Volcanics to the southeast.

Talus east of Easy Pass includes abundant blocks of metaconglomerate of the metamorphosed Virginian Ridge Formation(?) and blocks of hornblende-phyric volcanic breccia. The breccias could be an extrusive equivalent of the hornblende porphyries. The association of these lithologies raises the possibility that the hornblende-phyric igneous rocks are penecontemporaneous with the conglomerates, though interbedding of volcanic breccia and metaconglomerate has not been seen in talus blocks or on the outcrop.

SUPRACRUSTAL ROCKS OF UNCERTAIN AFFINITY IN THE FOGGY DEW FAULT ZONE

The belt of metamorphic rocks with protoliths similar to strata of the Methow basin is cut out southeast of the intersection of the North Creek fault, Twisp River fault, and Foggy Dew fault (Fig. 2). Mylonitized supracrustal rocks lie within the ~1-km-wide Foggy Dew Fault Zone, which separates the Methow basin from the Oval Peak batholith (~65 Ma, Miller and Bowring, 1990) and the Twisp Valley Schist of the Cascades core (Fig. 2). These rocks have been described by Miller and Bowring (1990), and we briefly summarize them here.

Amphibolite, greenschist, biotite schist, and mylonitic gneiss derived at least in part from the Oval Peak batholith are the dominant constituents of the fault zone. Local amphibolites displaying relict coarse-grained igneous texture are probably in part metamorphosed gabbros, but the typically small grain size of the amphibolites and greenschists suggests that most are metamorphosed volcanic rocks. Whole-rock major and trace element abundances, including those of the rare earth elements, indicate that the protolith was basaltic and probably island-arc tholeiite (Geary and Christiansen, 1989; Miller and others, 1993b). The biotite schists were mostly fine-grained sandstone and siltstone with minor shale. The protolith age for these supracrustal rocks is greater than 65 Ma, the age of the Oval Peak batholith which intrudes amphibolite in a few localities (Miller and Bowring, 1990).

We cannot confidently correlate the supracrustal rocks in the fault zone with any other units in the region. The chemistry of the metabasalts is unlike that of the nearby Twisp Valley Schist, which is described below, and metachert and calc-silicate rock characteristic of the Twisp Valley Schist are absent (Miller and others, 1993b). The Cascade River unit of the Chelan Mountains terrane contains considerable clastic rock, but the metaconglomerate of that unit is absent in the Foggy Dew Fault Zone. Furthermore, metavolcanic rocks in the Cascade River unit range from felsic to mafic and are predominantly intermediate in composition (for example, Tabor and others, 1989; Brown and others, 1993), in contrast to the island-arc tholeiites of the fault zone. Protoliths of the metasedimentary rocks in the Foggy Dew Fault Zone are similar to rocks in the Methow basin (for example, Twisp Formation and Lower and Middle Jurassic Ladner Group), but metavolcanic rocks in the basin are dominantly intermediate in composition (for example, Barksdale, 1975). Better candidates for Methow rocks are local intermediate to felsic metavolcanic rocks in the fault zone (Miller and Bowring, 1990). These rocks, however, have not been chemically analyzed. Triassic basalts that form the basement to at least part of the Methow terrane (Ray, 1986) display the chemistry of mid-ocean-ridge basalts, in contrast to the island-arc tholeiites in the Foggy Dew Fault Zone. In summary, the association of supracrustal rocks in the fault zone is distinctive, and these rocks may be highly allochthonous.

SUPRACRUSTAL ROCKS OF THE CHELAN MOUNTAINS TERRANE IN THE ROSS LAKE FAULT ZONE

Twisp Valley Schist

The Twisp Valley Schist (Adams, 1961) represents the largest continuous exposure of supracrustal rocks in the northeastern part of the Cascades core. The schist ranges in metamorphic grade from the greenschist facies to the sillimanite zone of the amphibolite facies. The Foggy Dew and

Twisp River faults separate it from lower grade rocks of the Methow basin and the North Creek Volcanics, respectively (Libby, 1964; Miller and Bowring, 1990). The southwestern contact of the schist with Cretaceous and Paleocene orthogneiss of the Skagit Gneiss Complex is marked by the high strain zone of the Gabriel Peak tectonic belt (Libby, 1964; Miller, 1987; Miller and Bowring, 1990; Miller, 1994). The Black Peak batholith and Oval Peak batholith intrude the Twisp Valley Schist and provide the only age control on this pre-90 Ma unit.

The Twisp Valley Schist is an assemblage of dominantly siliceous schist, considerable metabasite (amphibolite and greenschist), calc-silicate rock and marble, and minor metaperidotite and metasandstone (Miller and Bowring, 1990; Miller and others, 1993b). Siliceous schists, which were probably originally cherts and siliceous mudstones, dominate large parts of the unit. They locally grade into, and are interlayered with, more pelitic rocks that contain porphyroblasts of garnet, staurolite, and locally andalusite, sillimanite, and kyanite. Sections containing significant amounts of metamorphosed fine-grained sandstone, siltstone, and shale occur along the east side of the outcrop belt (Miller and others, 1993b). Widespread lenses of metabasite typically are less than 25 m thick, and calc-silicate rock and marble also generally occur as small pods. Metaperidotites are concentrated near the western contact of the unit in the Gabriel Peak tectonic belt. The small size (<20 m wide) of the ultramafites and their intercalation with a wide range of supracrustal rock types lead to the inference that they are pre- or syn-metamorphic fault slices (Miller and others, 1993b).

Analysis of rare earth and other immobile trace elements suggests that the Twisp Valley Schist contains both mid-ocean ridge and oceanic island basalts that occur in separate belts (Miller and others, 1993b). The inferred tectonic setting of the basaltic protoliths, combined with the overall association of rock types, is compatible with deposition in either an open ocean or marginal basin. The oceanic island basalts and their commonly associated calc-silicate rocks and marbles probably represent seamount(s) and accompanying sediments that formed in the marine basin (Miller and others, 1993b). The complex intercalation of rock types suggests that the unit was tectonically disrupted before Cretaceous and Paleogene metamorphism, perhaps during underplating in an accretionary wedge.

The Twisp Valley Schist is lithologically very similar to the Napeequa unit of the Chelan Mountains terrane, and we consider the two to be correlative (Tabor and others, 1989; Miller and others, 1993b). Rock types in the Twisp Valley Schist are also similar to those in the weakly metamorphosed Hozameen Group (Misch, 1966) of the Bridge River–Hozameen terrane, which lies structurally above metamorphosed Methow-like rocks (Fig. 1) and extends northward into British Columbia (for example, McTaggart and Thompson, 1967; Haugerud, 1985; Ray, 1986). The association of oceanic island basalts and limestones in the Bridge River–Hozameen terrane (Haugerud, 1985; Ray, 1986) provides further support for this correlation (Miller and others, 1993b), and by extension, the Napeequa unit as a whole probably correlates with this terrane. Radiolaria recovered from the Hozameen Group are Permian to Middle Jurassic (Haugerud, 1985), but the correlative Bridge River Complex extends back to the Mississippian (Cordey and Schiarizza, 1993), and the protoliths of the Twisp Valley Schist may similarly have accumulated during this interval.

Rainbow Lake Schist

A narrow, typically less than 100 m thick, discontinuous belt of metasedimentary and metavolcanic rocks extends for about 35 km within the Gabriel Peak tectonic belt. These rocks lie nearly on strike with the western boundary of the Twisp Valley Schist, and the two units occupy a similar structural position at the northeastern contact of the Skagit Gneiss Complex (Figs. 2 and 5). For most of its length, the supracrustal belt is in tectonic contact with the southwestern margin of the Black Peak batholith. Adams (1961) mapped a nearly continuous segment of metasupracrustal rocks in the southern part of the belt and named it the Rainbow Lake Schist. On the basis of our mapping and that of Peter Misch (oral commun. and unpub. map), we extend this unit northwestward and apply Adams' term to the entire belt.

The Rainbow Lake Schist is a heterogeneous sequence in which individual lithologies commonly form laterally discontinuous bodies that are typically less than 3 m thick (Adams, 1961). The dominant rock type is quartz-rich biotite schist, some of which contains porphyroblasts of garnet and staurolite. Andalusite and sillimanite occur in rare metapelite. The siliceous schist grades into metachert. Other clearly metasedimentary rocks are rare calc-silicate rock and marble. Amphibolites are widespread, and their fine grain size combined with reconnaissance geochemical data (R. B. Miller and L. S. Nicholson, unpub. data) indicate that they are probably metabasalts. Adams (1961) noted that some amphibolites are interlayered on a fine scale with siliceous biotite schists and are probably metamorphosed mafic tuffs rather than flows. Hornblende-biotite schists are also present in significant amounts; these may be derived from immature graywackes or from volcanic rocks of intermediate composition. Local small (<3 m thick) pods of ultramafic rock (metaperidotite and talc schist) probably represent pre- or syn-metamorphic fault slices within the unit.

Distinctive muscovite-bearing, leucocratic, and commonly pegmatitic sills and dikes intrude the Rainbow Lake Schist and are deformed with it. In some areas, numerous intrusive sheets of orthogneiss similar to that in the Skagit Gneiss Complex invade the schist. We mapped the contact with the Skagit at the southwestern limit of schist rafts in the orthogneisses.

The Rainbow Lake Schist records polyphase folding and multiple generations of foliation and lineation. Mylonitic structures are well developed in more quartz-rich and micaceous lithologies and include S-C fabrics and extensional crenulation cleavage that generally record dextral shear of the Gabriel Peak tectonic belt.

The overall association and mixing of rock types in the Rainbow Lake Schist is similar to that of the Twisp Valley Schist, and the two units are probably correlative (Miller and others, 1993b). The main differences between the units are that hornblende-biotite schist is present in significant quantities in the Rainbow Lake Schist, but is scarce in the Twisp Valley Schist, and calcareous rock is much less common in the Rainbow Lake unit. Like the Twisp Valley Schist, the Rainbow Lake Schist probably represents a deep marine assemblage that underwent tectonic mixing before or during initiation of metamorphism. The presence of metamorphosed immature graywacke or intermediate volcanic rock (hornblende-biotite schist) hints at proximity to a volcanic arc during deposition.

Napeequa Unit in the Gabriel Peak–Elijah Ridge Area

Hornblende-bearing schists on westernmost Elijah Ridge (Fig. 4), with minor ultramafite and locally extensive gabbro, appear to be continuous with similar schists on eastern Ruby Mountain that Tabor and others (1989) assigned to their Napeequa unit. These rocks are nearly on strike with the Rainbow Lake Schist, which contains similar rock types, and these units are probably correlative.

INTRUSIONS INTO THE NAPEEQUA UNIT AND UNITS OF METHOW AFFINITY ON ELIJAH RIDGE

Leucocratic gneissic biotite tonalite and granodiorite on central Elijah Ridge intrude early Late Cretaceous metagabbro to the east and metagabbro, meta-ultramafite, and hornblende-bearing schist of the Napeequa unit to the west (Fig. 4). The tonalite and granodiorite are characterized by scattered phenocrysts of K-feldspar, layers and lenses of greenish mafic schist that are metamorphosed mafic dikes, and a variously developed, mylonitic L-S fabric.

All mapped units in the Gabriel Peak–Elijah Ridge area are intruded by orange-weathering, unmetamorphosed quartz-two feldspar porphyry dikes related to the Golden Horn batholith. Less distinctive, but also abundant, are greenish plagioclase porphyry dikes. The leucocratic gneissic tonalite and granodiorite on central Elijah Ridge are intruded by small bodies of directionless biotite-hornblende granitoid that are probably correlative with parts of the Ruby Creek Heterogeneous Intrusive Complex of Misch (1966) (Fig. 5).

SKAGIT GNEISS COMPLEX

The least studied part of our transect lies between the Gabriel Peak tectonic belt on the northeast and the Holden assemblage (Fig. 2) on the southwest. Much of this area has been mapped as Skagit Gneiss (Adams, 1961; Libby, 1964), and it lies on strike to the southeast of the Skagit Gneiss Complex (Haugerud and others, 1991; Misch, 1966). Our mapping of a few areas in detail and at a reconnaissance level over a larger region, combined with our interpretation of previous studies, suggest that this tract is dominated by orthogneiss.

In the Gabriel Peak tectonic belt, and for at least several kilometers to the southwest, the orthogneisses range in composition from leucogranodiorite to hornblende quartz diorite. They are mostly variously deformed and metamorphosed biotite leucogranodiorites and biotite trondhjemites (Wade, 1988; Miller and Bowring, 1990) that in general are more felsic than the tonalitic orthogneisses (Misch, 1966) that dominate elsewhere in the Cascades core. Most are sheet-like bodies, and individual sheets commonly range from 1 to 10 m in thickness (Miller, 1992). The widespread sheeting imparts a distinctly heterogeneous appearance to the rocks. The relatively mafic sheets are typically intruded by the more felsic varieties.

A few of the orthogneisses have been dated by the U-Pb (zircon) method. Two tonalitic orthogneisses in the Gabriel Peak tectonic belt yield ages of 87 Ma and 68 Ma (Hoppe, 1984: Miller and others, 1989; Miller and Bowring, 1990) that are interpreted to be the crystallization age of the parent tonalites. Many of the volumetrically dominant felsic orthogneiss sheets resemble the Lake Juanita leucogneiss of the tectonic belt, which has a probable crystallization age of about 60 Ma (Miller and Bowring, 1990). The orthogneisses are also intruded by small, unfoliated, weakly metamorphosed to unmetamorphosed bodies that range from K-feldspar phyric quartz monzonite, reminiscent of parts of the Eocene (48 Ma) Cooper Mountain batholith (Miller and Bowring, 1990), to diorite.

We have observed many tens of rafts of supracrustal rocks within the orthogneisses in and a few kilometers southwest of the Gabriel Peak tectonic belt. Many of the rafts are less than 2 m thick, but the largest raft reaches nearly 20 m in thickness and extends for more than a kilometer (Libby, 1964). Clusters of rafts less than 1 m in width make up more than 50 percent of the outcrop in several areas more than 30 m across.

The rafts consist mainly of interlayered quartz-plagioclase-biotite schist, calc-silicate schist, amphibolite, and hornblende-biotite schist with rare quartzitic schist and nearly pure marble (Libby, 1964; Wade, 1988). The calc-silicate rocks form a distinctive lithology in this part of the Skagit Gneiss Complex. They have highly varied modal abundances but are characterized by significant amounts of diopside (10–70 percent) and plagioclase. Hornblende and titanite are typically present, and garnet, calcite, and quartz are found in many samples. The protolith for these calc-silicate rocks may have been calcareous mudstone or argillaceous limestone interlayered with argillaceous or psammitic sediment (now biotite schist) and basalt (now amphibolite). The hornblende-biotite schist is probably metamorphosed graywacke or volcanic rock of intermediate composition.

