SELECTED PAPERS ON THE GEOLOGY OF WASHINGTON

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This volume is in memory of

Randall L. Gresens
(1935-1982)

E. Bates McKee, Jr.
(1934-1982)

and in honor of

Julian D. Barksdale
(1904-1983)

Howard A. Coombs

Peter Misch

A. Lincoln Washburn

and

Harry E. Wheeler
(1907-1987)

for their service to the University of Washington,
Department of Geological Sciences, and
their contributions to the geological sciences.

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PREFACE

In the past few decades the geology of virtually every region of Washington has been so extensively revised that most previous descriptions are badly out of date. Much of the credit for these revisions belongs either directly or indirectly to five long-time faculty members of the Department of Geological Sciences at the University of Washington: J. D. Barksdale, H. A. Coombs, P. Misch, A. L. Washburn, and H. E. Wheeler. They either were directly responsible for much of the new data and interpretations through their own research or were indirectly responsible through the work of their colleagues and former students.

In 1980 Eric S. Cheney conceived the idea of a symposium on the geology of Washington to honor these five, each of whom had recently retired from the University of Washington. He discussed the idea with E. Bates McKee, Jr., who was in the process of revising *Cascadia*, his famous textbook on the geology of the Pacific Northwest. McKee was equally enthusiastic, and together they framed the concept that each speaker should involve as many co-authors as possible and should present a regional overview of some part of the State. Cheney and McKee approached the Cordilleran Section of the Geological Society of America about such a symposium and were immediately encouraged to proceed by Margaret Woyski, the Section's Program Chairperson.

Cheney was primarily responsible for the symposium, and McKee was to be the editor for a volume incorporating the papers of the symposium. V. E. (Ted) Livingston, State Geologist, and J. Eric Schuster, Assistant State Geologist, volunteered that the Washington State Division of Geology and Earth Resources would be interested in assisting in publication of the symposium volume. The symposium volume was to include brief biographies (prepared by Barksdale) of the five honorees. These biographies are included at the end of this volume, but two of the honorees (Barksdale and Wheeler) have since died, so their biographies have been revised and expanded by Cheney.

The symposium was held on April 20, 1982, as part of the 78th annual meeting of the Cordilleran Section of the Geological Society of America in Anaheim, California. The program of the symposium is reproduced here as Table 1. The abstracts were published by the Society (Geological Society of America Abstracts with Programs, v. 14, no. 4).

The symposium was highly successful. A meeting room capable of seating perhaps 75 to 100 people had been allotted for the symposium. The room was filled to overflowing by the time the symposium began, and after Cheney's opening remarks the gathering was moved to a larger hall. Attendance probably approached 500 people. Papers were delivered as shown in Table 1, except that H. A. Coombs was unable to attend because of illness. In his place, R. B. Waitt, Jr., gave a paper on the March-April 1982 eruptions of Mount St. Helens.

Chairpersons of the sessions were other major contributors to the geology of the State. The late V. E. Livingston had to withdraw as a chairperson for reasons of health just before the program was printed. Likewise, the late A. E. Weissenborn of the U.S. Geological Survey in Spokane had to withdraw after the program was printed. The sessions ultimately were chaired by McKee, John T. Whetten (a former chairperson of the Department of Geological Sciences, University of Washington), and Schuster.

The emotional apogee of the symposium came after Barksdale delivered his paper on the biographies of the five honorees. Characteristically, Barksdale had left almost no time for his autobiography. He was foiled by Whetten, who used his prerogative as chairperson to insert a few remarks. When Whetten had finished, there was hardly a dry eye in the hall, and the five honorees were given a standing ovation.

Three months after the symposium Bates McKee, Randy Gresens, and Randy's wife, Mimi, were killed in a plane crash near Wenatchee. Cheney then took on the responsibility of seeing that a symposium volume would be published. The five retired faculty members suggested that the volume be dedicated to McKee and Gresens, and, accordingly, their memorials are included with the biographies or memorials of the five original honorees at the end of this volume.

Cheney and the new state geologist, Raymond Lasmanis, issued a statement on November 29, 1982, that confirmed future publication of a symposium volume and called for papers. Since then there has been no doubt that the volume would be published, but it has taken longer than expected to carry through on the promise. Response was positive from the majority of the symposium speakers, and papers began to be prepared in spite of busy schedules. Cheney's schedule, including work in South Africa, prevented him from

Table 1.–Symposium Program

TUESDAY, APRIL 20, 1982

SYMPOSIUM: THE REGIONAL GEOLOGY OF THE STATE OF WASHINGTON, IN HONOR OF J. D. BARKSDALE, H. A. COOMBS, P. MISCH, A. L. WASHBURN, AND H. E. WHEELER, PART I: Costa Mesa Room 11 & 12, Convention Center, 0800 hours

A. E. Weissenborn and J. E. Schuster, Presiding

1 Eric S. Cheney*: OPENING REMARKS 0755

2 Darrel S. Cowan*: THE CORDILLERAN TECTONIC SETTING OF WASHINGTON 0800

3 Raymond A. Price*: MID-PROTEROZOIC TO OLIGOCENE CORDILLERAN TECONIC EVOLUTION, NORTHEASTERN WASHINGTON AND ADJACENT BRITISH COLUMBIA 0820

4 William A. Rehrig*, Stephen J. Reynolds, Richard Lee Armstrong: GEOCHRONOLOGY AND TECTONIC EVOLUTION OF THE PRIEST RIVER CRYSTALLINE/METAMORPHIC COMPLEX OF NORTHEASTERN WASHINGTON AND NORTHERN IDAHO 0840

5 E. S. Cheney, L. A. Oliver, K. E. Orr*, B. P. Rhodes: KETTLE AND OKANOGAN DOMES AND ASSOCIATED STRUCTURAL LOWS OF NORTH-CENTRAL WASHINGTON 0900

6 J. R. Snook, M. A. Ellis, J. W. Mills, A. J. Watkinson*: THE KOOTENAY ARC IN N. E. WASHINGTON 0920

7 Mark R. Cole*, Marilyn E. Tennyson: LATE MESOZOIC METHOW SEQUENCE, WASHINGTON 0940

BREAK 1000

8 Peter Misch*: NORTH CASCADES GEOLOGY 1020

9 R. W. Tabor*, R. E. Zartman, V. A. Frizzell, Jr.: POSSIBLE ACCRETED TERRANES IN THE NORTH CASCADES CRYSTALLINE CORE, WA 1040

10 V. A. Frizzell, Jr.*, R. W. Tabor, R. E. Zartman, D. L. Jones: MESOZOIC MELANGES IN THE WESTERN CASCADES OF WASHINGTON 1100

11 Randall L. Gresens*: EARLY- TO MID-CENOZOIC GEOLOGY OF CENTRAL WASHINGTON 1120

12 Joseph A. Vance*: CENOZOIC STRATIGRAPHY AND TECTONICS OF THE WASHINGTON CASCADES 1140

* Speaker

SYMPOSIUM: THE REGIONAL GEOLOGY OF THE STATE OF WASHINGTON, IN HONOR OF J. D. BARKSDALE, H. A. COOMBS, P. MISCH, A. L. WASHBURN, AND H. E. WHEELER, PART II Costa Mesa Room 11 and 12, Convention Center, 1330 hours

J. T. Whetten and B. McKee, Presiding

1 Julian D. Barksdale*: FIVE ABBREVIATED BIOGRAPHIES 1330

2 Darrel S. Cowan*, Mark T. Brandon: THE GEOLOGY AND REGIONAL SETTING OF THE SAN JUAN ISLANDS 1350

Table 1.–Symposium Program (continued)