The overall association of rock types in the rafts is different than that of the Twisp Valley Schist, Rainbow Lake

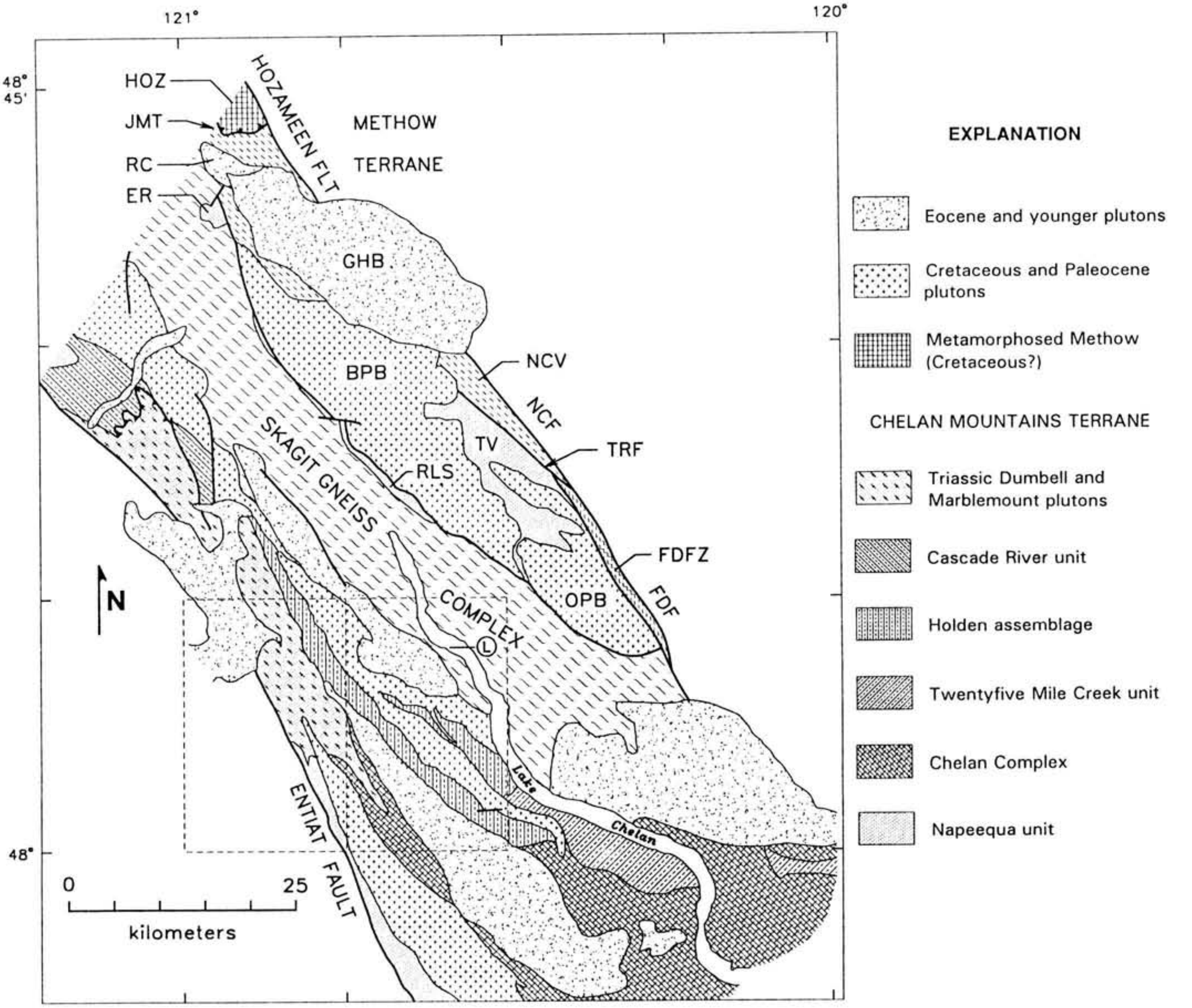

Figure 5. Sketch map of the northeastern Cascades. BPB, Black Peak batholith; ER, Elijah Ridge; FDFZ, Foggy Dew Fault Zone; GHB, Golden Horn batholith; HOZ, Permian–Jurassic Hozameen Group; JMT, Jack Mountain thrust (Cretaceous); L, Lucerne; NCF, North Creek fault; NCV, North Creek Volcanics; OPB, Oval Peak batholith; RC, Ruby Creek heterogeneous plutonic belt of Misch (1966) (Eocene); RLS, Rainbow Lake Schist; TRF, Twisp River fault; TV, Twisp Valley Schist. Boxes defined by dashed lines show location of Lucerne (east) and Holden (west) quadrangles discussed in text. Modified from Haugerud and others (1991) and Miller (1994).

Schist, and other Napeequa unit-like rocks in and near the Ross Lake Fault Zone. In particular, quartzitic schists are much less common in the rafts, the calc-silicate rocks in the rafts contrast with the purer marbles of the Napeequa-like units (Libby, 1964), and ultramafites have not been observed in the rafts. These differences lead us to infer that the rafts represent a separate unit and that they were juxtaposed with the Napeequa-like units by movement along the Gabriel Peak tectonic belt. The rafts may be part of the Cascade River unit, which is characterized by metamorphosed sandstones and intermediate volcanic rocks and contains some carbonates (Tabor and others, 1989). We note that the rafts most closely resemble the "younger gneissic rocks of the Holden area" (Cater and Crowder, 1967; Cater and Wright, 1967), which may also correlate with the Cascade River unit. We more fully address possible correlations in a later section.

Farther southwest in the Lucerne 15-minute quadrangle (Figs. 5 and 6) rocks on strike with the Skagit Gneiss Complex were mapped as Swakane Gneiss by Cater and Wright (1967). This correlation has been questioned in several recent compilations because the Swakane in its type area to the southwest (Fig. 1) is on the average more aluminous than the Skagit Gneiss Complex and has a more uniform appearance. The Swakane also contains Precambrian zircons (Mattinson, 1972) and has high initial Sr values (for

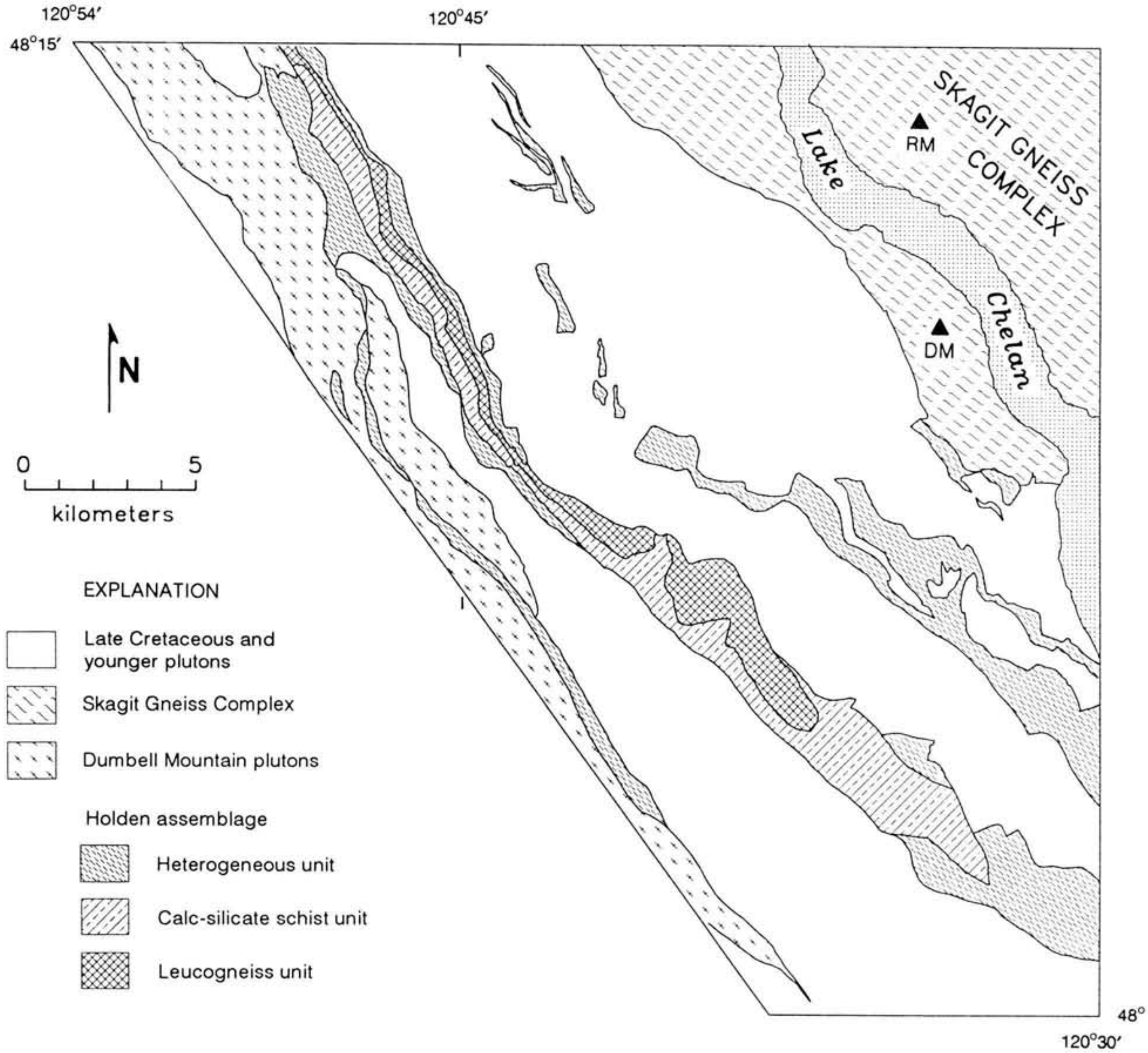

Figure 6. Sketch map emphasizing units of the Holden assemblage in parts of the Lucerne and Holden quadrangles. DM, Domke Mountain; RM, Round Mountain. Modified from Cater and Crowder (1967) and Cater and Wright (1967).

example, Ford and others, 1988) that are not readily compatible with the dominantly arc-like and oceanic-type terranes that are the protoliths for the Skagit Gneiss Complex (for example, Tabor and others, 1987a, 1989; Haugerud and others, 1991). In its type area, the Swakane Gneiss is mostly uniform biotite gneiss that was originally either an arkosic sequence (Waters, 1932) with a dominantly Middle Proterozoic source (Rasbury and Walker, 1992) or a dacitic volcanic complex (C. A. Hopson, *in* Mattinson, 1972; Cater, 1982).

We have examined the rocks designated as Swakane Gneiss by Cater and Wright (1967) only in reconnaissance (transect from Round Mountain to Domke Mountain, Fig. 6). This transect near its northeastern limit crosses a complex of dominantly biotite tonalite to granodiorite orthogneiss; most individual bodies are less than 50 m thick. To the southwest an approximately 300-m-wide zone consists of biotite paragneiss with local, garnet-bearing, impure quartzite. These metasupracrustal rocks have been pervasively intruded by orthogneiss and by sills, dikes, and irregularly shaped masses of pegmatite. As a result, areas of injection migmatite are well developed. Pegmatite is progressively more abundant to the southwest. Several rock types occur as screens within the pegmatite, including abundant biotite leucogneiss, hornblende-biotite quartz diorite and tonalite orthogneiss, medium-grained biotite tonalite/granodiorite that displays a weak to moderately developed linear fabric, and biotite paragneiss.

Our observations are in general similar to those of Webb (1957) who mapped about 13 km^2 northeast of Lake Chelan. Webb (1957) described "gneissic biotite granodiorite", which is presumably equivalent to the granodiorite orthogneiss we mapped, as the most common rock type. Sills of finer grained granodiorite to quartz monzonite, 3 to 15 m in width, and dikes of pegmatite intrude the gneissic granodiorite. Biotite-garnet schist and amphibolite form narrow lenses within the gneissic granodiorite (Webb, 1957) and presumably represent inclusions of wall rock.

Southwest of Lake Chelan, medium-grained quartz-plagioclase-biotite ± hornblende ± garnet gneiss and less abundant hornblende gneiss that grade into amphibolite are injected by leucocratic igneous rock. A coarse-grained biotite granodiorite/tonalite orthogneiss, more than 200 m wide, intrudes the paragneiss and is in turn cut by numerous pegmatite veinlets and dikes, typically 10–20 cm thick.

In summary, the unit mapped as Swakane Gneiss in the Lucerne quadrangle by Cater and Wright (1967) consists of variously metamorphosed plutonic rocks that contain screens of biotite paragneiss, probably derived from metaclastic rock, and less abundant hornblende gneiss and amphibolite that are presumably metamorphosed intermediate and (or) mafic volcanic rocks. The sparsity of paragneiss and its commonly migmatitic appearance preclude any simple correlation of its protolith. The position of the unit on strike with the migmatitic Skagit Gneiss Complex, combined with the sparsity of metachert, are perhaps most compatible with its being Cascade River unit that was extensively injected by the orthogneiss protolith.

HOLDEN ASSEMBLAGE

A heterogeneous series of supracrustal rocks metamorphosed to the amphibolite facies was mapped by Cater and Crowder (1967) and Cater and Wright (1967) in the Holden and Lucerne 15-minute quadrangles (Figs. 5 and 6). They referred to these rocks as the "younger gneissic rocks of the Holden area". We suggest that this is a somewhat misleading name because a significant percentage of these rocks are not gneissic and, although they may structurally overlie rocks mapped as Swakane Gneiss by Cater and Wright (1967), their original age relative to other gneissic units in the area is not clear. We thus use the informal name Holden assemblage in the following discussion.

The Holden assemblage is dominated by hornblende-bearing rocks that include amphibolite, hornblende gneiss, and hornblende-biotite schist. Calc-silicate rock, leucocratic gneiss, and plagioclase-biotite schist are less abundant constituents, and marble, pelitic schist, and metaconglomerate occur locally. On the basis of our ongoing studies, we subdivide the Holden assemblage into three major units; this represents a minor deviation from divisions defined by previous workers. Our units are the (1) heterogeneous unit, (2) calc-silicate schist unit, and (3) leucogneiss unit. They form northwest-trending belts that commonly occur as septa between elongate intrusive bodies of Triassic, Cretaceous, and Eocene age (Figs. 5 and 6) (Cater and Crowder, 1967; Cater and Wright, 1967). Dikes and sills ranging from granite to gabbro, and from unmetamorphosed to strongly metamorphosed, invade the assemblage and are particularly abundant in its southeastern part.

Heterogeneous Unit

These diverse rocks make up the largest unit in the Holden assemblage. We include in this unit rocks mapped as "heterogeneous gneiss, schist, and quartzite" and "hornblende schist and gneiss" by Cater and Crowder (1967) and Cater and Wright (1967). We do not follow the usage of these workers because their units are not lithologically distinct. Hornblende gneiss, amphibolite, and hornblende-biotite schist and gneiss dominate, but biotite schist, biotite gneiss, and calc-silicate (diopside-biotite-quartz) schist are also abundant. Layers of marble, calcareous gneiss, and quartzite are widespread, but they make up less than 5 percent of the unit. Rare constituents are pelitic schist, granitoid gneiss, hornblendite, and metaconglomerate with granitoid and quartz pebbles in a biotite-hornblende gneiss matrix.

The interlayering of the hornblende-bearing rocks with calcareous rock, pelite, metaconglomerate, and biotite schist suggests that the hornblende-bearing rocks are at least in part derived from sedimentary and (or) volcanic rocks, although some may have been sills or dikes. The mineralogy and modes of the amphibolites and hornblende-plagioclase gneisses are compatible with a basalt or basaltic andesite parent. The hornblende-biotite schists are of less certain origin. In some localities, they are interlayered with biotite schist on the scale of 4–8 cm, suggesting to us an original sequence of interbedded tuffs and clastic sediments. In addition, some of the hornblende-biotite schists may be metamorphosed flows or immature graywackes. Coarser grained hornblende-biotite gneisses, which increase in abundance to the southeast, are probably metatonalites that intruded the supracrustal rocks before or during metamorphism.

Fairly pure marbles reach approximately 20 m in thickness. Many of the marbles mapped by Cater and Crowder (1967) and Cater and Wright (1967) are actually calc-silicate rocks and are far more abundant than the pure marbles. The calc-silicate schists and gneisses probably have a range of protoliths, as they consist of highly varied proportions of diopside, quartz, tremolite, plagioclase, titanite, epidote-clinozoisite, biotite, grossularite, and calcite. These rocks may represent calcareous siltstones and mudstones that were interbedded, in places on the scale of a few tens of centimeters to several meters, with the inferred tuffaceous and siliciclastic rocks.

Calc-Silicate Schist Unit

We follow Cater and Crowder (1967) and Cater and Wright (1967) by mapping a unit containing abundant clinopyroxene (diopside)-biotite-quartz schist that extends discontinuously for most of the length of the Lucerne and Holden quadrangles (Fig. 6). This unit is distinctive in the field because of its sulfide-induced rusty weathering. However, it contains most of the rock types found in the heterogeneous unit, and it is distinguished from that unit by the higher percentage of calc-silicate schist and biotite schist.

Leucogneiss Unit

The leucogneiss unit (biotite gneiss unit of Cater and Crowder, 1967, and Cater and Wright, 1967) lies between the calc-silicate schist unit on the southwest and the heterogeneous unit on the northeast (Fig. 6). It has a structural thickness between 300 and 500 m. The leucogneiss unit is dominated by medium-grained gneiss consisting mainly of quartz (~60 percent), plagioclase (~30 percent), and biotite (~10 percent) with accessory garnet, titanite, epidote/clinozoisite, and white mica. Interlayered with the leucogneisses are less abundant, but widespread amphibolites that range from several centimeters to about 30 m in thickness.

The leucogneisses lack relict structures, and their protoliths are difficult to ascertain. We note that if metamorphism was isochemical, a sedimentary (impure quartzose sandstone?) parent is indicated by the abundance of quartz and by the high SiO_2 contents (77–78 percent) in two analyzed samples of leucogneiss (R. B. Miller, unpub. data). In contrast, the medium grain size and relative homogeneity of the leucogneisses are compatible with a volcanic (rhyolite?) protolith (C. A. Hopson, *in* Mattinson, 1972). The amphibolites presumably represent basalt (or basaltic andesite) flows, sills, or dikes. If the leucogneisses are metarhyolites, and if the leucogneiss and amphibolite protolith were coeval, the overall association suggests a bimodal volcanic suite.

Contact Relations and Correlations

The rocks of the Holden assemblage are strongly foliated and lineated, transposed, and in places mylonitic. Despite this intense deformation, the intimate interlayering of rock types suggests that the assemblage was originally a single depositional sequence. A possible exception is the leucogneiss, which forms a discrete belt and is only interlayered with amphibolite. The leucogneiss unit may record a major change in provenance and (or) in volcanism within the depositional sequence. Alternatively, it may be a pre- or synmetamorphic fault slice. There is no obvious repetition of units, however, as might be expected from imbricate thrusting or large-scale folding.