3 E. S. Cheney*: POST-EOCENE TECTONICS OF THE NORTHERN PUGET LOWLAND 1410

4 R. W. Tabor*: GEOLOGY AND STRUCTURAL EVOLUTION OF THE OLYMPIC MOUNTAINS, WASHINGTON 1430

5 John M. Armentrout*: CENOZOIC GEOLOGY OF SOUTHWESTERN WASHINGTON 1450

BREAK 1510

6 P. R. Hooper*, D. A. Swanson: EVOLUTION OF THE EASTERN PART OF THE COLUMBIA PLATEAU. 1530

7 Robert D. Bentley*, Newell P. Campbell: MIOCENE-PLIOCENE STRATIGRAPHY OF WESTERN COLUMBIA PLATEAU, WASHINGTON 1550

8 A. M. Tallman*, K. R. Fecht, J. T. Lillie, N. P. Campbell, L. G. Hanson, E. P. Kiver, D. F. Stradling, G. D. Webster: POST COLUMBIA RIVER BASALT STRATIGRAPHY OF THE COLUMBIA PLATEAU. 1610

9 Howard A. Coombs*: ENGINEERING GEOLOGY PROBLEMS IN THE COLUMBIA PLATEAU 1630

10 Stephen C. Porter*, Geoffrey A. Clayton: PLEISTOCENE GLACIATION AND VOLCANISM IN THE WASHINGTON CASCADES . . . 1650

11 Don J. Easterbrook*, D. Blunt, N. Rutter: PLEISTOCENE GLACIAL AND INTERGLACIAL CHRONOLOGY IN WESTERN WASHINGTON 1710

* Speaker

taking as active a part in preparation of the volume as he would have liked, and most of the day-to-day work of manuscript editing, corresponding with authors, and layout and planning for the volume fell to three editors or geologist/editors of the Division who worked under the direction of Schuster. These were Laura Bray, who served until her retirement in March 1985; Larry N. Stout, who worked from fall 1985 until the spring of 1986; and Katherine M. Reed, who has worked on the volume for the past year.

A comparison of this volume's table of contents with the original symposium program (Table 1) indicates that this is not a proceedings volume: it neither contains every paper presented at the symposium nor is it limited to papers presented at the symposium. Instead, some authors who gave papers at the symposium were unable to prepare papers for this volume; several of the papers contain updated and expanded information that was not available at the time the original papers were given. Some papers in this volume discuss topics not presented at the symposium, but are included here because they offer regional treatment of an important facet of the geology of Washington that otherwise would not be covered. The papers by Tennyson and Cole and by Gresens are reprinted with permission from a symposium volume published by the Society of Economic Paleontologists and Mineralogists and from Northwest Science, respectively. For these reasons we have chosen to title this volume "Selected Papers on the Geology of Washington" instead of giving it a title that would more closely link the volume with the symposium from which it grew.

Many people, in addition to those already mentioned, have assisted in the preparation of this volume. Among them are the following staff geologists of the Division of Geology and Earth Resources who reviewed drafts of most of the papers: Bonnie B. Bunning, Michael A. Korosec, Robert L. Logan, William M. Phillips, Weldon W. Rau, Henry W. Schasse, Keith L. Stoffel, and Timothy J. Walsh. Nancy E. Herman prepared final versions of many of the illustrations, Connie J. Manson verified many of the references, and Barbara E. Burfoot, Loretta M. Andrake, and J. C. Armbruster typed and coded the papers for typesetting. Most authors requested their colleagues to review their papers before they were submitted to the

Division. These reviewers are acknowledged in the papers they reviewed. We thank all for their patient efforts, without which this volume would have been impossible.

We believe that this volume fills a gap in the geologic literature and will prove useful to a broad cross section of geologists who work in the Pacific Northwest. Equally important, it pays tribute to seven, Gresens, McKee, Barksdale, Coombs, Misch, Washburn, and Wheeler, who did much to advance our knowledge of the geology of Washington.

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Seattle, Washington

April 2, 1987

A TECTONIC AND GEOCHRONOLOGIC OVERVIEW OF THE PRIEST RIVER CRYSTALLINE COMPLEX, NORTHEASTERN WASHINGTON AND NORTHERN IDAHO

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ABSTRACT

The Priest River crystalline complex, like the Kettle and Okanogan uplifts to the west, displays many structural and geochronological characteristics of Cordilleran metamorphic core complexes. Fundamentally, these characteristics include a core of deformed metamorphic and plutonic rocks with upwardly overprinted mylonitic textures, overlain in part by shallow-dipping detachment faults. Above the detachments, upper plates consist of variably tilted and faulted rocks lacking the penetrative deformation of the lower plate. Additionally, detachment structures like those along the central and eastern margins of the Priest River crystalline complex demark a thermal boundary between Eocene (40-50 million years) radiometric resetting in the lower plate and original Precambrian to Tertiary ages in the upper plate. Along the less well defined western margin of this complex, apparent age changes are gradational and not sharp.

Rocks of the Priest River crystalline complex generally consist of Precambrian gneiss and Phanerozoic granitoid rocks. Recent isotopic dating now establishes the age of oldest gneisses (Hauser Lake Gneiss) as pre-Belt ($>1{,}500$ million years). Plutonic rocks are either of Cretaceous age (90-140 million years) and thus part of the northwest batholithic terrane, or are smaller Eocene ($\approx$40-55 million years) higher level stocks, plugs, or dikes. Proterozoic, Belt-equivalent, stratified rocks occur in the complex as lower-plate, high-grade metamorphic rocks west and southwest of the Newport fault and, more commonly, as relatively unmetamorphosed sedimentary rocks in upper plates of the Newport and Purcell trench detachment faults.

The recognition of several distinct styles and relative ages of deformational/metamorphic foliation through the Priest River crystalline complex indicates that original, steeply dipping Precambrian fabrics have been overprinted by two shallow-dipping foliations: (1) a largely crystalloblastic, deep-seated foliation, uniformly affecting pre-Belt gneisses and Cretaceous plutons; and (2) a younger, less continuous, mylonitic foliation developed beneath detachment faults. Plutons either dated as or believed to be Eocene in age appear to separate development of the two low-angle foliations.

The Priest River crystalline complex is thus a part of outermost North American basement that has experienced nearly complete tectonic, metamorphic, and plutonic reconstitution and subsequent uplift through overlying Proterozoic supracrustal cover. The Lewis and Clark structural zone, elements of which form a southern boundary to the complex, links the Priest River crystalline complex with the Bitterroot metamorphic core complex to the east. This old fault zone, reactivated during Eocene tectonism, represents a regional tear or transform fault that accommodated different amounts of extension to the north and south.

INTRODUCTION

The Priest River crystalline complex, along with the neighboring Kettle and Okanogan complexes to the west (Fig. 1), possesses many of the characteristics common to metamorphic core complexes of the North American Cordillera. In essence these characteristics include a central domal terrane of metamorphic and plutonic rocks that is overprinted by an overlying and/or flanking carapace of mylonitic fabric. At least one flank of the mylonitic arch has a zone of detachment marked by pervasive brecciation, propylitic alteration (chloritic breccia and microbreccia), and a superjacent

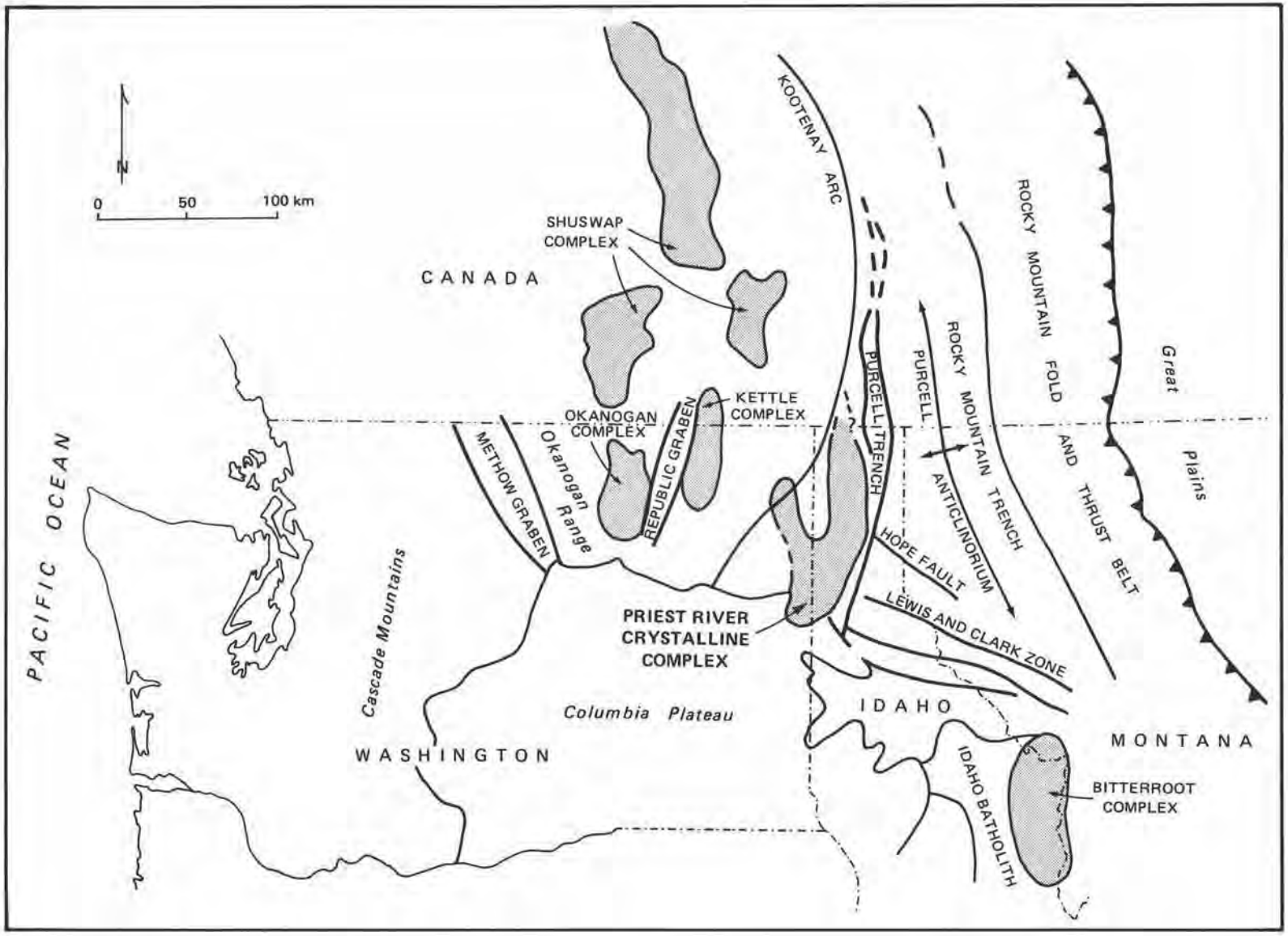

Figure 1.–Regional tectonic setting of the Priest River crystalline complex among major tectonic features of the Pacific Northwest. Metamorphic core complexes are shaded.

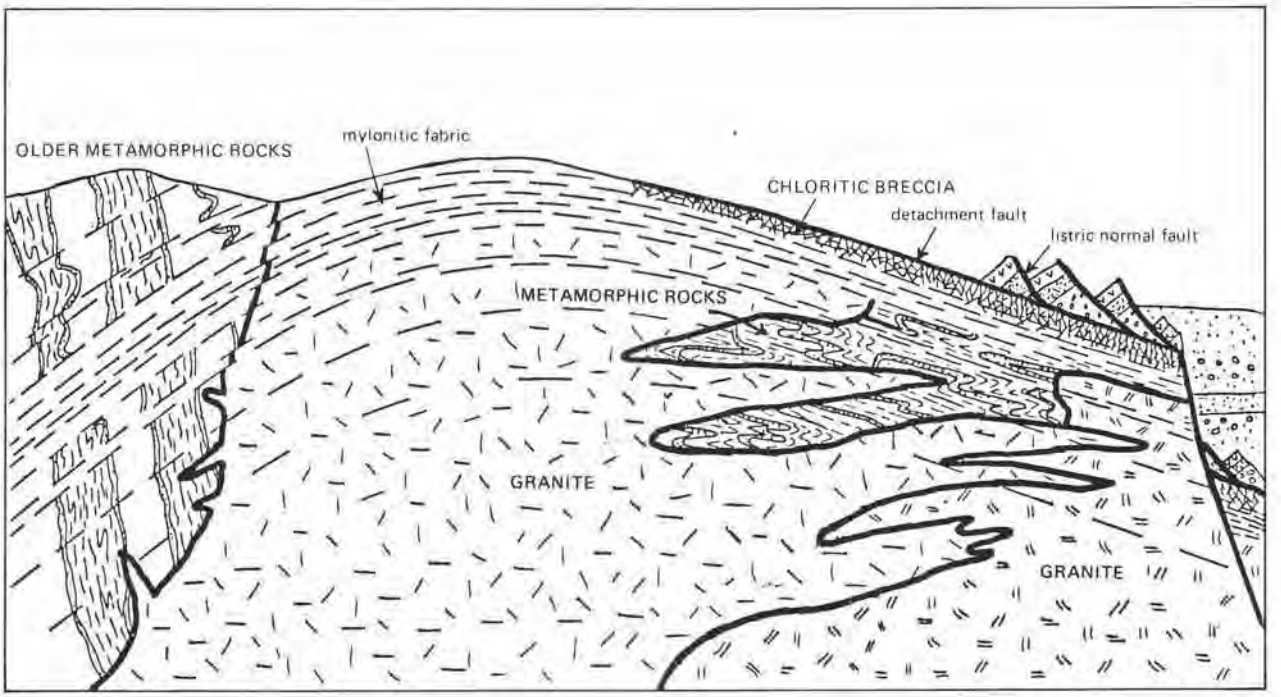

Figure 2.–Diagrammatic cross section of a Cordilleran metamorphic core complex. Large granitic bodies with superimposed mylonitic fabric commonly are of Tertiary age, thus fixing the time of deformation. The transition from mylonite to chloritic breccia usually occurs only on one flank of the arch. Mylonitization, detachment, and listric faulting predate late Tertiary, high-angle, Basin and Range faulting.

detachment fault (Fig. 2) (Crittenden and others, 1980; Armstrong, 1982). Rocks above the fault (upper plate) generally include Cenozoic and older sedimentary and volcanic sequences that are variably rotated or tilted on normal faults, but they lack the penetrative ductile to brittle deformation of lower-plate rocks. Lower-plate rocks yield Tertiary cooling ages nearly the same as ages of supracrustal volcanic rocks of the upper plate.

The Priest River crystalline complex was first recognized as a metamorphic core complex analogous to those in the southwestern United States in 1976 and 1977, when the first two authors were engaged in reconnaissance uranium exploration in Arizona and northeastern Washington. The proprietary nature of that work precluded publication of the findings until several years later (Reynolds, 1980; Rehrig and Reynolds, 1981; Reynolds and others, 1981; Rehrig and others, 1982). During that time, the similarities of the Priest River crystalline complex to other metamorphic core complexes were also noted by other workers (F. K. Miller, personal commun., 1980; Cheney, 1980). The Priest River crystalline complex has been variously referred to as the Selkirk complex and the Spokane dome by those workers.

The purpose of this paper is to present a brief overview of the Priest River crystalline complex and its tectonic and geochronologic framework. Geochronologic details are presented by Armstrong and others (this volume).

GENERAL CHARACTERISTICS OF THE PRIEST RIVER COMPLEX

The Priest River crystalline complex comprises the mountainous terrain west of the Purcell trench, extending from Coeur d'Alene Lake north at least to the southern British Columbia border (Fig. 1; Reynolds, 1980). The southwest margin is marked by the Columbia River volcanic plain. Much of the west and northwest borders of the complex lacks physiographic expression.