Cater (1982) stated that rocks of the Holden assemblage are intruded by the Late Triassic (~220 Ma; Mattinson, 1972) quartz dioritic and tonalitic Dumbell Mountain plutons. We concur with this observation and note that xenoliths of amphibolite exceeding 2 m in diameter occur in these plutons and that stringers of intrusive rock extend into the amphibolitic wall rock. However, the plutonic rocks and amphibolites display parallel foliations and lineations, in contrast to Cater's (1982) assertion that dynamothermal metamorphism of the amphibolites predated intrusion of the Dumbell bodies.

Mattinson (1972) estimated a crystallization age of 265 ± 15 Ma for zircons in the leucogneiss unit. If the leucogneiss is metamorphosed quartz keratophyre, as suggested by C. A. Hopson (*in* Mattinson, 1972), then the protolith age for part of the sequence may be Permian. This age is compatible with the constraint that the Holden unit is Late Triassic or older. However, the zircon age is based on discordant U-Pb analyses of a single fraction, at least some (all?) of these zircons may be detrital, and thus the interpretation that the protolith age is Permian is speculative.

The Holden assemblage is continuous along strike to the southeast with the migmatitic Chelan Complex and the Twentyfive Mile Creek unit (Fig. 5) (Tabor and others, 1987b). The increase in orthogneiss to the southeast within the heterogeneous unit is compatible with the higher metamorphic grade and greater percentage of plutonic rock in the Chelan Complex (for example, Hopson and Mattinson, 1971; Tabor and others, 1987a). The protoliths for supracrustal rocks in the Chelan Complex are apparently similar to those for the Holden rocks (Tabor and others, 1987a, 1987b). The Twentyfive Mile Creek unit, which consists of amphibolite, biotite schist, quartzitic schist, and rare marble, resembles the Holden assemblage but may contain more quartzitic schist.

Cater (1982) suggested that the Holden assemblage is correlative with the Napeequa unit, which lies across strike to the southwest (Fig. 5) in the Holden quadrangle. We do not agree with this correlation. The Napeequa unit consists dominantly of amphibolite and quartzite that probably represent an oceanic assemblage of metamorphosed basalt and chert according to Tabor and others (1988, 1989). Ultramafic rock is also widespread in the unit, and marble and calc-silicate rock occur locally. In the Holden assemblage, quartzite is scarce and ultramafite is absent. The Napeequa unit, in turn, contains much less hornblende-biotite schist than the Holden assemblage, and the leucogneiss-amphibolite unit is apparently not represented in the Napeequa unit.

We propose that the Holden assemblage is either correlative with the Cascade River unit of Tabor and others (1989) or, less likely, is a tectonostratigraphic unit with no direct correlative elsewhere in the Cascades crystalline core. Map relations suggest that the Cascade River unit is on strike with the Holden assemblage to the northwest (Fig. 5) (Libby, 1964; Grant, 1966; Dragovich, 1989; Tabor and others, 1989; Dougan, 1993). The Cascade River unit is a varied sequence of metavolcanic and metaclastic rocks that include hornblende-biotite schist, biotite schist, amphibolite, metaconglomerate with granitoid clasts, and minor calcareous rock (for example, Tabor and others, 1988, 1989). The lithologies within the Holden and Cascade River units are broadly similar, both are inferred to contain metamorphosed tuff and immature sandstone, and both are associated with Triassic plutons. Metavolcanic and calc-silicate rocks are apparently more common in the Holden unit, however, whereas metaconglomerates are much more abundant in the Cascade River unit. The granitoid clasts in the Cascade River metaconglomerates are probably derived from intrusions of the Late Triassic Marblemount Meta-Quartz Diorite (Misch, 1966; Tabor and others, 1989). These intrusions are the same age as, and lie on strike with, the Dumbell Mountain plutons that intrude the Holden assemblage. Metatuff in the Cascade River unit is nearly coeval with these Late Triassic intrusions (Tabor and others, 1988; Cary, 1990), and the Cascade River unit is probably intruded locally by the Marblemount plutons (Brown and others, 1993). These relations have been interpreted to indicate broadly coeval Triassic plutonism and volcanism, and erosion and unroofing of the plutons shortly after emplacement, while volcanism continued (for example, Tabor and others, 1988, 1989). If the Cascade River unit–Holden assemblage correlation is correct, then at least parts of the Holden assemblage may also broadly overlap in time Late Triassic plutonism.

Many workers have interpreted the Marblemount plutons and Cascade River unit to represent the intrusive roots and supracrustal cover of a magmatic arc (for example, Dragovich, 1989; Tabor and others, 1989; Cary, 1990; Brown and others, 1993). We similarly infer that the Holden assemblage formed in and adjacent to a magmatic arc on the basis of its abundant metamorphosed tuffaceous rocks of intermediate composition, clastic sedimentary rocks, local calcareous rocks, and association with possibly broadly coeval tonalite and quartz diorite plutons.

DISCUSSION

Are the Cascade River Unit and Napeequa Unit Separate Terranes?

Tabor and others (1988, 1989) inferred that the Cascade River unit and Napeequa unit are part of a single terrane, the Chelan Mountains terrane. They based this conclusion on several observations and interpretations: (1) the Cascade River unit contains granitoid clasts thought to be derived from the Late Triassic Marblemount Meta-Quartz Diorite (Misch, 1966); (2) the Cascade River unit probably unconformably overlies the Marblemount; (3) the Napeequa unit may have been intruded by the Late Triassic Dumbell Mountain plutons, which are of similar ages and lie in the same belt as the Marblemount plutons; and (4) the Cascade River unit may thus unconformably overlie both the plutons and the Napeequa unit. This conclusion is reasonable from a tectonic perspective in that an arc assemblage (Cascade River unit) is built on oceanic lithosphere (Napeequa unit) (Tabor and others, 1989).

Tabor and others' (1989) interpretation that the Napeequa unit was intruded by the Dumbell plutons (point 3) is based on Cater's (1982) description of intrusive contacts between these plutons and the Holden assemblage. Our interpretation that the Holden assemblage is part of the Cascade River unit, not the Napeequa unit, or is a previously unrecognized group of rocks, weakens the inferred premetamorphic stratigraphic ties within the Chelan Mountains terrane. Our interpretation instead suggests that the Marblemount–Dumbell intrusions are linked along the length of their outcrop belt to the Cascade River unit. Furthermore, if we have correctly correlated the Twisp Valley Schist and the Napeequa unit with the Mississippian to Jurassic Bridge River–Hozameen terrane (Tabor and others, 1989; Miller and others, 1993b), which consists of ocean-floor rocks and entrained oceanic islands, then the Napeequa–Twisp Valley rocks probably represent a terrane separate from the Triassic arc-type rocks of the Cascade River unit and Holden assemblage.

Our interpretations are in general accord with the observations of Brown and others (1993) from the Cascade River area along strike to the northwest of the Holden area. They note that the Cascade River unit and Napeequa unit are found only in tectonic contact and that the juxtaposition of these units probably occurred on pre-metamorphic low-angle faults. Brown and others (1993) concluded that the primary relations of these units remain uncertain. We similarly conclude that more work is needed to establish the nature of the pre-metamorphic contacts between the Napeequa and Cascade River units and to trace lithologic units from the Cascade River area to the Holden area.

Is the Ross Lake Fault Zone a Major Tectonic Boundary?

The geology of the Gabriel Peak–Elijah Ridge area (Fig. 4) is central to resolving the controversy over the tectonic significance of the Ross Lake Fault Zone. Along strike to the northwest and southeast, there is little question that faults of the Ross Lake zone form distinct breaks in metamorphic grade and lithology, separating low-grade rocks of the Methow basin on the northeast from medium- to high-grade oceanic rocks characteristic of the Napeequa unit on the southwest (Haugerud, 1985; Miller and Bowring, 1990; Miller, 1994). Kriens and Wernicke (1990) propose, however, that almost all of the supracrustal rocks in the Gabriel Peak–Elijah Ridge area belong to the same unit (Newby Group of Barksdale, 1975) of the Methow basin and that there is thus no significant tectonostratigraphic break across the Ross Lake zone in this area. They interpret the apparent abrupt gradient in metamorphic pressures on western Elijah Ridge (Whitney, 1992) to result from a cryptic, pre-90 Ma normal fault unrelated to the Ross Lake zone.

We disagree with several aspects of the Kriens and Wernicke (1990) hypothesis, noting that the following observations favor a tectonic contact within the Gabriel Peak–Elijah Ridge area:

(1) Rocks in the western part of the area belong to the oceanic Napeequa unit (abundant mafic schist, minor ultramafite, marble, metagabbro), whereas the eastern part contains probable metamorphosed equivalents of Methow basin fill (metaclastic rocks with distinctive chert-pebble conglomerate). Local interlayering of these rocks on western Elijah Ridge is compatible with imbrication of Methow and Napeequa rocks prior to metamorphism, as suggested by Kriens and Wernicke (1990), but they contend that the Napeequa rocks are minor components and that their Newby Group protoliths continue west of Elijah Ridge well into the northeastern Cascades core.

(2) High-pressure amphibolite-facies assemblages are for the most part confined to the oceanic rocks, whereas rocks of Methow basin affinity were metamorphosed at

lower pressures and temperatures (for example, Whitney, 1992).

(3) Many rocks in the area are mylonitic.

Kriens and Wernicke (1990) also infer that the high (8–9 kbar) pressures recorded by metamorphic rocks on western Elijah Ridge predate intrusion of the Black Peak batholith. The high-pressure assemblages, however, probably formed in the Late Cretaceous or Paleocene, after intrusion of the relatively shallow (emplaced at 1–3 kbar) Black Peak batholith at about 90 Ma and before or during emplacement of the much deeper (emplaced at ~6 kbar) Oval Peak batholith at about 65 Ma (Miller and others, 1993a, 1993b).

In summary, we propose for the following reasons that a major tectonic boundary lies within the Gabriel Peak–Elijah Ridge area:

(1) The Napeequa unit and probable metamorphosed equivalents of Methow basin strata (Virginian Ridge and Harts Pass or Panther Creek formations) are tectonically imbricated.

(2) The probable correlative of the Napeequa unit, the Bridge River–Hozameen terrane, is in tectonic contact with rocks of the Methow basin over a broad region (for example, McGroder, 1989, 1991).

(3) The Methow terrane is juxtaposed against rocks correlative with the Napeequa unit (Twisp Valley Schist) in the Ross Lake zone southeast (Miller and Bowring, 1990) of the Gabriel Peak–Elijah Ridge area.

We cannot, however, unequivocally demonstrate that this boundary has large displacements across it because the chert-rich facies of the Virginian Ridge Formation was derived from, and may have in part overlain, an intrabasinal high of Bridge River–Hozameen terrane (Garver, 1992). Thus, imbrication of metamorphosed Methow strata and the Napeequa unit may represent a late stage in the Cretaceous accretion of a western source terrane (for example, Monger and others, 1982), or it may record inversion of the Methow basin—that is, the Methow terrane may have a heterogeneous basement of both Spider Peak and Napeequa–Hozameen elements.

The timing of juxtaposition of supracrustal units in the Gabriel Peak–Elijah Ridge area is tightly bracketed by the probable Cenomanian or Turonian age (maximum of ~97 Ma) of the rocks inferred to represent metamorphosed Virginian Ridge Formation and by the 88–91 Ma Black Peak batholith that stitches the boundary to the southeast of the area (Misch, 1966; Kriens and Wernicke, 1990; Miller, 1994). Early Tertiary faults of the Ross Lake Fault Zone, which are evident to the northwest and southeast as discussed above, have probably been obliterated by younger intrusions, particularly the leucocratic tonalite and granodiorite of central Elijah Ridge.

SUMMARY

Our transect of the northeastern Cascades crosses several major tectonostratigraphic elements that are, from northeast to southwest:

(1) The Methow basin, which consists of Jurassic and Lower Cretaceous (perhaps as young as Cenomanian) mostly marine clastic and volcanic strata of the Methow terrane deposited at least in part on a basement of Triassic ocean ridge basalt (Ray, 1986). These rocks are overlain by the Upper Cretaceous Pasayten Group, an overlap assemblage which ties the Methow terrane to a western oceanic source terrane.

(2) Medium-grade, metamorphosed volcanic and clastic rocks, some of which probably correlate with the Upper Cretaceous Pasayten Group of the Methow basin. These rocks are separated from little metamorphosed strata of the Methow basin by the Hozameen fault and North Creek fault of the Ross Lake Fault Zone. These faults largely predate the 49 Ma Golden Horn batholith (for example, McGroder and others, 1990; Miller, 1994).

(3) The Twisp Valley Schist, Rainbow Lake Schist, and rocks on the western part of Elijah Ridge, which formed in a deep marine basin. They lie west of the Foggy Dew fault, Twisp River fault, and intrusions that obliterated the Ross Lake fault and are bounded on the southwest by the Gabriel Peak tectonic belt. We correlate these rocks with the Napeequa unit of the Cascades core and the Mississippian to Jurassic Bridge River–Hozameen terrane.

(4) Orthogneiss of the Skagit Gneiss Complex, which contains rafts of supracrustal rock that probably correlate with the Holden assemblage and Cascade River unit. These rocks lie southwest of the Gabriel Peak tectonic belt, which may represent along at least part of its length the present boundary between the Napeequa unit and Cascade River unit.

(5) Thoroughly intruded paragneisses of the Skagit Gneiss Complex, which are possibly correlative with the Cascade River unit. These rocks were originally mapped as the Swakane Gneiss.

(6) Supracrustal rocks of the Holden assemblage, which also correlate with the Cascade River unit. The difference between this assemblage and the Skagit Gneiss Complex is probably the amount of contained intrusive material. Correlation of the Holden assemblage with the Cascade River unit weakens the depositional link between the Cascade River unit and Napeequa unit proposed by Tabor and others (1989) and suggests that the Chelan Mountains terrane of Tabor and others (1987a) includes two separate terranes.

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Tectonic Evolution of the Cascades Crystalline Core in the Cascade River Area, Washington

E. H. Brown, J. A. Cary, B. E. Dougan, J. D. Dragovich[1], S. M. Fluke, and D. P. McShane
Department of Geology
Western Washington University
Bellingham, WA 98225
[1]Washington Division of Geology and Earth Resources
P.O. Box 47007, Olympia, WA 98504-7007

ABSTRACT

This study of metamorphism and plutonism in the Cascade River area of the Cascades crystalline core addresses questions concerning the causes of the profound mid- to Late Cretaceous orogeny that affected this belt, as well as the entire 1,800-kilometer-long Coast Plutonic Complex of which the study area is a part.

The two chief country rock units are: (1) a Triassic (220 Ma) arc sequence that includes a plutonic basement, the Marblemount Meta-Quartz Diorite, and a cogenetic supracrustal volcanic and sedimentary cover named the Cascade River unit, and (2) a structurally chaotic package of oceanic rocks of unknown age, the Napeequa unit. These rock units were mutually juxtaposed along low-angle faults prior to high-grade metamorphism and plutonism.

Metamorphism in the area is diachronous and characterized by sharp gradients. The area can be divided into two blocks of different metamorphic character, the boundary being approximately marked by the post-metamorphic Entiat fault. Within the northeastern block, metamorphic grade rises abruptly *across strike* to the northeast, passing into the Skagit complex; grade ranges from chlorite zone semi-schists to 9-kilobar staurolite-kyanite schists within 8 to 10 kilometers. Metamorphic loading was developed in this block between emplacement times of the shallow-level Eldorado pluton (88 to 90 Ma) and the deep-level Marble Creek pluton (76 Ma). Southwest of the Entiat fault, metamorphic grade increases *along strike* to the southeast, ranging from 3–4-kilobar chlorite zone schists to 9-kilobar rocks of the amphibolite facies within a map distance of about 30 kilometers. This metamorphic gradient was established at or before 92 to 96 Ma, the age range of deep-level, syntectonic plutons emplaced in this zone. Metamorphic fabrics across the study area are dominated by steep, northwest-striking foliation and shallow, strike-parallel stretching and mineral lineations. Dextral-shear-sense features are common. Much less common is fabric characterized by weak down-dip mineral lineation or by oblate deformed clasts not accompanied by mineral lineation.

The observed metamorphic patterns are not readily interpreted in terms of a thrust loading mechanism. We suggest that the steep gradients, different directions of the gradients, and different ages of the gradients are best explained as the product of localized loading of country rock by plutons emplaced at higher levels in the crust. The predominant metamorphic fabric appears to be the result of strike-slip dextral shear not directly related to the metamorphic loading mechanism, but perhaps focused by the weakening effect of plutonism.