Fundamentally, the complex consists of an uplifted, north-trending arch of metamorphic and plutonic basement that has been affected by magmatic, metamorphic, and deformational events during Precambrian, Mesozoic, and Eocene times. The oldest rocks occur mainly in the south half of the complex and are mainly metasedimentary. Plutonic rocks underlying most of the north part of the complex have

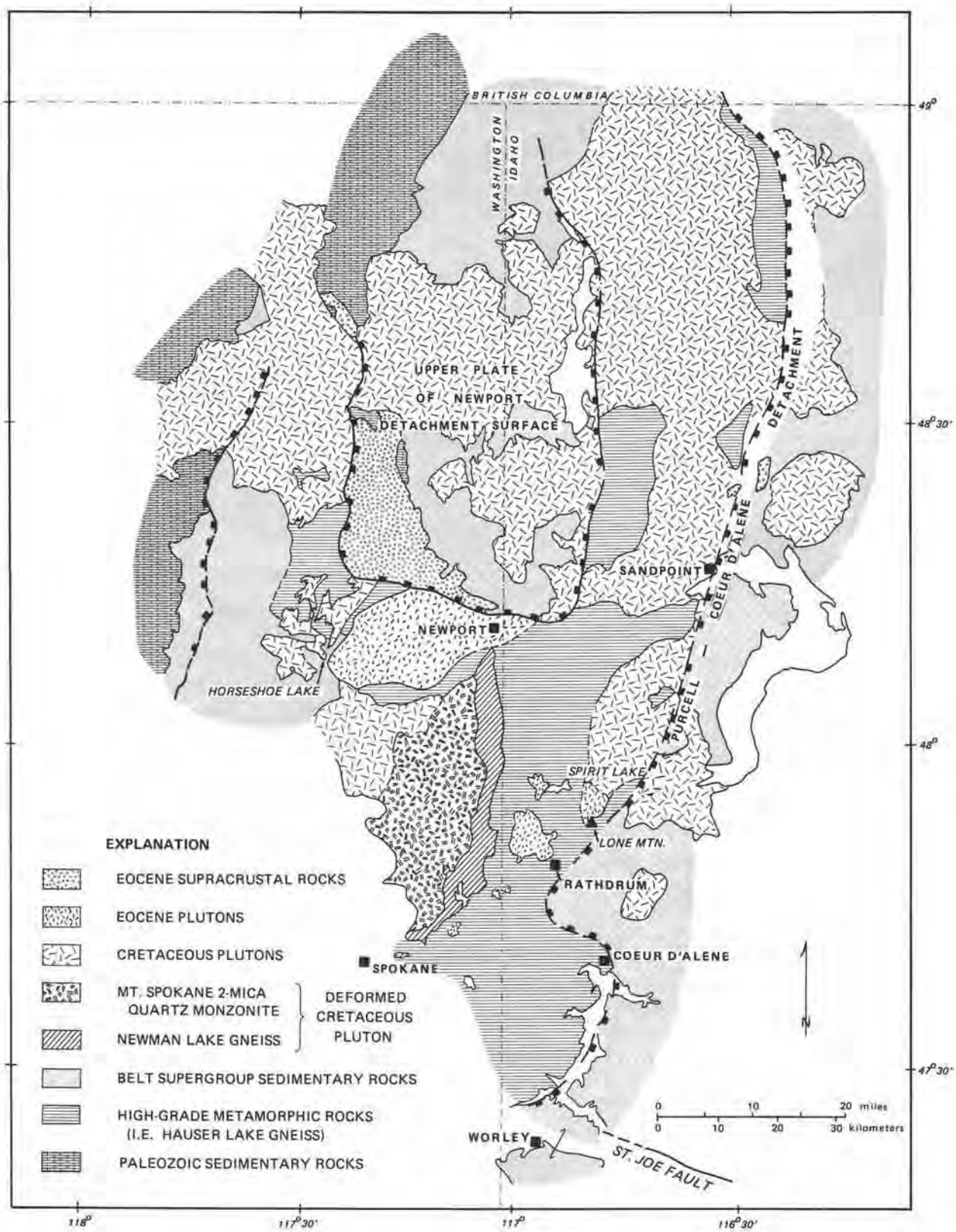

Figure 3.–Generalized geologic map of the Priest River complex. Eocene supracrustal rocks include continental sediments and the Pend Oreille Andesite. Cretaceous plutons include foliated bodies.

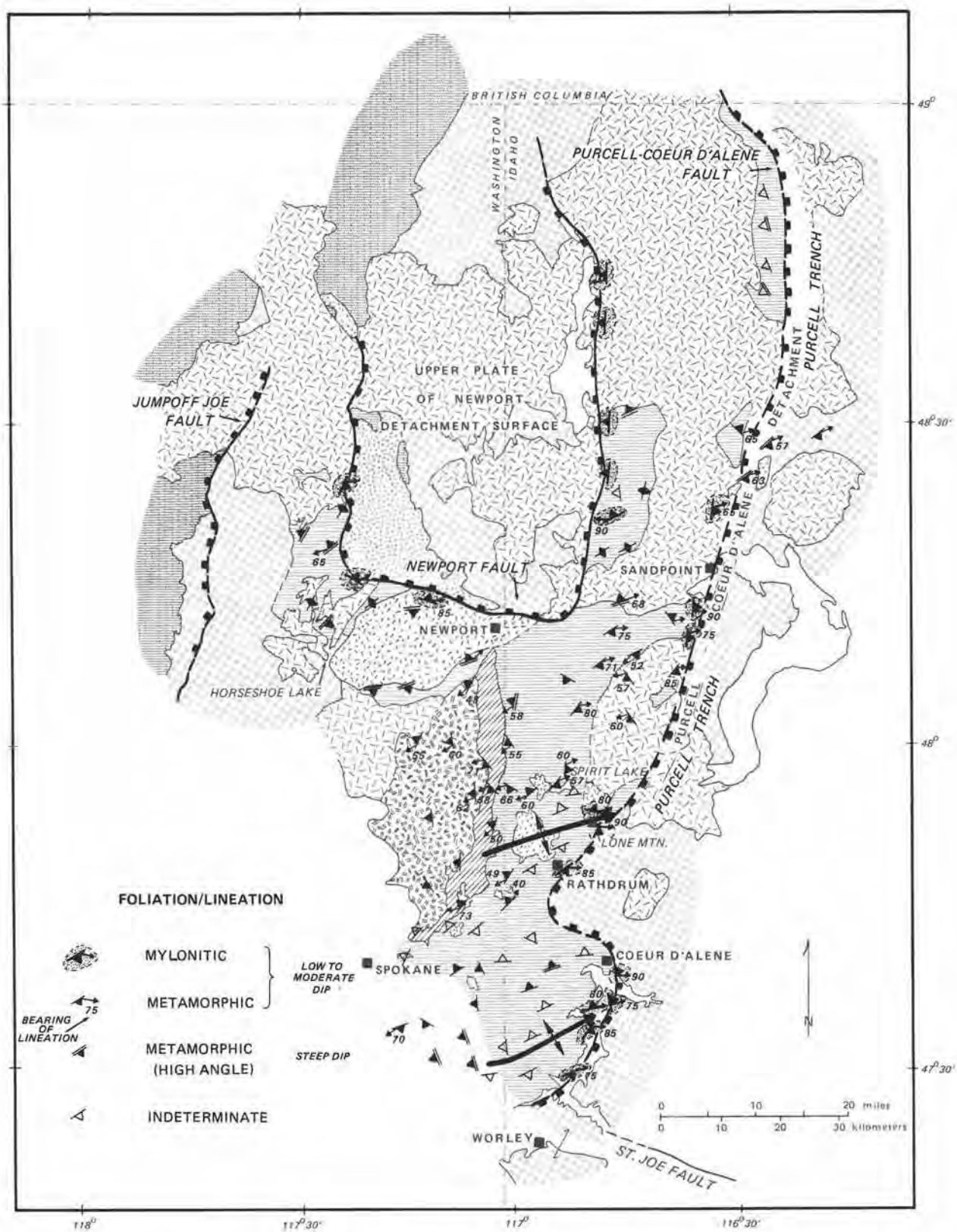

Figure 4.–Structural map of the Priest River complex, showing detachment faults, northeast-trending foliation arches, various foliation types, and lineations. (See text.)

been called the Kaniksu batholith (Ross, 1936; Park and Cannon, 1943) or the Selkirk igneous complex (Miller, 1982a,b).