INTRODUCTION

The Cascades crystalline core of Washington is a metamorphic-plutonic complex representing the southeast end of the 1,800-km-long Coast Plutonic Complex (CPC) that extends through British Columbia into Alaska. The CPC developed across the boundary between the Insular superterrane (Wrangellia terrane in the region of interest to this study; Fig. 1) and North America (the Intermontane superterrane in Washington). Thus, terrane accretion is a tectonically important event in the history of the complex. In addition, subduction has been a significant process in this region because a great amount of oceanic plate was consumed under the western margin of North America during formation of the CPC. Current hypotheses to explain the metamorphism, deformation, and plutonism in the CPC vary widely, from a purely contractional/collisional model in which these orogenic processes are attributed to accretion-related tectonic thickening (Monger and others, 1982; Whitney and McGroder, 1989; McGroder, 1991) to an arc model where emplacement of subduction-derived magmas is inferred to be the fundamental cause of the orogenesis (Brown and Walker, 1993).

Compared to many other parts of the CPC, the Cascades crystalline core has relatively more country-rock schist and less plutonic rock. It therefore provides an opportunity to trace a tectonic record from protolith formation through the peak of orogeny. In this paper, we report on stratigraphy, structure, and metamorphism in the Cascade River area of the crystalline core (Fig. 2). Our primary goal is to discern the cause of orogeny: tectonic thickening, arc plutonism, or other processes. The rock record in this region is complex and in some respects remains indecipherable. Our findings indicate that arc and oceanic protolith materials, in part of Triassic age, were assembled in their present association

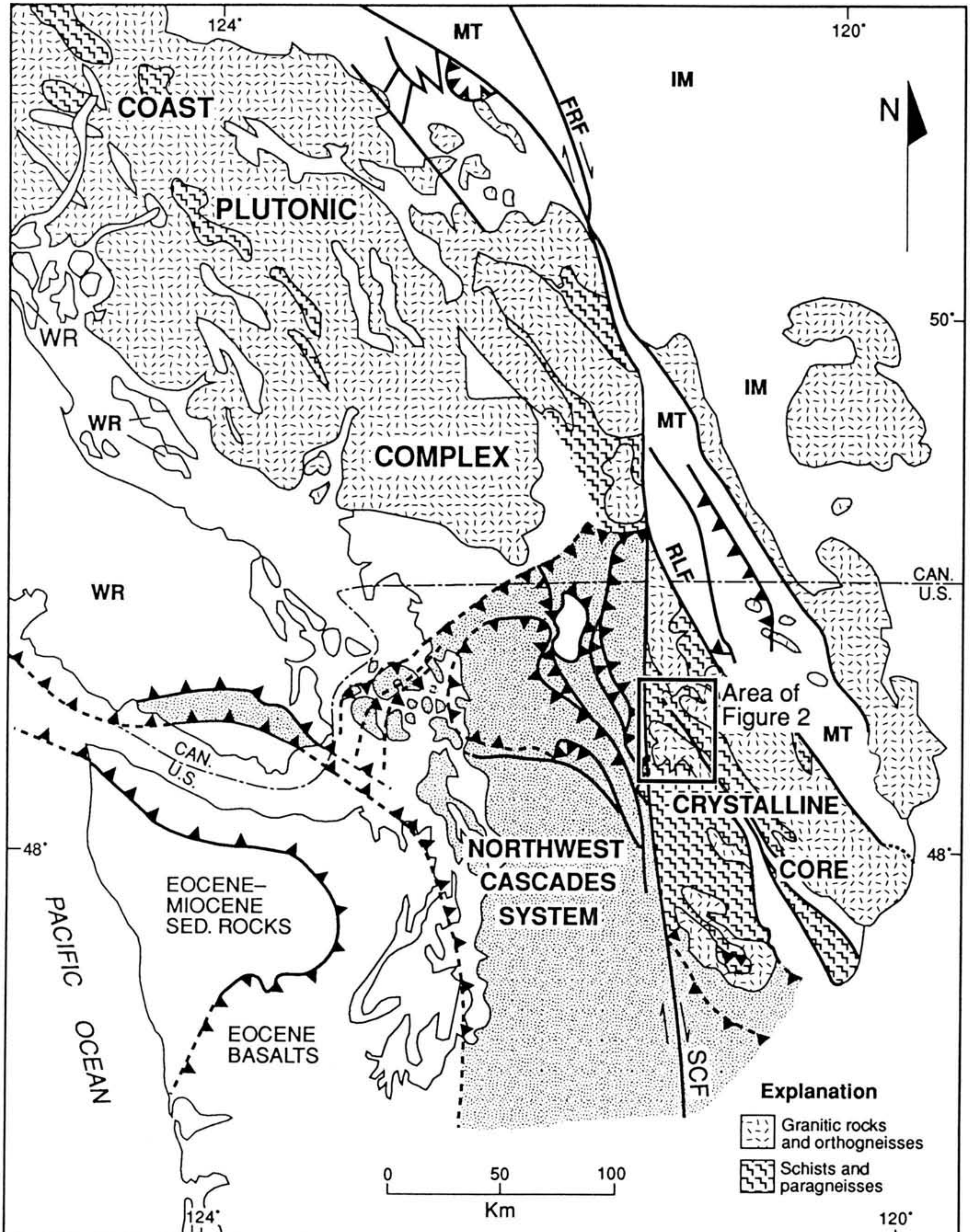


Figure 1. Regional geologic setting of the Cascade River area. FRF, Fraser River fault; IM, Intermontane superterrane; MT, Methow–Tyaughton terrane; RLF, Ross Lake Fault Zone; SCF, Straight Creek fault; WR, Wrangellia terrane.

prior to mid-Cretaceous plutonism and metamorphism. Isograd patterns and baric gradients are not compatible with a thrust-loading orogenic mechanism, and we suggest that the metamorphism is directly related to plutonism.

Regional Geologic Setting

The Cascade crystalline core (CC) is dominated by syntectonic plutons and high-grade schists and gneisses, all formed in mid-Cretaceous to Eocene time. The CC is offset from the CPC along the Tertiary Straight Creek–Fraser River fault system, which truncates structural trends in the CC and forms its western flank (Fig. 1). The northeast side of the CC is faulted (Ross Lake Fault Zone) against, and may be transitional to, low-grade rocks of the Methow area (Tabor and others, 1989; Kriens and Wernicke, 1990). A number of terranes have been distinguished within the CC on the basis of their age and lithology (Tabor and others, 1987a); however, criteria proving disparate origins of these map units in the sense of the terrane concept (for example, Coney and others, 1980) are not yet well established. Useful divisions for this report (following Tabor and others, 1989) are the Chelan Mountains terrane and the Nason terrane (Chiwaukum Schist in the study area; Fig. 2). Grouped within the Chelan Mountains terrane are the Cascade River unit and Napeequa unit.

The Cascade River drainage, which is the center of the study area (Fig. 2), lies in an area of low-grade metamorphic rocks between two high-grade culminations within the CC. To the northeast is the Skagit Gneiss complex, consisting dominantly of migmatite and orthogneiss with high-pressure metamorphic signature (8–9 kb) and mainly Late Cretaceous through early Tertiary plutons and fabric (Misch, 1966; Mattinson, 1972; Babcock and Misch, 1989; Tabor and others, 1989; Whitney and McGroder, 1989; Miller and Bowring, 1990; Haugerud and others, 1991; Whitney, 1992). To the southwest of the Cascade River, a high-grade belt is made up of schists, gneisses, and migmatites recording pressures as high as 9 kb and intruded by abundant synmetamorphic mid-Cretaceous plutons (Magloughlin, 1986; Tabor and others, 1987b, 1988, 1989; Evans and Berti, 1986; Ford and others, 1988; Walker and Brown, 1991; Brown and Walker, 1993; Fluke, 1992; Dougan, 1993).

Because the Cascade River study area straddles a low-grade zone between the two high-grade belts of the CC, the transition from low to high grade can be observed in this region, allowing documentation of relatively unaltered protolith materials and structural and metamorphic features progressive into the orogenic culminations. Data from this area form a basis for evaluation of the cause of the high-grade metamorphism.

General Geology of the Cascade River Area

Early work in the region by Bryant (1955), Tabor (1961) and Misch (1966, 1979) led to the recognition of two major country-rock units: the Cascade River Schist, which includes all the supracrustal rocks, and the Marblemount Meta-Quartz Diorite (MMQD), a premetamorphic plutonic unit. Clasts of quartz diorite in conglomerate beds of the Cascade River Schist led Tabor (1961) and Misch (1966) to conclude that the MMQD is unconformably overlain by the Cascade River Schist. A Late Triassic U-Pb zircon age (220 Ma) was determined for the MMQD by Mattinson (1972).

More recently, Tabor and others (1988) recognized two separable lithologic suites in the Cascade River Schist as originally defined: a metamorphosed, dominantly clastic assemblage, in part conglomeratic, which they named the Cascade River unit, and a structurally dismembered package consisting mostly of quartz-rich mica schist and amphibolite with pods and lenses of ultramafic rock, metagabbro, and marble, named the Napeequa unit. Tabor and others (1988) regard the Cascade River unit as unconformably overlying both the MMQD and the Napeequa unit, and they assign all these units to the Chelan Mountains terrane.

A focus of our study has been examination of features indicative of the primary relations among these units. Our interpretation, based on observations given below, is that the MMQD and Cascade River unit are cogenetic parts of the same Triassic magmatic arc, and that primary relations of this package with the Napeequa unit remain uncertain.

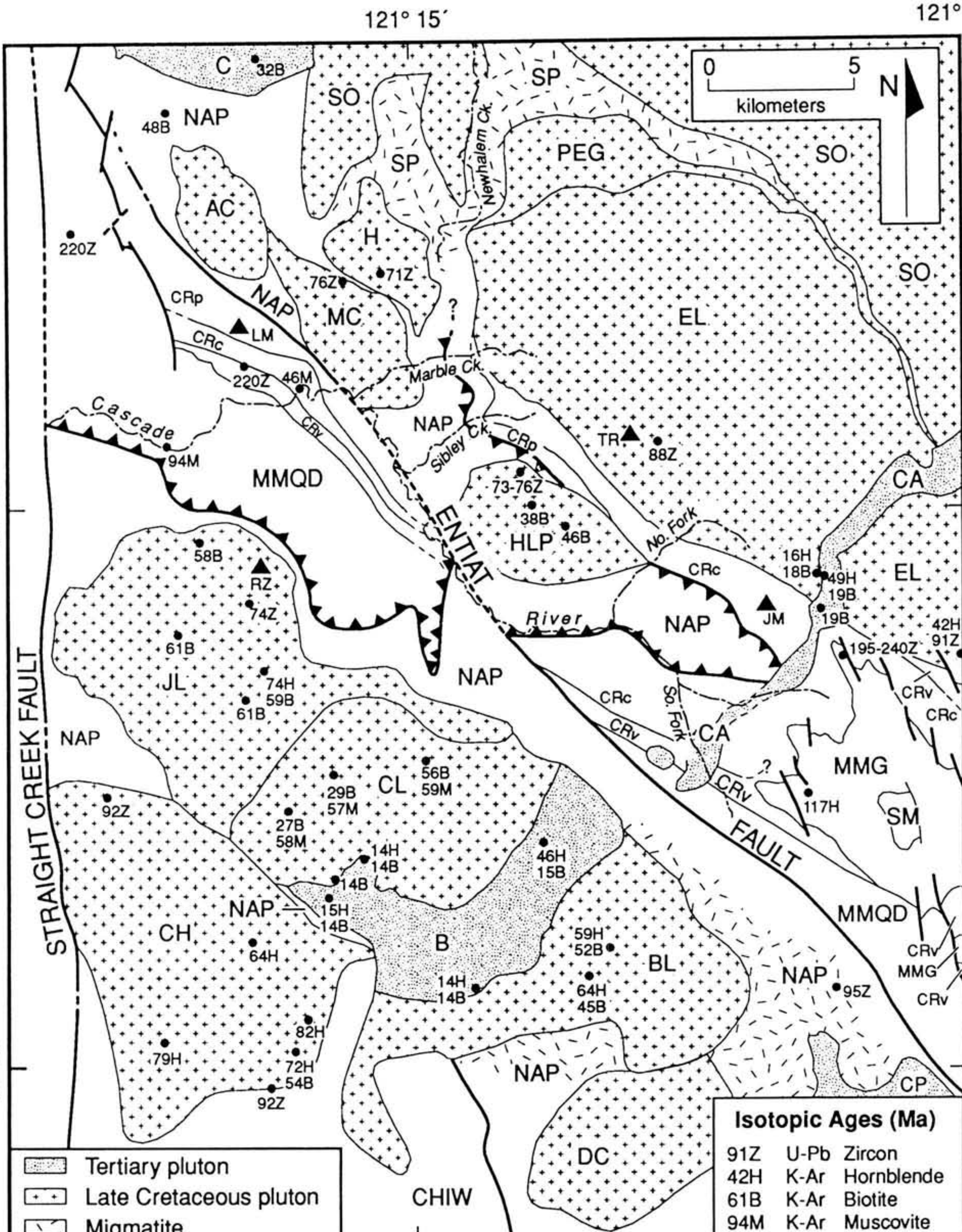
The country-rock units extend well outside the Cascade River area. The MMQD is part of a belt of Triassic plutons, named the Marblemount belt by Misch (1966), that extends for about 90 km, mostly to the southeast of the study area. The Cascade River unit is also traceable for many tens of kilometers on strike to the southeast. The Napeequa unit may be the dominant country-rock unit in the Skagit Gneiss. This unit is also found on the northeast flank of the Nason metamorphic belt in the southern part of the study area and farther southeast.

A number of plutons intrude country-rock units in the study area (Fig. 2). On the northeast side of the Cascade River drainage, the Eldorado (88–90 Ma), Marble Creek (75 Ma), and Haystack (71 Ma) plutons are foliated and have been interpreted as syntectonic, whereas the Hidden Lake Peaks (74 Ma) pluton is unfoliated and is thus suggested to be post-tectonic (Misch, 1966; Dragovich and others, 1989; Haugerud and others, 1991; McShane, 1992). On the southwest side of the Cascade River drainage, the undated Downey Creek, Bench Lake, and Cyclone Lake plutons are foliated, whereas the Jordon Lake pluton (74 Ma) is not foliated and crosscuts country-rock fabric and has been regarded as post-tectonic (Dragovich and others, 1989). In a broad region around the Bench Lake and Downey Creek plutons, the country-rock Napeequa unit is host to a dike/sill complex of tonalitic material comprising an injection zone migmatite (termed the "banded tonalite gneiss" by Tabor and others, 1988). A deformed sill of this material yields a zircon age of 95 Ma (N. W. Walker, Univ. of Texas, Austin, oral commun., 1991).

Stratigraphy and Lithology

Our study began with an idea derived from map patterns reported by Tabor (1961) and Misch (1966, 1979) that a useful stratigraphy might be recognized within the Cascade River unit. Detailed work in low-grade zones across the south face of Lookout Mountain (Cary, 1990), on northeast-facing slopes opposite the mouth of Marble Creek (Dragovich, 1989), and near the South Fork Cascade River has yielded a stratigraphy that is mappable through at least parts of the area (Fig. 2). Lithologic units in these areas are nearly vertical, strike northwest, and are younger to the northeast.

Lowest in the section is the Marblemount Meta-Quartz Diorite (MMQD), described by Bryant (1955), Tabor



(1961), and Misch (1966). Following Tabor and others (1988), the Magic Mountain Gneiss (MMG) at the southeast end of the study area is correlated with the MMQD, although these two units are lithologically somewhat different. The MMQD is derived from mostly massive, uniform quartz diorite. In contrast, the MMG is a metamorphosed layered complex of felsic tonalite and greenschist. The MMG has been alternatively interpreted as a thrust slab emplaced over the Cascade River unit (Tabor, 1961) or a dike/sill complex intrusive into the Cascade River unit (Dougan, 1993).

The contact of the MMQD with the Cascade River unit is metamorphosed and deformed, and although in places its primary character cannot be discerned, at localities on Lookout Mountain the MMQD appears to be gradational into, and to have mutually intrusive relations with, hypabyssal and volcanic rocks of the Cascade River unit (Fugro NW, 1979; Cary, 1990).

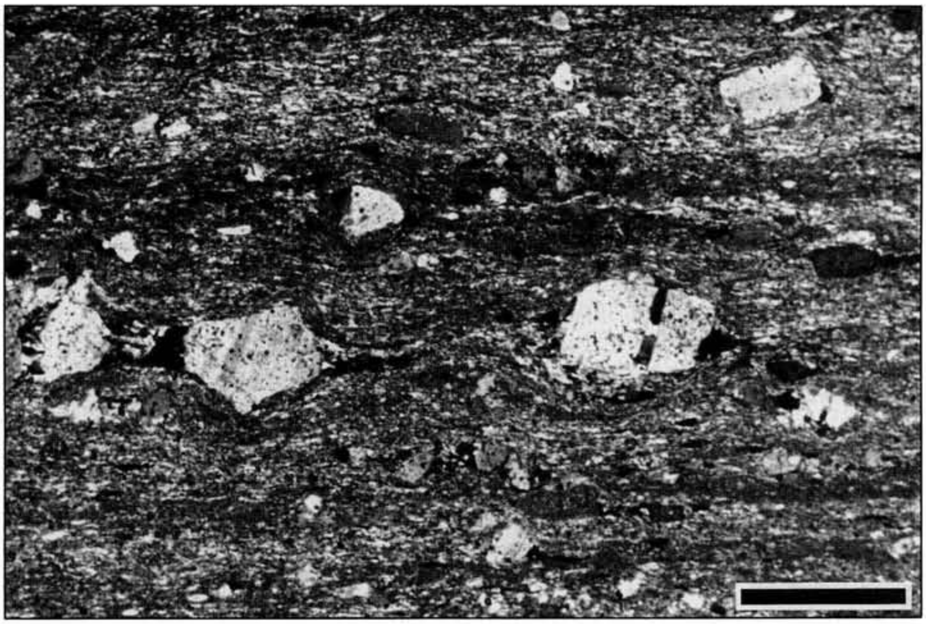


Figure 3. Metadacite with relict phenocrysts from the volcanic petrofacies of the Cascade River unit on the south side of Lookout Mountain. Scale bar is 2 mm.

The volcanic part of the Cascade River unit is composed of metamorphosed predominantly pyroclastic deposits and lesser dikes and flows. Compositions range from basalt to rhyolite, with dacite and andesite most abundant. Relict igneous features are locally well preserved in low-grade areas; these include crystal and lithic lapilli and porphyritic-aphanitic textures (Fig. 3). This unit is 500–600 m thick. Chemical compositions are comparable to those of the MMQD; both units are of calc-alkaline character (Cary, 1990).

The volcanic unit gives way gradually upward to a poorly sorted coarse clastic section containing volcanic and plutonic fragments of the underlying units and interbeds of pyroclastic rock. Clasts are as long as 2 m, but most are in the range of 4–20 cm. Much of the clastic rock is matrix supported, and it may represent, in part, volcanic debris flows. The clastics fine to the northeast (upward). On Lookout Mountain, the coarse clastic unit is about 600 m thick. Along strike to the southeast, the clastic unit thickens to more than 1,000 m near the South Fork Cascade River.

The youngest unit of the section is a pelitic and marly metasedimentary rock with interlayers of tuffaceous volcanic rock. This unit is also locally quartzose (on Lookout Mountain and in the upper Sibley Creek drainage), probably representing a chert protolith. One extensive marble bed has been traced discontinuously for about 5 km from the upper Sibley Creek area (Dragovich, 1989) to the north across Marble Creek (McShane, 1992). In areas of high-grade metamorphism, layers of amphibolite in this unit probably represent tuffaceous interlayers seen at lower grade.

Although the contacts between these units are metamorphosed and in many places sheared, enough excellent exposures of relatively undeformed rock are available to confirm the mutually gradational character of adjacent units and thus the coherence of the stratigraphic section. Support for this conclusion is given by a U-Pb zircon age for dacite in the volcanic unit on Lookout Mountain determined by John Stacy of the U.S. Geological Survey (data reported in

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Figure 2. *(facing page)* Geologic units, localities, and isotopic ages in the Cascade River area. AC, Alma Creek pluton; B, Buckindy pluton; BL, Bench Lake pluton; C, Chilliwack pluton; CA, Cascade Pass pluton; CH, Chaval pluton; CHIW, Chiwaukum Schist; CL, Cyclone Lake pluton; CP, Cloudy Pass pluton; CRc, Cascade River unit, coarse-clastic petrofacies; CRp, Cascade River unit, pelitic petrofacies; CRv, Cascade River unit, volcanic petrofacies; DC, Downey Creek pluton; EL, Eldorado pluton; H, Haystack pluton; HLP, Hidden Lake Peaks pluton; JL, Jordan Lake pluton; JM, Johannesburg Mountain; LM, Lookout Mountain; MC, Marble Creek pluton; MMG, Magic Mountain Gneiss; MMQD, Marblemount Meta-Quartz Diorite; NAP, Napeequa unit; PEG, pegmatite associated with the Eldorado pluton; RZ, Razorback Mountain; SM, Spider Mountain schist; SP, paragneiss of the Skagit complex; SO, orthogneiss of the Skagit complex; TR, The Triad. Geologic map relations from Bryant (1955), Tabor (1961), Misch (1979), Fugro NW (1979), Ford and others (1988), Tabor and others (1988, 1989), and this study. Isotopic ages from Engels and others (1976), Mattinson (1972), Tabor and others (1988), Haugerud and others (1991), Walker and Brown (1991), John Stacy (USGS, written commun., 1988), and N. W. Walker (Univ. of Texas, Austin, oral commun., 1991).

Cary, 1990). The age is 220 ± 3 Ma, the same (within uncertainty limits) as that previously obtained by Mattinson (1972) for the underlying MMQD.

A magmatic arc setting for the Cascade River unit is suggested by the intermediate composition of the igneous rocks and the relative abundance of pyroclastic deposits. Thus, the MMQD represents subarc plutonic material, and the Cascade River unit a volcanic and sedimentary edifice forming the upper part of the section.

The Napeequa unit of Tabor and others (1988, 1989) is made up of an assemblage of lithologies of probable ocean-floor origin. Most abundant are quartzose biotite schist and centimeter-scale interlayered quartzite and mica schist. The latter lithology is apparently metamorphosed ribbon chert. Next in abundance is amphibolite of basaltic composition. Much of the amphibolite is thoroughly recrystallized; however, some contains a relict gabbroic plutonic texture. Relatively unaltered gabbros are also exposed. Ultramafic rock occurs in lenses and pods from centimeters to tens of meters or more in length in which most primary textures and mineralogy have been obliterated. Marbles occur as discontinuous layers and pods a few meters thick. Fault zones must be plentiful, juxtaposing unrelated lithologies (especially ultramafic bodies with other rocks), but exposure generally does not permit observation.

LARGE-SCALE STRUCTURE

Structural Relations of the Cascade River and Napeequa Units

Some progress toward delimiting large-scale structure of country rock in the area is possible on the basis of map relations of stratigraphy in the MMQD–Cascade River unit and the relation of this complex to the Napeequa unit. Nonetheless, significant uncertainties remain for future generations of hardy bushwhacking geologists to unravel.

In most of the study area, the Cascade River unit lies in two subparallel belts. On the south side of Lookout Mountain, the Cascade River unit with its basement of MMQD forms a nearly vertical, northwest-striking homoclinal sequence. If we accept the stratigraphic order described above, the sequence here becomes younger to the northeast. In the Sibley Creek area, the other belt of the Cascade River unit dips southwest and bears graded bedding indicative of stratigraphically younger rocks lying to the southwest (Dragovich, 1989; McShane, 1992). The Napeequa unit lies between the two belts of the Cascade River unit along the Cascade River valley. In the vicinity of Johannesburg Mountain, the Napeequa unit ends, and the two belts of the Cascade River unit join. A few kilometers farther southeast, a broad, subhorizontal sheet of the MMG extends over and across nearly the entire breadth of the merged belts of the Cascade River unit. The geometry of structures responsible for this map pattern is not easily discerned, partly because outcrop-scale structures are related to a later metamorphic overprint, and partly because mapping is incomplete in this rugged and remote area.

The Napeequa unit in the structural complex described above seemingly must be thrust against (over, under, or into) the Cascade River unit. We think the Napeequa unit cannot underlie the Cascade River unit because the contact between the units is well up in the section of the Cascade River unit and the Cascade River unit is progressively younger toward the Napeequa unit. The contact between the Napeequa and Cascade River units in this structure is not discernible in the field. Not only is the contact poorly exposed, but it is also difficult to determine to which unit rocks belong—marbles, amphibolites, metacherts, and metapelites occur in both the Napeequa and the uppermost part of the Cascade River unit. Our interpretation, based on outcrop distribution across areas of high topographic relief in the Sibley Creek and Johannesburg Mountain areas, is that the Napeequa overlies the Cascade River unit. R. W. Tabor (USGS, written commun., 1991) forms a different conclusion from the map patterns and suggests that the Cascade River unit overlies the Napeequa unit.

We infer that the Napeequa/Cascade River unit contact is a thrust, not from direct observation, but based on (1) our interpretation that the Napeequa overlies the Cascade River unit, and (2) the much greater structural dismemberment of the Napeequa unit and the presence in the Napeequa of ultramafic lenses that we think are tectonically emplaced oceanic mantle fragments. If the Napeequa is a mélange of ocean-floor and upper mantle materials, then the possibility that such an assemblage was *deposited* on top of the coherent arc stratigraphy of the Cascade River unit seems remote.

The Napeequa unit and MMQD are in contact along an observable thrust in the west-central part of the map area near Razorback Mountain (Fig. 2). This structure, mapped by Bryant (1955) and Tabor and others (1988), is well exposed on ridges and is characterized by a mylonitic fabric. The fault dips under the MMQD at angles of 30–45 degrees. Our attempts at kinematic analysis of the mylonite have not yielded consistent results. Lineations vary from shallow to steep, even at the same locality. In the Razorback Mountain area, relatively brittle fabric in the thrust zone shows left-lateral strike-slip motion and extends into the Jordan Lake pluton (74 Ma) (Fluke, 1992). In this area, dikes of the pluton crosscut ductile fabric in the Napeequa, which is parallel to the MMQD–Napeequa shear zone. Metamorphic isograds related to the regional metamorphism (described in a later section) cross the contact. These relations suggest a prolonged and complex movement history for this contact.

The geometry of the large-scale, premetamorphic structure in the Cascade River unit is not yet discernible from present mapping. Several alternative models are discussed in the following paragraphs:

(A) The Cascade River unit is a near-vertical, undisrupted homoclinal sequence, younger northeastward across its entire breadth, including the northeast belt in the Sibley Creek area. The near-horizontal Magic Mountain slab of plutonic and hypabyssal rock is a dike complex rotated from an originally vertical position. Two problems with this model are: (1) the graded bedding in the northeast belt that indicates younger rocks to the southwest (described above), a matter that needs to be addressed by further study in the northeast belt in the Sibley and Marble Creek areas; and (2) the structural position of the Napeequa unit apparently deeply embedded within the Cascade River unit section, as indicated by the trace of the Napeequa–Cascade River unit contact across steep topography.

(B) In a variant of model A, the Cascade River unit is a homoclinal thrust sequence that has the Napeequa unit imbricated as a thick, central thrust packet pinching out to the southeast. In this model, the Magic Mountain slab is a horizontal thrust emplaced over the imbricated vertical sequence, as suggested by Tabor (1961). However, we have not found outcrop evidence for this thrust. At the contact of the Cascade River unit with the MMG, the two units are interlayered on a scale of 0.2–2 m and mylonite fabric is lacking; the contact appears to be intrusive rather than tectonic (Dougan, 1993).

(C) The Cascade River unit is overthrust by the Napeequa, and both units are folded into an upright syncline. This model accommodates evidence for reversal of the stratigraphic sequence between the northeastern and southwestern belts of Cascade River unit and the map pattern of the Napeequa–Cascade River contact. Because the Magic Mountain slab is a horizontal sheet lying across this structure, it would have to represent a later thrust event.

(D) The Cascade River unit together with the MMQD and MMG are folded into a recumbent syncline. The southwest homoclinal sequence represents the steeply dipping hinge area, and the Magic Mountain slab is the inverted upper limb. The Napeequa unit is thrust over the top. Problems with this model are that it does not account for evidence that the Napeequa unit is embedded in the Cascade River unit, and the trace of contacts across topography is not consistent with a smooth curvature around a macroscopic fold hinge from the southwest homoclinal sequence to the Magic Mountain slab

These models need evaluation by further work, especially in the high country around the headwaters of the Cascade River and the outcrop area of the MMG, before a definitive explanation of the regional structure is possible.

Entiat Fault

The Entiat fault (LeConte fault of Tabor, 1961) is a near-vertical, northwest-trending, throughgoing structure. It offsets country-rock map units, it bounds and thus may cut the southwest side of the Marble Creek pluton, and it offsets metamorphic isograds (Figs. 4–7). Offset on the Entiat fault is partly defined by displacement of stratigraphy in the Cascade River unit. If the offset is purely strike-slip, restoration of stratigraphy requires about 5 km of dextral motion. Comparison of metamorphic facies across the fault, with or without 5 km of dextral motion, suggests a scissors type of dip-slip motion: in the northwest part of the map area, high-grade rocks on the northeast side of the fault are juxtaposed against low-grade rocks to the southwest; along strike to the southeast, near the limit of the map area, high-grade rocks lie on the southwest side of the fault and lower grade rocks are on the northeast side. An accurate assessment of dip-slip from metamorphic discontinuity is difficult owing to inexact calibration of strong pressure–temperature gradients observed on both sides of the fault (described in the next section), but displacement of as much as 20 km seems likely.

Metamorphic Facies

Strong metamorphic gradients in the study area represent a range from low greenschist facies to middle amphibolite facies. A low-grade metamorphic zone lies more or less along the Cascade River valley (Figs. 4–7). From this zone, grade increases sharply northeast into the Skagit belt, in places with discontinuity across the Entiat fault. On the southwest side of the Entiat fault, grade rises to the southeast, parallel to regional structural trends. Metamorphic zones overprint large-scale structures in the area, but not the Entiat fault, which postdates metamorphism.

In the lowest-grade zone, metamorphism is characterized by the mineral assemblage albite + epidote + chlorite + actinolite + quartz. Absence of biotite in this zone (Fig. 4) is interpreted to reflect sub-biotite zone conditions of metamorphism; however, rock composition is possibly also a factor.

Away from the lowest-grade zone, we have mapped isograds on the basis of first appearance of biotite, garnet, oligoclase, hornblende, and staurolite, and the last occurrence of albite (Figs. 4–7). The location of zones defined by albite vs. oligoclase and actinolite vs. hornblende is based on microprobe analyses. A sizable area of overlap in the occurrences of albite and oligoclase, possibly representing the peristerite solvus, occurs in the southeast end of the map area. In the staurolite zone, kyanite was found at three localities, and andalusite pseudomorphed by kyanite was recorded at one locality (described below). To summarize the zonal sequence: biotite appears at lower grade than other index minerals; garnet, oligoclase and hornblende all appear at about the same grade above the biotite zone; and staurolite and kyanite come in at a still higher grade. This sequence is typically Barrovian and suggests a medium pressure/temperature field gradient. (See, for example, Miyashiro, 1973.)

Retrograde assemblages are common throughout the higher grade zones, notably along shear zones. A late-stage overprint also occurs near the Miocene Cascade Pass pluton (Fig. 2) in the form of andalusite and cordierite in mica schists, as well as patches of actinolite in more mafic schists; these effects are observed as much as several hun-

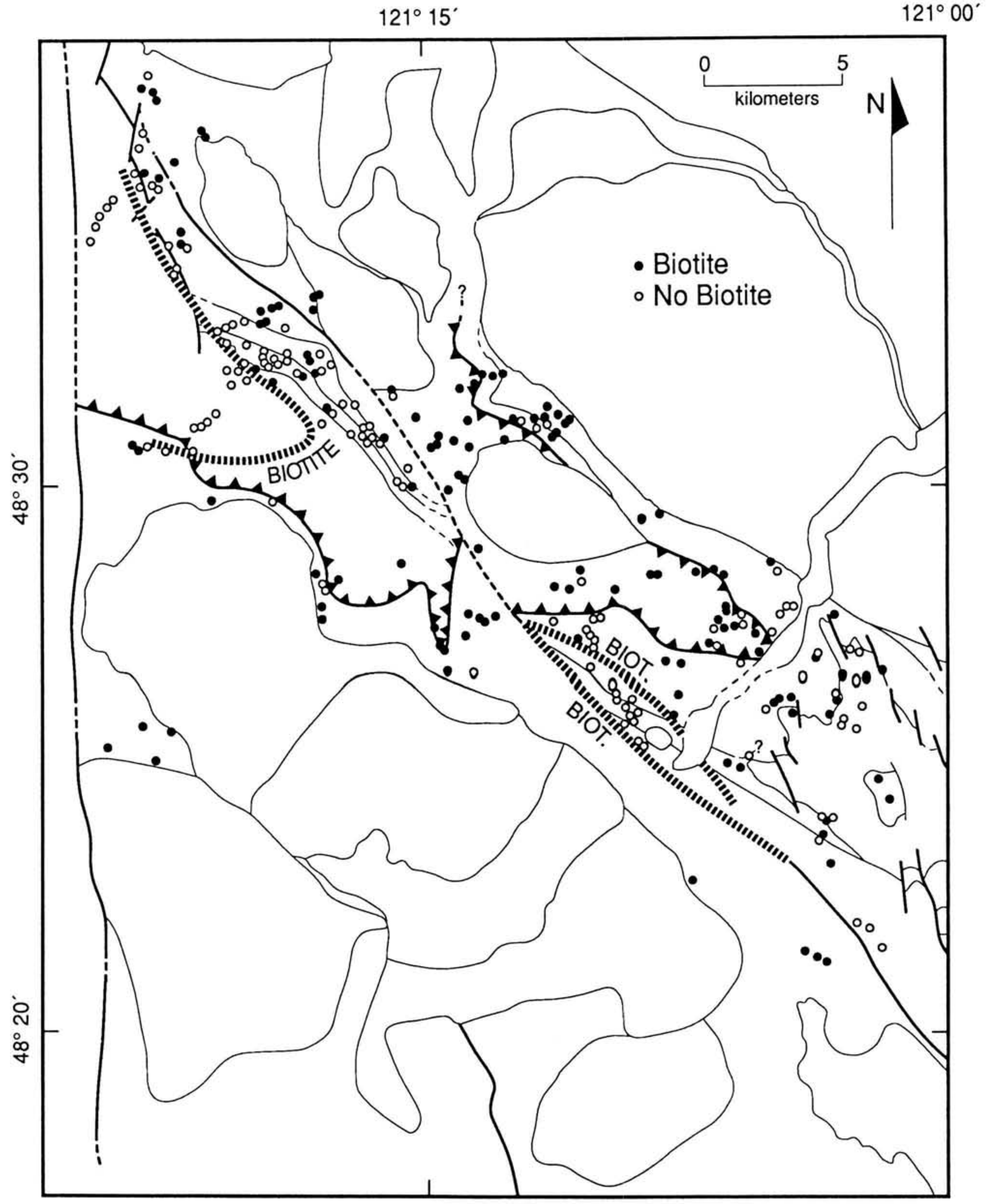

Figure 4. The distribution of biotite and the biotite isograd in the Cascade River area.

dred meters from exposed Cascade Pass plutonic rock, and they partially obscure regional metamorphic isograds through this region. (See also Tabor, 1961, 1963.)