The Priest River crystalline complex is flanked on the east by the Purcell trench, where high-grade metamorphic rocks of the complex are in contact with Belt Supergroup metasedimentary rocks of much lower grade (Fig. 3). As discussed further in this report, a regionally significant fault is suspected at this contact.

The west margin of the complex is more difficult to define precisely because of the westward gradation from high-grade metamorphic rocks yielding Eocene cooling ages to lower grade rocks that give Cretaceous or older cooling ages (Miller and Engels, 1975; Miller and Clark, 1975). The north-central boundary of the complex is defined by the Newport fault, a spoon-shaped, low-angle detachment fault overlain by a thick plate of Belt Supergroup rocks, Cretaceous plutons, and Eocene volcanic and sedimentary rocks. On the upper plate, the Belt section is thick (9,000 m minimum) and is unconformably overlain by lower Paleozoic rocks (Fig. 3) (Miller, 1974b; Harms, 1982).

The oldest rocks of the complex are high-grade metamorphic rocks consisting of quartzofeldspathic gneiss, garnet amphibolite, migmatite, and quartz-mica schist with sillimanite and kyanite (Fig. 3). These rocks have been collectively mapped as the Hauser Lake Gneiss (Weissenborn and Weis, 1976), Laclede augen gneiss (Clark, 1964, 1967), and correlative units (Weis, 1968; Miller, 1974c,d, 1982a,b; Barnes, 1965; Nevin, 1966). These gneisses have been considered to be either pre-Belt basement or metamorphosed Belt sedimentary rocks. Griggs (1973) mapped these rocks as metamorphosed Prichard Formation. A 1.5-b.y. zircon date from orthogneiss near Laclede (Clark, 1973) and Rb-Sr and U-Pb data for the Hauser Lake Gneiss (Armstrong and others, this volume) show that much of the metamorphic rock of the Priest River complex represents pre-Belt basement with an age in excess of 1,500 m.y.

In part of the complex west of the Newport fault, contact relationships between the high-grade metamorphic rocks and recognizable metamorphosed Belt-age rocks farther west are ambiguous. The problem is brought to focus in the Horseshoe Lake–Little Roundtop area, just west of the Silver Point pluton (Fig. 3). Mapping by Miller (1974c) shows that high-grade metamorphic rocks with steeply dipping, northeast-striking foliation are in contact with Prichard-like metasedimentary rocks containing metadiabase sills that strike WNW. The metasedimentary rocks dip toward the south and grade up-section into recognizable Belt formations. It is possible that this area contains an unconformity between basal Belt strata and their underlying crystalline basement. Similar unconformable relationships may occur farther north, structurally below and west of the Newport fault (Harms, 1982). If this unconformity hypothesis is correct, then the Horseshoe Lake locality represents one of the few places where the depositional base of the Belt sequence is exposed.

Intruding the pre-Belt and metamorphosed Belt rocks of the complex are several types of variably foliated and nonfoliated plutons (Fig. 3). Miller and Engels (1975) and Miller (1980) have divided the intrusive rocks of the Sandpoint 1° x 2° quadrangle into hornblende-biotite and two-mica suites. Representatives of both suites occur both in the core of the complex and in the upper plate of the Newport fault. Large plutons of the two-mica suite are present in the Mount Spokane area (Miller, 1982a,b) and west of the Newport fault. Elevated initial strontium ratios and the presence of Precambrian xenocrystic zircons in the two-mica plutons indicate that they were formed by melting of continental crust. Intrusive contacts are typically diffuse and gradational with pre-Belt metamorphic rocks, and intrusion appears to have been closely associated with high-grade metamorphism.

Plutonic rocks of the complex have been radiometrically dated by a variety of methods. Numerous K-Ar determinations were reported by Miller and Engels (1975), and additional K-Ar, Rb-Sr, and U-Pb work has been carried out (Reynolds and others, 1981; Armstrong and others, this volume). Further U-Pb dating is in progress, and results should soon appear (D. W. Hyndman, personal commun., 1985). These data collectively indicate that a pervasive regional thermal and/or cooling event affected most of the crystalline complex between 40 and 50 m.y. ago. Rocks outside the complex, including upper-plate rocks above the Newport fault, have escaped this Eocene thermal pulse or uplift-induced cooling. Along the gradational west margin of the complex, K-Ar cooling ages of granitic rocks systematically increase westward in correspondence with decreasing intensity of metamorphism (Miller and Engels, 1975). Ultimately, the cooling ages approach the probable emplacement ages of plutons (90-100 m.y.). The boundary between reset (Eocene) and non-reset (mid-Cretaceous) ages is sharp across structural boundaries of the complex like the Newport fault and Purcell trench. The 90- to 100-m.y. K-Ar dates from rocks bordering the west margin of the complex are concordant for muscovite, biotite, and hornblende, and therefore appear to give the true emplacement ages of the plutons (Miller and Engels, 1975).

Establishing original plutonic ages within the core of the complex is difficult because of the pervasiveness

of the Eocene thermal or cooling event. Thus, the concordant 41- to 54-m.y. K-Ar dates in the Selkirk igneous complex are not definitive (Miller, 1982a,b; Price and others, 1981), and they may not represent true emplacement ages. Recent work by Archibald and others (1984) strongly suggests that much of the Kaniksu batholith is of mid-Cretaceous age. Instructive in this regard are Rb-Sr and U-Pb zircon dates on the two-mica Mount Spokane quartz monzonite and Newman Lake orthogneiss northeast of Spokane (Fig. 3). Whereas K-Ar apparent ages for these rocks are Eocene, Rb-Sr whole-rock dates from 86 to 165 m.y. and U-Pb discordant zircon dates of 94 to 143 m.y. indicate the likelihood of a large composite pluton of mid-Cretaceous age (Armstrong and others, this volume). Tectonic aspects of the pluton are discussed below.

Accompanying the Eocene thermal or cooling event was the emplacement of equigranular, biotite-hornblende-bearing stocks of intermediate composition such as the Silver Point Quartz Monzonite and fine-grained quartz monzonite to granodiorite plutons in the Lone Mountain–Spirit Lake area (Fig. 3). Whole-rock Rb-Sr analyses (Reynolds and others, 1981; Armstrong and others, this volume) confirm the Eocene age for the Silver Point pluton (Miller and Engels, 1975). Limited Sr-isotopic initial-ratio data on the other stocks are not incompatible with Tertiary origins. However, Rhodes (1986) cites isotopic evidence that the intrusions are no younger than about 65 m.y.

The only post-Paleozoic supracrustal rocks associated with the Priest River crystalline complex are the Sandpoint Conglomerate north of Sandpoint, Idaho, and the Tiger Formation and Pend Oreille Andesite west and northwest of Newport (Fig. 3). The Sandpoint Conglomerate crops out sparingly along the Purcell trench, where it is moderately tilted to the east. Harrison and others (1972) considered this continental arkosic rock to be of Cretaceous age, but an Eocene age is not precluded.

The Tiger Formation, which crops out along the west margin of the Newport upper plate, is Eocene as dated by palynomorphs (Harms, 1982) and consists of coarse clastic deposits with minor lignite horizons (Gager, 1982). The formation is tilted 15° to 30° westward. The Tiger Formation overlies flows and flow breccias of the Pend Oreille Andesite dated at 50 to 51 m.y. (Pearson and Obradovich, 1977).

DIAGNOSTIC FEATURES OF THE PRIEST RIVER COMPLEX

The rocks described above contain evidence for a complicated sequence of plutonism, metamorphism, and intense deformation. This Mesozoic-Cenozoic history has resulted in formation of a metamorphic core complex similar to those elsewhere in the Cordillera. Diagnostic features of the Priest River metamorphic core complex are described below.