Thermobarometry in the Country Rock

Absolute values of pressure (P) and temperature (T) were estimated using a host of mineral equilibria calibrated in various independent ways. Comparisons based on these

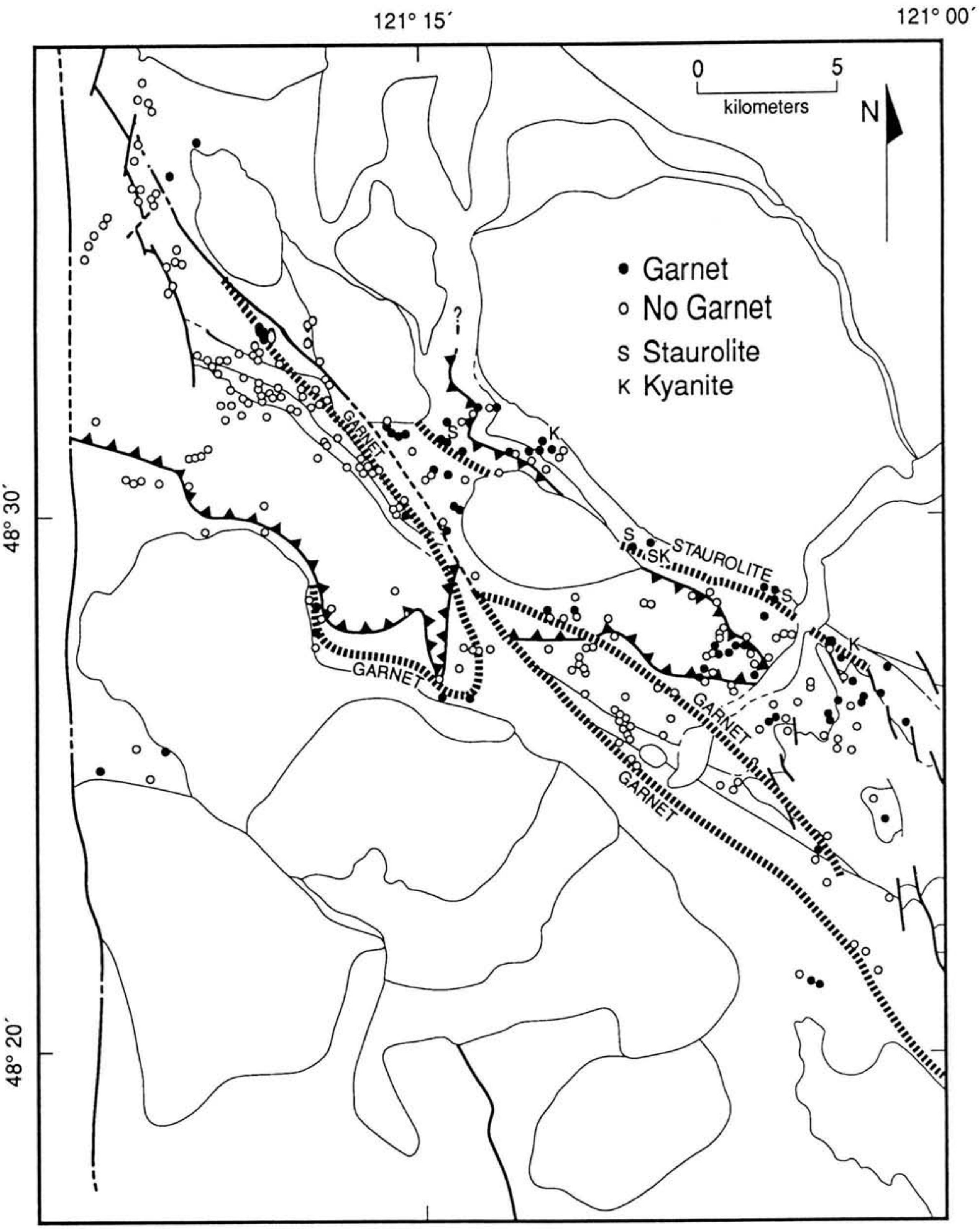

Figure 5. The garnet and staurolite isograds in the Cascade River area.

methods of P-T estimation must allow uncertainties of at least ±1.0 kb and 50°C. Within this range of uncertainty, pressures determined by independent mineralogic equilibria are generally in agreement; exceptions are discussed below. The thermobarometric results are also in general agreement with P-T values estimated for Barrovian metamorphism (for example, Yardley, 1989). The estimated P-T values are shown on Figure 8 together with our interpreta-

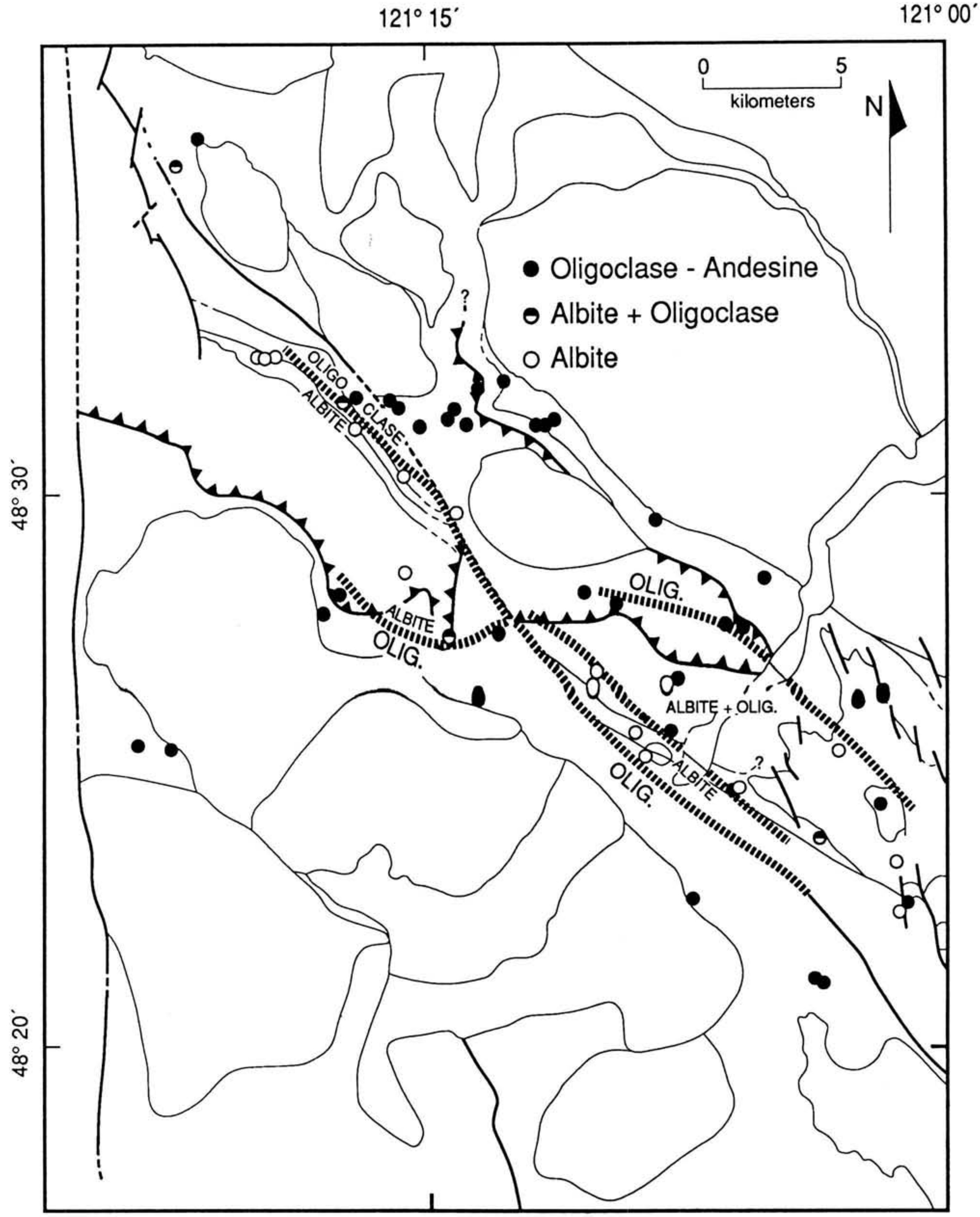

Figure 6. The albite and oligoclase isograds in the Cascade River area. Mineral identification is based on microprobe analyses.

tion of the distribution of isobars representing peak metamorphic conditions. Constraints for the inferred baric patterns are fairly sparse, but except for post-metamorphic offset on the Entiat fault, the patterns would be expected to be smooth and continuous, giving some credibility to the extrapolations between data points. However, this analysis of the distribution of metamorphic pressures would benefit from more data.

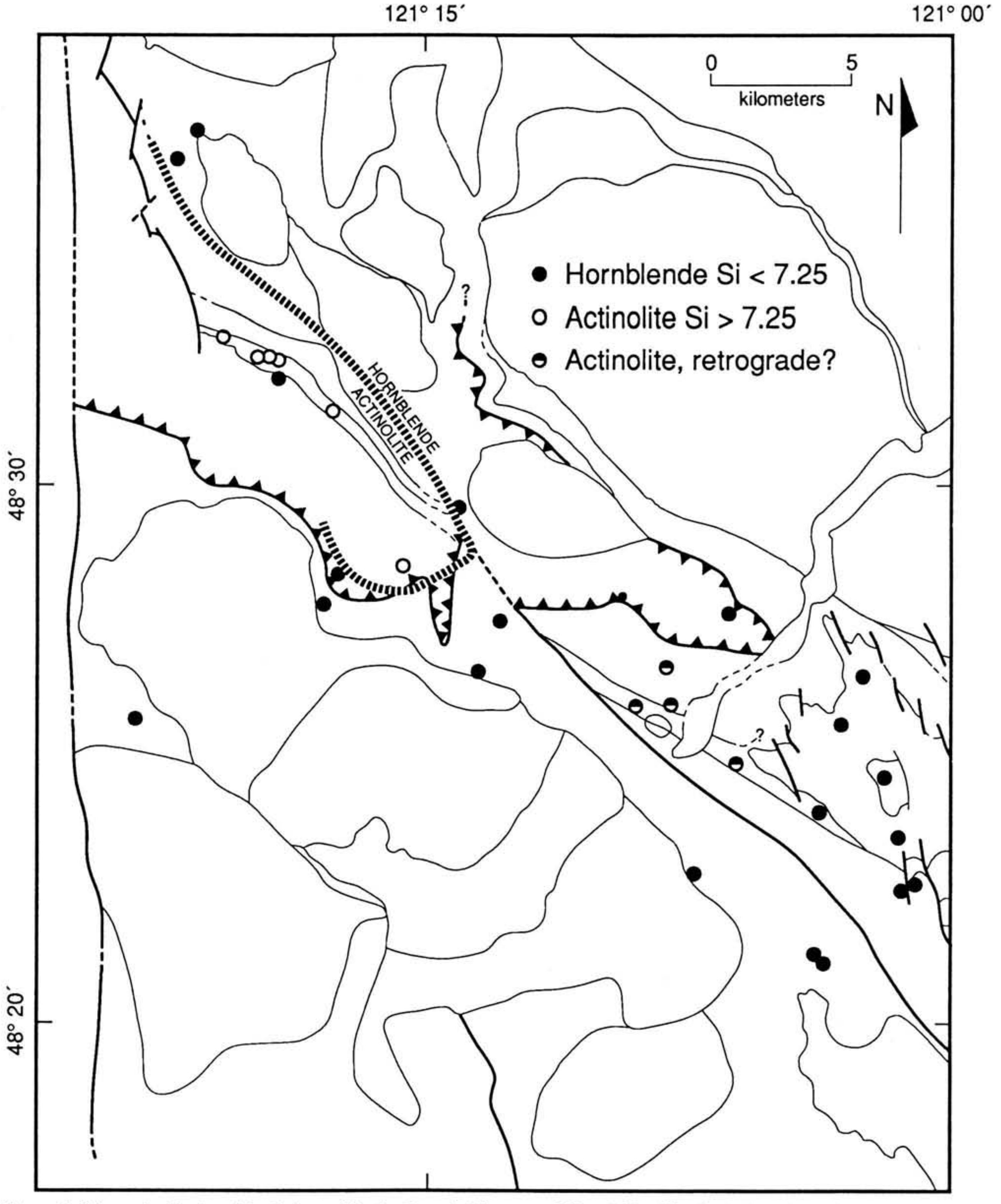

Figure 7. The actinolite–hornblende isograd in the Cascade River area. Mineral identification is based on microprobe analyses.

In the rocks having the lowest grade, pressure is in the 3–4 kb range as determined from a very low Na-amphibole component in actinolite in a buffering assemblage of quartz + albite + chlorite + epidote + magnetite (Brown, 1977). Temperature in the lowest grade rocks is judged to be less than 400°C from evaluation of the P-T conditions of the biotite, garnet, albite–oligoclase, and actinolite–hornblende isograds. (See, for example, Brown, 1978; Maruyama and others, 1983.)

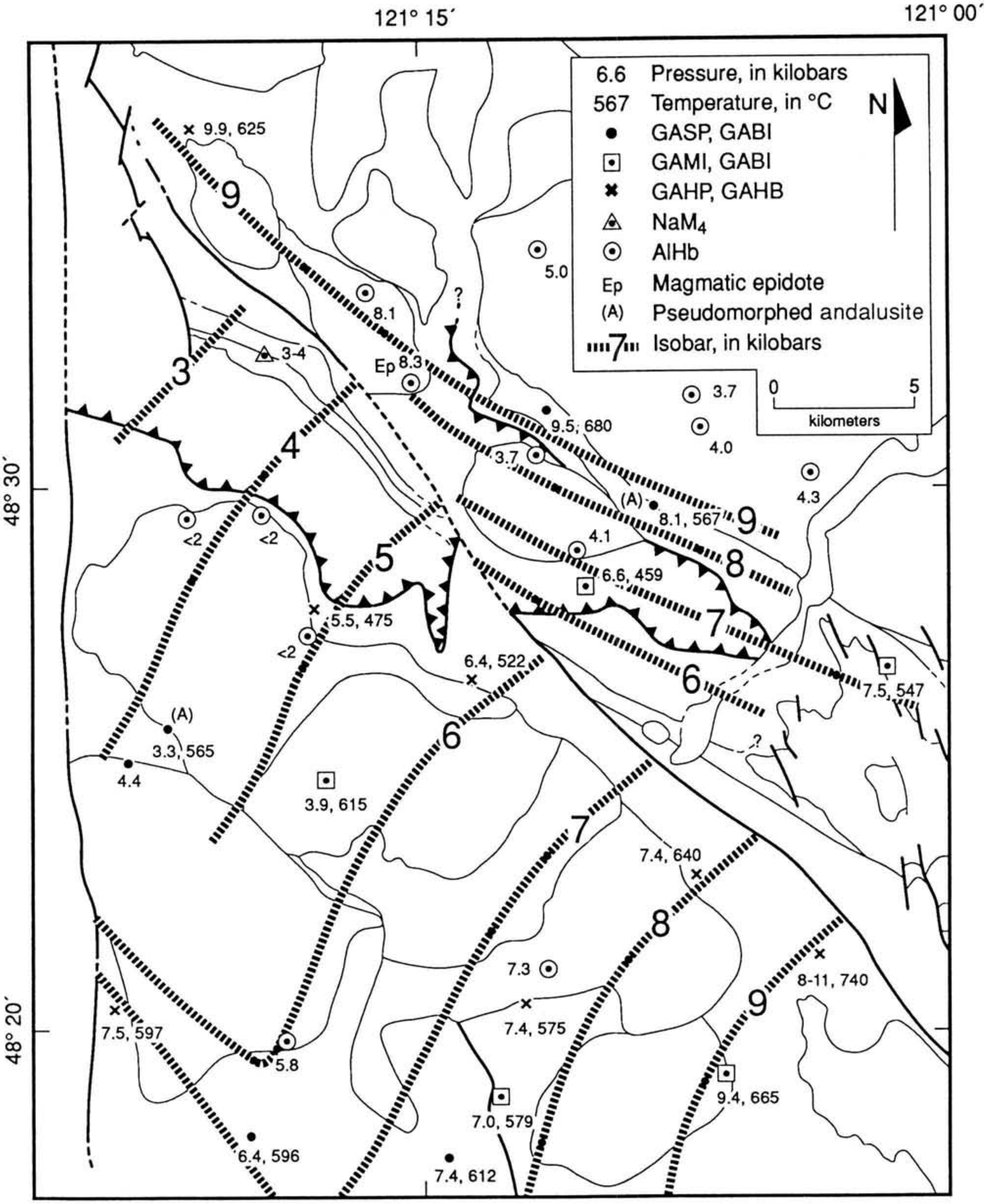

Figure 8. Thermobarometric data and inferred isobars for the Cascade River area. GASP, garnet - aluminum silicate - plagioclase equilibrium, calibration of Berman (1991); GABI, garnet - biotite Fe-Mg exchange, calibration of Berman (1991); GAMI, garnet - biotite - muscovite - plagioclase equilibrium, calibration of Berman (1991); GAHP, garnet - hornblende - plagioclase equilibrium, Kohn and Spear (1990); GAHB, garnet - hornblende Fe-Mg exchange, calibration of Graham and Powell (1984); NaM_4, crossite content of Na-amphibole, Brown (1977); AlHb, aluminum-in-hornblende barometer, calibration of Johnson and Rutherford (1989). See Table 1, at the end of the text, for details of mineral compositions.

Thermobarometry in higher grade rocks was carried out by applying a number of mineral equilibria involving garnet. These are listed in Table 1 (at the end of the text) together with microprobe analyses and references to publications from which the calibrations were obtained. For zoned garnets, we used compositions representing peak metamorphic conditions; in nearly all garnets this composition is at the rim or near the rim. We tried to work only with samples showing textural evidence of equilibrium among the phases of interest.

On the northeast side of the Entiat fault, the high-grade schist of the staurolite zone represents the southwest flank of the Skagit belt, the interior of which is reported by Whitney (1992) to have crystallized at approximately 9 kb and 720°C. We obtained comparably high pressures, 9.5–9.9 kb, but somewhat lower temperatures, 625° to 680°C, along the northeast side of the present study area (Fig. 8). From the area below the staurolite isograd but above the albite-out isograd, two specimens gave pressures of 6.6 and 7.5 kb and temperatures of 459° and 547°C, respectively (Fig. 8; Table 1). The transition to low-grade rocks is cut off by the Entiat fault in the northwest part of the map area, but it is apparently unfaulted along strike to the southeast. Here, in the vicinity of Johannesburg Mountain, the metamorphic gradient appears to be impossibly abrupt. Nine-kilobar rocks are only 5 or 6 km map distance from 6-kb rocks, and lower greenschist facies rocks are only a few kilometers away. Throughgoing faults seem unlikely because distinctive map units extend across this transition. Perhaps our barometry is in error. Possibly there is a kinetic explanation; for example, the gradient may represent an incompletely developed late high-pressure overprint on a regionally extensive lower pressure terrane.

On the southwest side of the Entiat fault, metamorphic grade increases to the southeast from the chlorite zone to the amphibolite facies along the strike of the regional structure. Pressures range from 3 to 4 kb on Lookout Mountain to more than 9 kb in the southeast corner of the map area. There is no apparent post-metamorphic fault offset along this gradient. Isobars defining this gradient are shown on Figure 8 to swing sharply northwestward in the southwestern part of the area. This interpretation is based on one data point (7.5 kb) on the west side of the Chaval pluton and on metamorphic patterns and ages in the Chiwaukum schist south of the study area, as discussed in Brown and Walker (1993).

We found evidence for a late, high-pressure metamorphic overprint in schists at a locality on the southwest contact of the Eldorado pluton near the North Fork of the Cascade River (McShane, 1992). Here, relict porphyroblasts of andalusite preserved as idioblastic prisms 5 mm across are replaced by kyanite, staurolite, biotite, muscovite, plagioclase, and quartz. Garnets in this rock are strongly zoned and have idioblastic growth rims (Fig. 9). Rim compositions together with matrix plagioclase and biotite yield P-T values of 8.1 kb, 567°C (Figs. 8 and 9). Garnet cores, matched with plagioclase cores and assuming temperature within the andalusite stability field (Holdaway, 1971), give a pressure in the range of 2–3 kb. This sample provides evidence for an early low-pressure metamorphic event followed by a pressure increment on the order of 5 kb. The high pressure overprint is consistent with other high-pressure values nearby and suggests that the metamorphism of the Skagit belt has a similar history, a concept more fully developed later in this paper.

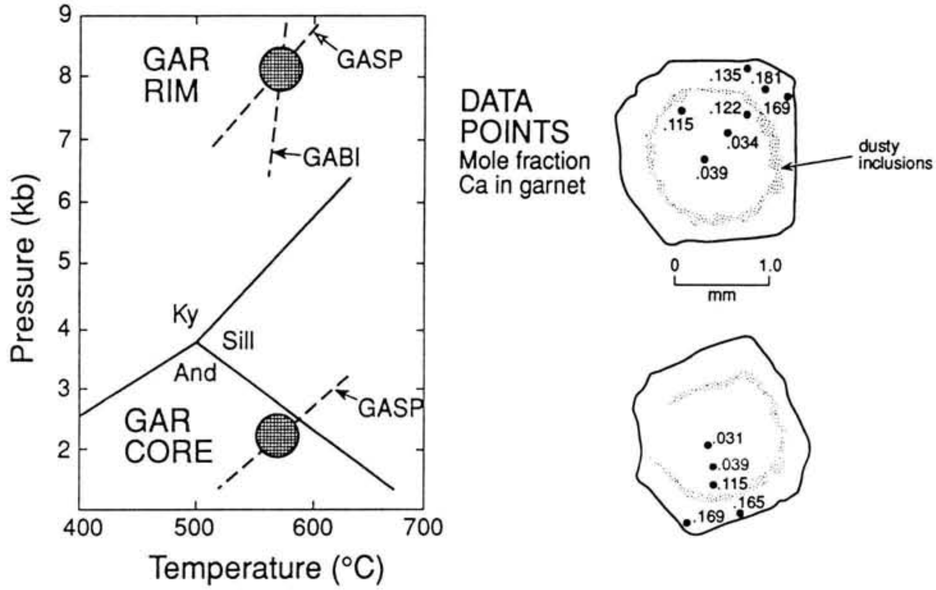

Figure 9. Zoned garnet from the contact aureole of the Eldorado pluton, North Fork of the Cascade River. See Table 1 and text for details of the thermobarometry, and McShane (1992) for further details.

METAMORPHIC FABRIC

Country-rock schists of the region are predominantly L-S tectonites with well-developed foliation striking northwest and dipping steeply southwest (Fig. 10) and with subhorizontal (strike-parallel) mineral and stretching lineations (Fig. 11). Much less common, but notable, are schists with flattening fabrics and no lineation or only a weak down-dip mineral lineation. The fabric characterized by strike-parallel lineation is found throughout the area and is defined by mineral assemblages of all metamorphic grades. Retrograde shear zones also typically bear this fabric. Shear sense features associated with the strike-parallel lineation in metamorphic fabrics are fairly abundant, and about 90 percent are dextral (Fig. 11). Metaclastic rocks show evidence of high strain in many areas; X/Y strain ratios of as

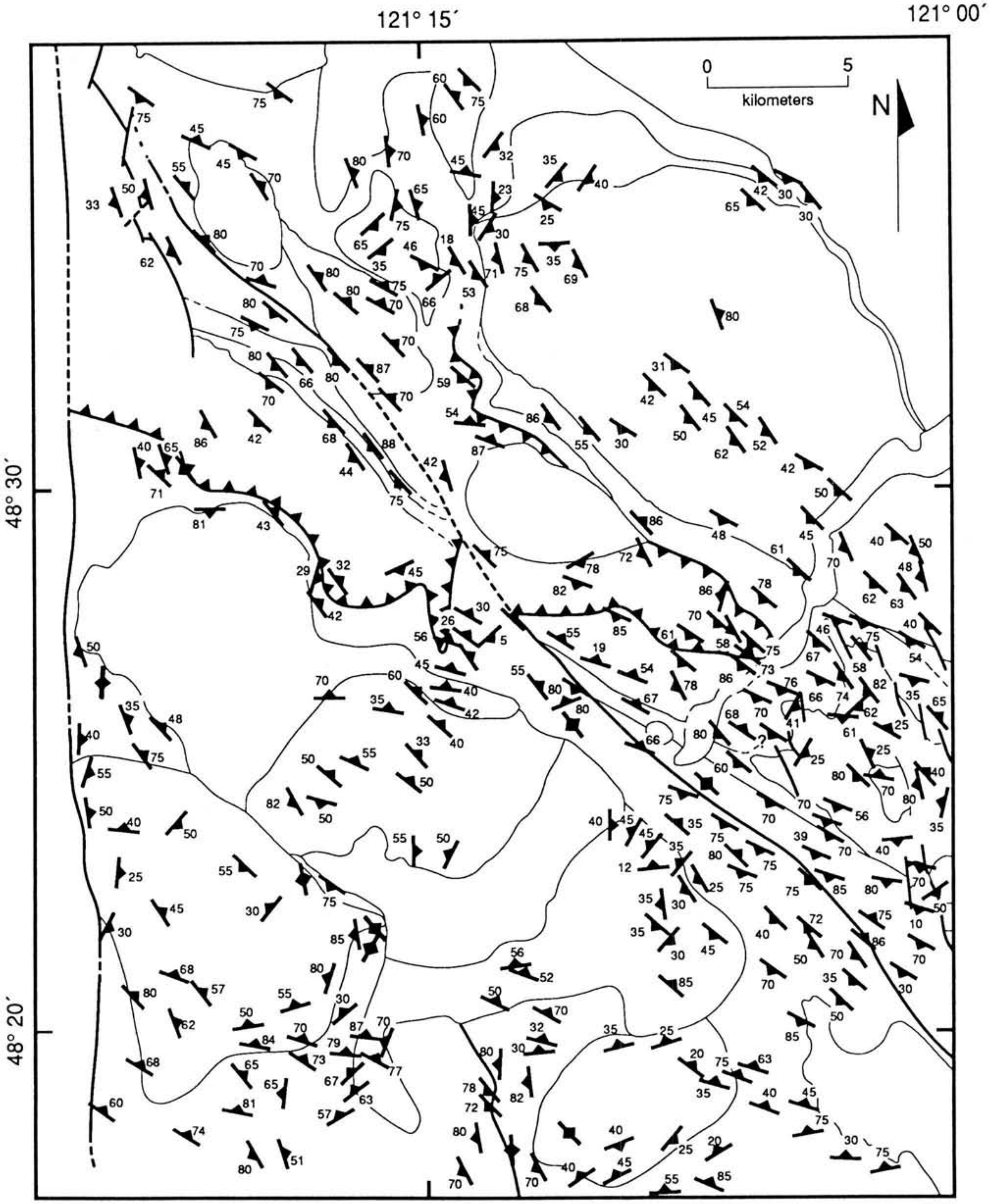

Figure 10. Foliation attitudes in the Cascade River area. Data from Tabor and others (1988), Bittenbender (1991), and this study.

much as 10:1 are observed (Fig. 12). The strain geometry in fabrics that have the strike-parallel lineation varies from plane strain to constrictional (Fig. 12). This lineation appears to represent a right-lateral strike-slip shear regime operative during much of the orogenic activity of the region.

The flattening fabric is characterized by pancake-shaped clasts in metaconglomerates and by unoriented

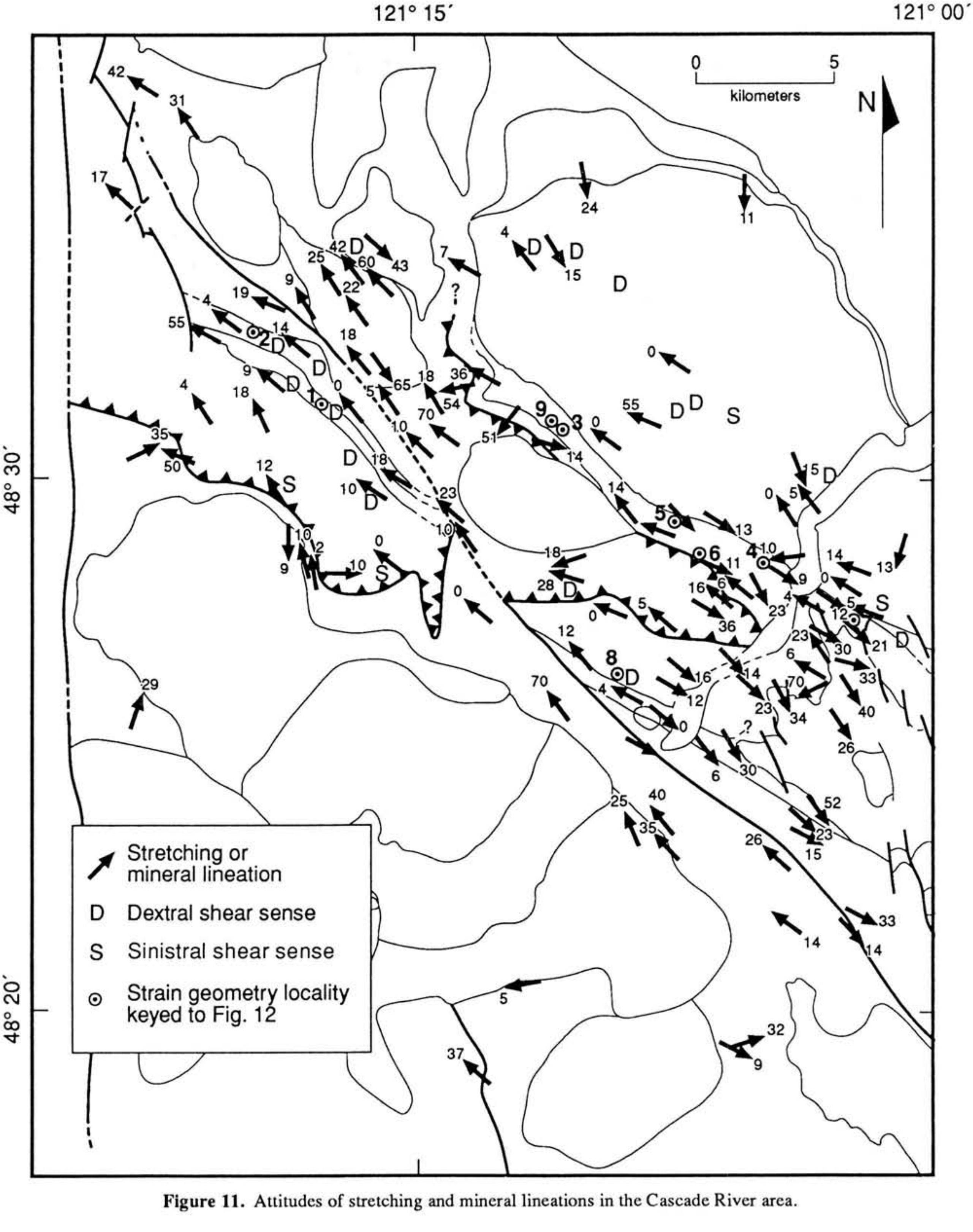

Figure 11. Attitudes of stretching and mineral lineations in the Cascade River area.

mineral laths lying in the foliation plane (garbenschiefer texture). Locally a weak down-dip mineral alignment is observed associated with this fabric. The flattening fabric is mostly restricted to the staurolite zone (Figs. 5, 11, and 12) and is defined only by high-grade, high-pressure minerals in the Sibley Creek area and on the north side of Johannesburg Mountain. The flattening fabric is in many places crosscut by dextral strike-slip shear zones along which retrograde minerals are developed.