Metamorphic and Mylonitic Foliations

Many rocks within the complex are foliated to various degrees (Fig. 4). We recognize at least three different styles of foliation. The oldest foliation (S_1) is the steeply dipping, generally north-northeast- to northeast-striking foliation found in pendants and more extensive outcrops of pre-Belt metamorphic rocks through the northern and western parts of the Priest River crystalline complex. Lineation is not pronounced in this fabric. Foliation in places is discordant with the boundaries of the mid-Cretaceous plutons; in other places it is concordant, indicating that fabric guided plutonic emplacement (Miller, 1974b,c).

Another widespread, but generally shallow-dipping ($<45°$) foliation (S_2), with variably developed lineation, is common to a large part of the complex south of Newport. This foliation is generally crystalloblastic, defined by oriented minerals and compositional or gneissic layering. Granitic sills and metamorphic differentiates are conformable with this foliation, and their textures indicate that deformation was synchronous with metamorphism. Mylonitic textures are present in these rocks near Hauser and Newman Lakes (Weissenborn and Weis, 1976), but it is still unclear whether S_2 is truly mylonitic, or whether a crystalloblastic foliation has been overprinted by the mylonitic S_3 foliation described below. Rhodes (1986) considers much of the foliation mylonitic and part of a low-pressure amphibolite-facies metamorphism.

The third foliation (S_3) is highly mylonitic and occurs in footwall rocks fringing the Newport fault and the Purcell trench. Feldspar in these rocks displays brittle deformation, whereas quartz has deformed plastically or has recrystallized, giving rise to fabrics displaying notably more intergranular strain and disruption than is common in S_2 foliation.

An important distinction between S_2 and S_3 foliations is their differing directions of lineation. For S_2 the trend is S. 50°-60° W., whereas for S_3 it is N. 80°-90° E., except along the eastern prong of the Newport fault where it plunges westward (Fig. 4). The small stocks of latest Cretaceous or Eocene equigranular quartz monzonite to granodiorite (Fig. 3) intrusive into metamorphic rocks in the Lone Mountain–Spirit Lake area further tend to distinguish the two foliations. The plutons are discordant with respect to S_2, yet subtly and discontinuously contain the S_3 foliation.

Detachment Faults

The Priest River complex is bounded on several sides by low-angle normal faults of probable extensional origin and Eocene age. The Newport fault (Miller, 1971), the best exposed and documented of these detachment faults, is a spoon- or snowshovel-shaped surface that separates metamorphic, plutonic, and mylonitic rocks of the lower plate from less metamorphosed, relatively undeformed rocks of the upper plate (Figs. 3 and 4; Reynolds, 1980; Rehrig and others, 1982; Johnson, 1981; Harms, 1982). Upper-plate rocks were not affected by the pervasive deformational and thermal processes that affected the lower plate, as evidenced by unperturbed K-Ar ages of plutonic rocks (Miller and Engels, 1975). Harms (1982) has documented normal movement both east and west of the U-shaped fault trace. Upper-plate strata have been rotated by normal displacement on both east and west sides of the fault, forming a broad antiformal structure, the Snow Valley anticline. Mylonitic textures are common in crystalline rocks underlying the detachment surface (Miller, 1971; Harms, 1982). Although there is considerable controversy regarding the age and tectonic origin of these mylonites (for example, Rhodes and Hyndman, 1984), relatively thin mylonite zones and well-developed chloritic breccias above occur within the Silver Point pluton of Eocene age, thus establishing part of the ductile and brittle deformation as extensional and of Eocene or later age. The overall characteristics of the Newport fault, including (1) lower-plate mylonitic rocks, (2) chloritic breccia, (3) the low-angle normal displacement, and (4) highly tilted upper-plate rocks, are very similar to those of detachment faults in other metamorphic core complexes, particularly those described in the southwestern United States (Davis, 1980; Davis and others, 1980; Rehrig and Reynolds, 1980; Reynolds and Rehrig, 1980).

Although an outcrop of an actual detachment fault surface has not been found along the Purcell trench, there are many indications that one is present there. First, the trench separates high-grade rocks of the core of the complex from Belt rocks of lower grade to the east, a relationship that suggests relative downfaulting of the Belt rocks. Also, crystalline and mylonitic rocks of the complex are increasingly fractured, faulted, and brecciated up-section to the east, toward the trench (Nevin, 1966). At the town of Rathdrum (Fig. 3), we have found chloritic breccia and striated or grooved fracture surfaces of shallow dip, diagnostic of the footwall just below a detachment fault. That this structure dips toward the east is suggested by the gentle easterly dip of mylonitic foliation in core rocks bordering the trench. In all, the data indicate that the Purcell trench is occupied by an east-dipping, low-angle normal fault, the Purcell–Coeur d'Alene detachment fault. The east-dipping Belt rocks and Sandpoint Conglomerate east of the fault reflect tilting by listric and low-angle normal faulting similar to that in the upper plates of other metamorphic core complexes, although pre-tilting attitudes for the Belt rocks may not have been horizontal.

West of the Priest River complex, Cheney (1980) raised the possibility that the Jumpoff Joe fault in the Chewelah area (Miller and Clark, 1975) might be a detachment fault. Although Miller and Clark mapped the structure as a thrust, one strand of the fault cutting the Phillips Lake Granodiorite (Cretaceous) displays a wide zone of shearing and cataclasis similar to chloritic breccias along detachment faults. While it is possible that the Jumpoff Joe fault zone had an original thrust history, part of it may also be a detachment-like structure. One should note, however, that neither mylonitic nor isotopically reset rocks are found in the footwall of the structure as they are beneath the Newport and Purcell detachment faults.

Eocene Cooling in Lower-Plate Rocks

Rocks in the core of the complex invariably yield 40- to 50-m.y. cooling ages (Miller and Engels, 1975; Armstrong and others, this volume). This also applies to the nearby Okanogan and Kettle complexes (Fox and others, 1976; Orr and Cheney, this volume), and to the metamorphic core complexes of Nevada and Arizona, except that cooling ages south of the Snake River Plain are Oligocene-Miocene (20-30 m.y.).

Synchroneity of Lower-Plate Plutonism and Upper-Plate Volcanism

As in other core complexes, lower-plate plutonism and upper-plate volcanism were synchronous in the Priest River complex, even though outcrops of upper-plate volcanic rocks in the Priest River complex are relatively rare. The Pend Oreille Andesite (Eocene) and related Tiger Formation were deposited synchronously with emplacement of lower-plate plutonic rocks like the Silver Point pluton, stocks in the Lone Mountain–Spirit Lake area, and numerous north- to north-northeast-trending dikes. Upper-plate extrusive rocks of Eocene age are far more abundant in the areas of the Kettle and Okanogan core complexes (Orr and Cheney, this volume).

TECTONIC DEVELOPMENT

The Priest River crystalline complex occurs within an expansive region of Belt Supergroup rocks. The Rocky Mountain fold and thrust belt lies to the east,

and the Paleozoic continental shelf-slope break, defined by the abrupt transition from miogeosynclinal to eugeosynclinal rocks along the Columbia River, is west of the complex. The overall position of the complex is slightly inboard of the Paleozoic edge of North America. In contrast, the adjacent Kettle and Okanogan complexes occur within eugeosynclinal terranes and may not be underlain by North American Precambrian basement.

With respect to major tectonic features of Canada, the Priest River complex lies along the transition between the Kootenay arc and Canadian thrust belt. Across the Canadian border, the Priest River complex merges in a complex and ill-defined manner with crystalline gneissic rocks of the Mesozoic Nelson batholith and gneisses of the Trail and Valhalla complexes (Okulitch, 1984). The Kettle and Okanogan complexes appear to be southerly extensions of the Omineca crystalline terrane and Shuswap metamorphic core complexes of British Columbia, which lie west of the Canadian part of the Kootenay arc (Fig. 1).