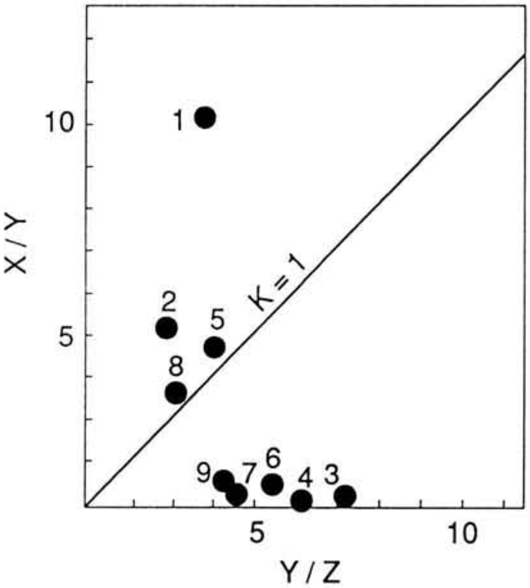

Figure 12. Flinn diagram for metaconglomerates in the Cascade River area. The numbers are keyed to localities on Figure 11. X, Y, and Z refer to the maximum, intermediate, and least principal strain axes, respectively. K = 1 line separates field of constrictional strain (above) from field of flattening strain (below).

STRUCTURAL AND METAMORPHIC SETTING OF THE PLUTONS

The major plutons, not including the pre-tectonic MMQD, all have intrusive contacts with country-rock schists. This finding for the Eldorado pluton (McShane, 1992) is a revision of previous reconnaissance mapping (Misch, 1966; Tabor and others, 1989) in which the Eldorado was considered to be fault bounded. Although contacts of this pluton are sheared against country rock in many places, intrusive relations can be seen in excellent exposures in the headwaters of Newhalem Creek and on the ridge west of The Triad. At these localities, plutonic rock crosscuts structure in the schist and occurs as apophyses in the schist; furthermore, xenoliths of schist are present in the plutonic rock.

Although crosscutting features are observed in detail at the contacts of all the Late Cretaceous plutons, country-rock foliations are generally conformable with pluton margins. The Bench Lake, Cyclone Lake, Downey Creek, Eldorado, Haystack, and Marble Creek plutons contain both metamorphic and igneous fabrics developed in different zones but of mutually similar orientations and of similar orientation to fabric in the country rock. The strike-parallel lineation, down-dip lineation, and flattening fabric observed in the country rock are seen also in the plutons. In the Eldorado pluton the fabric is much more strongly developed within a half a kilometer of the margin than in the interior of the body (as observed also by Tabor and others, 1988). These plutons are reasonably considered to be syntectonic (Misch, 1966). Several other Late Cretaceous plutons, the Jordan Lake, Alma Creek, and Hidden Lake Peaks, mostly lack directional fabric and are considered to be late- or post-tectonic (Misch, 1966; Haugerud and others, 1991; Tabor and others, 1988; this study).

Mineralogic barometers used for determining depth of pluton emplacement include: aluminum in hornblende; the equilibrium among garnet, muscovite, biotite, and plagioclase; and the presence of magmatic epidote, which is inferred to indicate a pressure of crystallization greater than 6 kb (Zen and Hammarstrom, 1984). See Table 1 for details.

Mineral barometry in the Chaval, Bench Lake, Downey Creek and Marble Creek plutons yields pressures comparable (within uncertainty limits) to those derived for country rocks (Fig. 8).

Hornblende barometry in the Jordan Lake, Hidden Lake Peaks, and Eldorado plutons gives significantly lower pressure values than are derived for the adjacent country rock (Fig. 8). These discrepancies possibly reflect faulty barometric tools, but we are inclined to think they are real. The Jordan Lake and Hidden Lake Peaks plutons lack metamorphic fabric and could well have been intruded at a shallow level after uplift of the regional metamorphic rocks. A problem with this explanation for the Hidden Lake Peaks pluton is that its zircon age (74 Ma) is similar to that of the neighboring deep-level Marble Creek and Haystack plutons. For the Eldorado pluton, we think regional burial followed pluton emplacement, as discussed below.

Shallow intrusion of the Eldorado pluton is indicated by several lines of evidence:

(1) Andalusite replaced by kyanite and other minerals (described above) occurs in the Eldorado contact aureole, indicating pressure less than 3.7 kb (aluminum silicate calibration of Holdaway, 1971).

(2) The Eldorado pluton lacks igneous epidote but contains the equivalent lower pressure assemblage of K-feldspar + hornblende; pressures were less than 6 kb.

(3) Aluminum in hornblende indicates pressures in the range of 3.7–5.0 kb at four localities in the relatively unaltered core of this pluton.

AGE OF METAMORPHISM

Because the deep-level Marble Creek pluton is younger (76 Ma) than the Eldorado pluton (88–90 Ma) and in view of the evidence in the contact aureole of the Eldorado pluton for a late high-pressure overprint (discussed above), we conclude that the major metamorphic loading event northeast of the Entiat fault occurred after 88–90 Ma and before 76 Ma. This finding is in accord with other studies elsewhere on both flanks of the Skagit complex that indicate development of the orogenic load between 90 and 75 Ma (Miller and others, 1993).

Southwest of the Entiat fault, metamorphism is synchronous with or older than the deep-level plutons that occupy this zone. The Chaval pluton (92 Ma) is a deep-level intrusive at its southern end. This pluton crosscuts country-rock fabric and bears an internal fabric that is dominantly of magmatic origin (Bittenbender, 1991). Orthogneiss in migmatite in the 9-kb zone in the southeast corner of the map area (Fig. 8) yields a zircon age of 95 Ma. Just outside the map area within the 9-kb zone to the southeast, the Sulphur Mountain pluton (96 Ma) contains magmatic epidote and thus was also emplaced at deep level. We conclude that crustal loading in this block occurred at or before about 96 Ma.

K-Ar ages of mica in Cretaceous or older rocks across the study area are mostly in the range of 45 to 60 Ma, indicating a considerable residence time at depth before uplift and cooling. An exception is a 94 Ma age of muscovite in the chlorite zone in the northwest part of the area (Fig. 2; age from Tabor and others, 1988). The temperature of metamorphism (≈400°C) was probably about the same as the Ar blocking temperature. Therefore, this age may be close to that of the metamorphism in this area and is in agreement with the age suggested above for related higher grade rocks to the southeast.

MECHANISMS OF OROGENY

Many hypotheses have been advanced to explain deformation and metamorphism in the Cascade crystalline core. Misch (1966) invoked a contractional model; he (p. 136) viewed the crystalline core of the Cascades as a "compressively uplifted crustal wedge" occupying a root zone area for thrusts verging northeast and southwest outward from this central zone, and he favored origin of the granitic rocks as (p. 141) "end products of regional metamorphism", a mechanism suggested also by Zen (1988).

Whitney and McGroder (1989), McGroder (1991), and Whitney (1992) interpreted the Cascades orogen in terms of a collisional model in which the Skagit belt represents the deepest (28–34 km) exposed part of a southwest-vergent thrust stack formed between the Insular superterrane and North America. They suggest that the Eldorado pluton was intruded at the peak of orogeny. In the McGroder (1991) model, the Nason terrane formed at a higher structural level and represents a thrust sheet overlying the Skagit belt; McGroder suggests that the primary accretionary suture between the Insular superterrane and North America lies between these two belts, that is, in the Cascade River area. The role of the granitic plutonic rocks is not addressed in this model.

Kriens and Wernicke (1990), working in the eastern part of the Skagit belt near Ross Lake, proposed a contractional orogenic model which attributes Skagit metamorphism to 90 Ma arc magmatism at middle crustal levels (<23 km) in an already structurally thickened crust; peak pressures (9–10 kb, 35 km) are inferred to have been attained prior to this event. The thickening event is not specifically addressed; however, they argue for east-vergent thrusting in the eastern Skagit region between 110 and 95 Ma. Latest Cretaceous and early Tertiary zircon ages of Skagit orthogneiss are interpreted to represent blocking ages of plutons intruded earlier, at about 90 Ma. These authors emphasize evidence for a continuity of section from the high-grade Skagit to low-grade rocks in the Methow basin and suggest that no structural break occurs on the northeast flank of the Skagit belt.

Brown and Talbot (1989) noted the prevalence of steep northwest-striking foliations and strike-parallel stretching lineations throughout the crystalline core and suggested that this fabric represents orogen-parallel strike-slip shear during the peak of orogeny. They interpreted the crystalline core as the exhumed root of a transpressional magmatic arc.

Walker and Brown (1991) noted contemporaneity of plutonism and metamorphism in the Nason belt on the basis of zircon geochronology and study of metamorphic fabrics. They suggested that the plutons are not the product of metamorphism related to crustal thickening but instead are arc related. Brown and Walker (1993), following suggestions of Evans and Berti (1986) and citing a variety of features including metamorphic patterns near the Spuzzum and Scuzzy plutons in the southeast part of the Coast Plutonic Complex in British Columbia, proposed that Barrovian metamorphism in the Cascades crystalline core is primarily the product of pluton emplacement rather than tectonic thickening.

How do our findings in the Cascade River area pertain to these orogenic models? Our contributions are as follows:

(1) Stratigraphy connects high- and low-grade rocks on the southeast flank of the Skagit belt. Therefore, displacements across this zone are limited, and the observed metamorphic gradient represents a more or less continuous crustal cross section. This finding is similar to the interpretation of Kriens and Wernicke (1990) for the northeast side of the Skagit belt.

(2) Metamorphic zones are developed across terrane-bounding thrusts within the study area, and therefore the metamorphism can be interpreted to postdate and not be related to the terrane-stacking thrust event.

(3) The observation of a low-grade metamorphic zone in the Cascade River area between the Skagit and Nason metamorphic culminations would seem to preclude the regional thrust stacking model of McGroder (1991), which places this entire region at lower crustal levels.

(4) Any thrust model encounters difficulty with the boundaries of the loaded panels as defined by metamorphic gradients. The Skagit block requires a steep-sided load rooting in the northeast, whereas the metamorphic gradient just on the other side of the Entiat fault in the Nason terrane requires a steep-sided load rooting to the southeast.

(5) Metamorphic ages indicate diachronous loading. The Nason belt was depressed at or before about 96 Ma, whereas the Skagit belt was not loaded until after about 88 Ma. These age and depth constraints pose another difficulty for the McGroder model, which places the Eldorado pluton under the Nason belt in a regional thrust stack. The timing of Skagit loading also contradicts the Kriens and Wernicke (1990) model, which invokes high-pressure metamorphism in the Skagit belt

prior to 90 Ma arc magmatism and requires uplift to middle crustal levels by 90 Ma for consistency with their inferred continuous crustal section from the Skagit belt into low-grade rocks.

The magmatic loading model of Brown and Walker (1993) was derived in part from geology in the Cascade River area. However, the full body of evidence for this interpretation includes features outside this area and is discussed separately (Brown and Walker, 1993). Briefly, this hypothesis explains the metamorphism of the Cascades crystalline core as being the product of crustal downwarping in zones of pluton ascent. As the magma rose to upper levels of the crust, country rock was displaced downward, taking the place of the magma in lower parts of the crust. We envisage both the Nason and Skagit metamorphic culminations to represent separate zones of voluminous magmatic ascent; they formed independently of one another and mark a shifting axis of magmatism. The Cascade River area lies between these zones. Uplift is interpreted to be isostatic, beginning as magmatism died out and erosion removed the crustal load.

Strain features in the Cascade River area are interpreted by us to represent two kinematic processes operating simultaneously. The strike-slip fabric records a transcurrent tectonic motion, perhaps localized by thermal weakening of country rock along a magmatic axis, and represents an oblique component of plate convergence. We do not envisage large strike-slip displacement because the stratigraphy across the Cascade River area is more or less continuous. The coeval flattening fabric and associated weak down-dip mineral lineation record flattening and vertical flow of country rock in response to pluton emplacement. We cannot rule out the possibility that this fabric represents a component of tectonic contraction, but note that the magmatic explanation is equally viable.

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Table 1. Database for thermobarometric analyses. (See Fig. 8 for locations.) References [see list of cited references for full citation]: 1, Dragovich, 1989; 2, E. H. Brown and N. Walker, unpubl. data; 3, McShane, 1992; 4, Dougan, 1993; 5, Fluke, 1992; 6, Bittenbender, 1991; 7, Cary, 1990

A. Equilibria involving garnet. Equilibria used: 1, GABI (garnet - biotite Fe-Mg exchange; Berman, 1991) and GASP (garnet - aluminum silicate - plagioclase - quartz; Berman, 1991); 2, GABI and GAMI (garnet - biotite - muscovite - plagioclase; Berman, 1991); 3, GAHB (garnet - hornblende Fe-Mg exchange; Graham and Powell, 1984) and GAHP (garnet - hornblende - plagioclase; Kohn and Spear, 1990). Ky, kyanite; And, andalusite

Sample number	Unit	Latitude (degrees and minutes)	Longitude (degrees and minutes)	Temp. (°C)	Pres. (kb)	Al_2SiO_5	Equil. used	Ref.
119-183	Cascade River Schist	48 31.2	112 11.0	–	9.5	Ky	1	1
119-134	Cascade River Schist	48 31.3	121 11.5	680	–	–	1	1
164-30	Napeequa Schist	48 28.1	121 11.0	459	6.6	–	2	1
164-32d	Napeequa Schist	48 36.4	121 21.3	625	9.9	–	3	2
164-37c	Napeequa Schist	48 25.7	121 22.4	565	3.3	And	1	2
164-73c	Napeequa Schist	48 26.5	121 14.5	522	6.4	–	2	2
164-335f	Chiwaukum Schist	48 20.4	121 23.9	597	7.5	–	3	2
169-HLF	Cascade River Schist	48 29.4	121 8.6	567	8.1	Ky	1	3
171-23b	Banded gneiss	48 21.4	121 4.0	740	8-11	–	3	4
174-8a	Banded gneiss	48 21.0	121 7.1	665	9.4	–	2	5
174-24	Napeequa Schist	48 20.7	121 12.2	575	7.4	–	3	5
174-33a	screen/Cyclone Lake	48 25.3	120 17.8	615	3.9	–	2	5
174-36c	Banded gneiss	48 22.7	121 7.5	640	7.4	–	3	5
174-45b	Chiwaukum Schist	48 18.5	121 10.3	579	7.0	–	2	5
174-118a	Napeequa Schist	48 27.9	121 19.0	475	5.5	–	3	5
OHM20	Chiwaukem Schist	48 17.6	121 19.7	596	6.4	Ky	1	6
ODM22	Chiwaukum Schist	48 2.5	121 14.6	612	7.4	Ky	1	6
RT48A58	Cascade River Schist	48 26.6	121 1.9	547	7.5	–	2	2

Mineral composition in molar units for samples in Table 1A (G, garnet; P, plagioclase; B, biotite; M, muscovite; H, hornblende)

	119-183		119-134			164-30				164-32d			164-37c		
	G	P	B	G	P	B	G	M	P	G	H	P	B	G	P
Si	3.01	–	–	3.01	–	5.95	5.87	6.80	–	6.00	6.49	–	5.91	6.02	–
Al^4	–	–	–	–	–	2.05	–	1.20	–	–	1.51	–	2.09	–	–
Al^6	1.96	–	–	1.96	–	1.49	4.03	4.53	–	3.96	0.96	–	1.75	3.98	–
Ti	–	–	–	–	–	0.37	–	0.06	–	–	0.04	–	0.22	–	–
Fe	1.30	–	2.48	2.12	–	3.03	3.22	0.33	–	2.83	1.69	–	2.48	4.63	–
Mg	0.62	–	2.04	0.35	–	2.00	0.19	0.32	–	0.58	2.36	–	2.57	0.84	–
Mn	0.81	–	–	0.13	–	0.04	1.05	–	–	0.77	0.05	–	–	0.22	–
Ca	0.30	0.30	–	0.36	0.27	–	1.71	–	0.18	1.89	1.90	0.28	–	0.29	0.34
Na	–	0.70	–	–	0.73	0.02	–	0.13	0.84	–	0.42	0.72	–	–	0.64
K	–	–	–	–	–	1.91	–	1.93	–	–	0.06	–	1.81	–	–

	164-73c				164-335f			169-HLF			171-23b		
	B	G	H	P	G	H	P	B	G	P	G	H	P
Si	5.96	5.98	6.46	–	5.92	6.29	–	5.91	5.99	–	5.98	6.29	–
Al^4	2.04	–	1.54	–	–	1.71	–	2.09	–	–	–	1.71	–
Al^6	1.52	3.98	1.08	–	3.96	1.00	–	1.70	4.01	–	3.94	0.76	–
Ti	0.24	–	0.06	–	–	0.12	–	0.23	–	–	–	0.14	–
Fe	2.46	3.77	1.99	–	3.73	1.78	–	2.30	4.19	–	3.32	2.61	–
Mg	2.80	0.60	2.01	–	0.88	2.19	–	2.77	0.68	–	0.38	1.46	–
Mn	0.02	0.73	0.01	–	0.29	0.03	–	0.01	0.13	–	0.18	0.02	–
Ca	–	0.96	1.73	0.34	0.12	1.86	0.44	–	0.98	0.32	2.25	1.93	0.33
Na	0.02	–	0.54	0.67	–	0.42	0.59	0.06	–	0.68	–	0.38	0.64
K	1.96	–	0.04	–	–	0.10	–	1.85	–	0.01	0	0.36	0.02