The Priest River complex contains evidence for at least two major episodes of post-Paleozoic plutonism, metamorphism, and deformation: one during the Mesozoic, and the other in the Eocene. The spatial coincidence of Mesozoic and Tertiary magmatic and low-angle tectonic events is typical of Cordilleran core complexes and has led to much confusion in sorting out causes for their origin (DeWitt, 1980; Mattauer and others, 1983). The Mesozoic tectonic episode in the Priest River complex involved emplacement of extensive two-mica granitoids, deep-seated regional metamorphism, and ductile deformation. The foliation and compositional layering formed during this event may be largely represented by the gently dipping crystalloblastic gneisses through most of the complex, as emphasized by Rhodes and Hyndman (1984) and Rhodes (1986).

Much of this S_2 fabric is present within the Hauser Lake Gneiss and Newman Lake-Mount Spokane plutonic complex from Newport south to at least Interstate Highway 90. The fabric in the pre-Belt gneisses is the same as that in the mid-Cretaceous composite pluton comprising the Newman Lake orthogneiss and Mount Spokane quartz monzonite (Weissenborn and Weis, 1976; Miller, 1974a,b). This fabric also cuts obliquely across the intrusive contact in places (Miller, 1974d). These observations indicate that the S_2 fabric (1) was developed after intrusion, (2) is not a protomylonitic magmatic phenomenon, and, most important, (3) is not the original fabric of the pre-Belt basement. Therefore, it must be a superimposed metamorphic fabric of mid-Cretaceous or younger age.

As mentioned, the structural association between foliations S_2 and S_3 is difficult to assess without more field research. Development of the S_3 mylonitic foliation appears to be the result of the most recent deformation in the Priest River metamorphic core complex. Its close spatial association with the Newport fault and postulated Purcell–Coeur d'Alene fault suggests a close genetic relationship to regional detachment. Because both mylonite and detachment breccias of the Newport structure cut the Silver Point stock of Eocene age, the age of deformation is assumed to be Eocene or younger (Miller and Engels, 1975; Reynolds, 1980; Harms, 1982). By analogy, and because possible Eocene stocks are variably mylonitic near the east margin of the complex (for example, Lone Mountain), we are assuming Tertiary deformation also along the Purcell–Coeur d'Alene detachment zone. What is not known with certainty is whether this detachment-related mylonitic foliation at a deeper level correlates with any of the west-dipping mylonitic fabric seen in the core or west parts of the complex. The fine-grained quartz monzonite stocks west of Spirit Lake and Rathdrum discordantly cut this west-dipping foliation with its deep-seated, metamorphic character. Because of a probable late Cretaceous to Eocene age for these stocks, we assume a pre-Eocene age for much of the deformation. Alternatively, however, the stocks could have been intruded during the waning stages of Eocene mylonitization, thereby not precluding the presence of Eocene mylonitic foliation within the wide zone of west-dipping foliation. The sorting out of the early and later fabrics is a difficult task, but where this has been done in other core complexes, the Tertiary mylonite zones have been found intermixed with and superimposed upon earlier fabrics (Davis and others, 1983; Reynolds, 1982). In this sense, the Priest River complex cross section would more closely resemble the idealized metamorphic core complex section of Figure 2.

As with Tertiary deformation in metamorphic core complexes elsewhere, the mylonitic fabric in the Priest River complex reflects stretching parallel to lineation, accompanied by subvertical flattening. An extensional origin for such fabric in the Priest River complex is suggested by the fact that north-striking Eocene dike swarms and extensional faults are oriented perpendicular to the ENE to east-west trends of mylonitic lineation and transport direction on normal faults (Harms, 1982). Furthermore, the S_3 mylonites appear to be related to extensional processes in the detachment and upper-plate low-angle normal faults (Harms, 1982).

The Eocene events in the Priest River complex were part of the intense regional extension and intermediate to silicic magmatism that occurred about 50 to 40 m.y. ago through the entire northwestern United

States and southwestern Canada (Armstrong, 1974; Ewing, 1980; Price and others, 1981; Rehrig and others, 1982). The combination of tensional stress, widespread magmatic activity, and elevated temperatures resulted in relatively flat zones that accommodated crustal extension. Rehrig (1986) favors formation of the mylonitic zones by first-order elements of ductile crustal stretching, as did Harms (1982) and Price and others (1981). Reynolds, however, considers the mylonites formed along large, ductile normal faults by simple shear (Davis, 1983; Davis and others, 1983). Regardless of the mechanical model, the Priest River complex appears to be capped on the north and east sides by major zones of Eocene mylonite and detachment that separate high-grade, ductilely deformed rocks below from low-grade, brittlely extended rocks above.

The overall degree of ductile and brittle extension in the complex seems to increase southward. Harms (1982) described the Newport structure dying out well before it reaches the Canadian border. From the Sandpoint area northward, only thin discontinuous, east-dipping mylonite zones are found. To the south, there is progressive amplification of the mylonitic east flank of the complex. Outcrops of mylonite gneiss on Lone Mountain and along the shores of Lake Coeur d'Alene are most impressive. Furthermore, this southern segment contains the only known occurrences of ENE-trending foliation arches oriented parallel to lineation (Fig. 4), so typical of other Cordilleran metamorphic core complexes (Keith and others, 1980; Rehrig and Reynolds, 1980).

If the north end of the Priest River complex is ill-defined and diffuse, the south end, although partially covered, appears to be rather abrupt. Belt metasedimentary rocks are present immediately south of the mylonitic gneisses near Worley (Fig. 3) and define a probable southern boundary of the complex. This boundary nearly aligns with the westward projection of the St. Joe fault, an important element of the Lewis and Clark fault zone (Griggs, 1973).

We have referred above to similarities between the Priest River metamorphic core complex and metamorphic core complexes of the southwestern United States. Figure 5 is a schematic cross-sectional comparison between the Priest River complex through the Mount Spokane area and the Harcuvar metamorphic core complex in western Arizona. Key points in their likeness are (1) the low-angle east-dipping detachment zones and subjacent Tertiary mylonites, (2) a probably earlier, west-dipping, largely crystalloblastic foliation that decreases in intensity up structural section, and (3) a large, sill-like, mid-Cretaceous pluton with alaskitic phases that intrudes host rock of quartzofeldspathic migmatitic gneiss that has shallow-dipping foliation.

SUMMARY

Salient points of this paper are as follows:

(1) Basement gneisses of the Priest River crystalline complex are pre-Belt, with ages in excess of 1,500 m.y.

(2) The Newman Lake orthogneiss is Cretaceous and is part of the 90-to-100-m.y.-old major batholithic terrane that characterizes much of northeastern Washington.

(3) The Priest River terrane is described as a typical Cordilleran metamorphic core complex with a deformational history probably beginning in mid-Cretaceous time, associated with deep-seated plutonism (two-mica granites), metamorphism, and generation of a low-angle ductile fabric. Eocene tectonism has most recently affected the core complex by overprinting earlier fabrics with gently dipping mylonitic foliations and zones of low-angle detachment associated with regional crustal extension and a pulse of calc-alkaline magmatic activity. The specific details and degrees of shallow-dipping mylonitic fabric, resulting from either Mesozoic or Eocene deformational events, still must be determined, as they must for most other Cordilleran metamorphic core complexes.

In their regional setting, the metamorphic core complexes of northern Washington and Idaho generally occupy the transition zone from cratonic shelf and miogeosyncline on the east to eugeosynclinal and oceanic terranes on the west. Specifically, the Priest River complex represents part of North American basement that has been tectonically raised through overlying Proterozoic supracrustal cover. The uplift is localized just inboard of the Kootenay arc and appears to have disrupted basement structures like faults of the Lewis and Clark zone, which cannot be traced through crystalline rocks of the core complex (Fig. 1). Despite this disruption, the Lewis and Clark zone forms a perplexing regional map connection between the Priest River complex and the Bitterroot metamorphic core complex in eastern Idaho (Fig. 1). Recently, much of the plutonism and mylonitization in the Bitterroot complex has been ascribed to Eocene extension (Rehrig and others, 1982; Chase and others, 1983; Garmezy, 1983). Therefore, the implications of thick, eastward-dipping mylonite zones in both complexes, the time congruency of this deformation, and the crude tectonic(?) connection by the Lewis and Clark zone pose an interesting

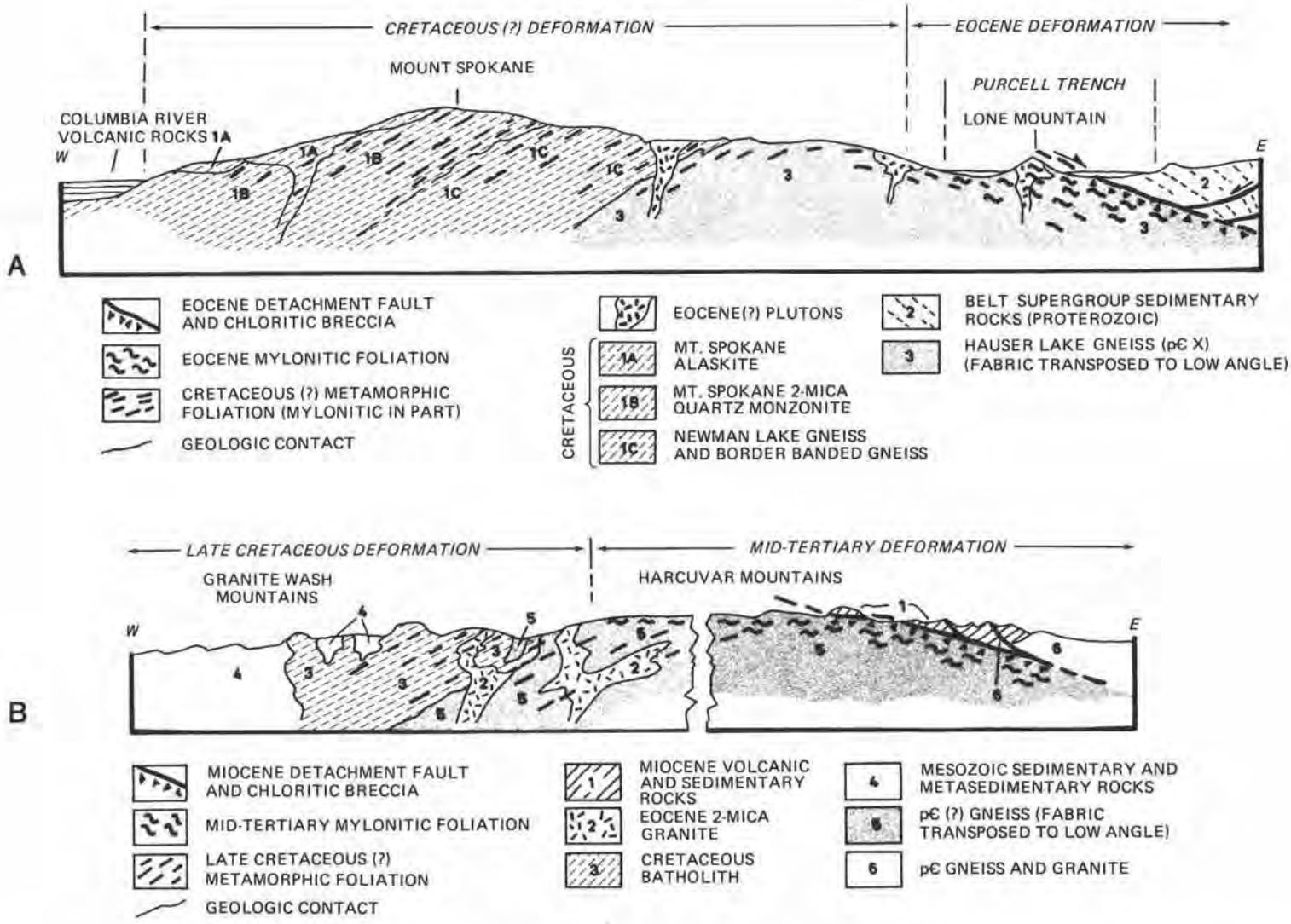

Figure 5.–Cross sections through the Priest River metamorphic core complex (A) and the Harcuvar metamorphic core complex, Arizona (B). Note the overall similarities in lithologies and structural configuration, even though Tertiary extensional deformation occurred during the Eocene in the Priest River complex, but during the Oligocene and Miocene in the Harcuvar complex. The Harcuvar cross section is taken from Rehrig and Reynolds (1980).

subject for future speculation and research. One appealing possibility is that the Lewis and Clark zone, originally a Precambrian structure, was reactivated as a strike-slip fault during Phanerozoic time. As interpreted by Sales (1968), it may have had considerable left-lateral offset during Laramide compressional tectonism. During Eocene extension, the zone apparently experienced right-lateral displacements (Harrison and others, 1974). The structure essentially connects the Priest River complex and the Bitterroot complex, and thus it may link extension along the Purcell–Coeur d'Alene detachment fault with that along the Bitterroot detachment fault. In this scheme, the Lewis and Clark zone was reactivated either as a tear fault separating different lobes of a regional, low-angle, ductile shear zone, or as a transform zone accommodating enhanced crustal stretching south of the zone.

Finally, the Priest River complex and neighboring metamorphic core complexes expose crystalline core zones that contain evidence for a pre-Tertiary history of profound plutonic and metamorphic orogeny. This thermal and structural mobility during the Mesozoic has been linked in general but direct ways to cratonic thrusting and supracrustal shortening to the east (Price, 1981; Scholten, 1982). Our present work gives some insight into the pervasiveness and severity of this deep-seated core-zone orogeny, even though details linking it mechanically to foreland shortening remain elusive. We have, however, emphasized here the overprint of a much later and less recognized extensional orogeny that seriously modified or destroyed the Mesozoic tectonic features.

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Rb-Sr AND U-Pb GEOCHRONOMETRY OF THE PRIEST RIVER METAMORPHIC COMPLEX–PRECAMBRIAN X BASEMENT AND ITS MESOZOIC-CENOZOIC PLUTONIC-METAMORPHIC OVERPRINT, NORTHEASTERN WASHINGTON AND NORTHERN IDAHO

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ABSTRACT

The Priest River metamorphic complex contains a variety of plutonic and metamorphic rocks. The oldest unit, the Hauser Lake Gneiss, is basement to Belt–Purcell sedimentary rocks that are at most 1.45 billion years old. The migmatitic Hauser Lake Gneiss is on the order of 2.0 billion years old by Rb-Sr and is crosscut by augen orthogneiss (also part of the Hauser Lake Gneiss) that is at least 1.44 to 1.58 billion years old by Rb-Sr and previously published U-Pb dates. Rb-Sr and Sr isotopic analyses of adjacent layers of migmatitic gneiss indicate that the migmatitic banding is of pre-Belt–Purcell age; the approximate time of that metamorphic event may be given by an upper-intercept U-Pb date of 1.67 billion years for zircons from migmatitic paragneiss. A deformed leucogneiss layer within migmatitic Hauser Lake Gneiss gives indeterminant Rb-Sr and U-Pb dates. While probably of the same general pre-Belt–Purcell age, the leucogneiss could be the result of Mesozoic anatexis of the Hauser Lake Gneiss. Gneisses at Davis Lake and southeast of Spokane in Washington and Idaho give scattered Rb-Sr whole-rock dates and are inferred to be pre-Belt–Purcell basement with a Phanerozoic plutonic-metamorphic overprint.

Muscovite granite (adamellite) of Tubbs Hill in Coeur d'Alene, Idaho, is a Cretaceous pluton (U-Pb date ~90-100 million years) with a very high $^{87}Sr/^{86}Sr$ initial ratio (~0.748) that contains xenoliths of muscovite pegmatite from pre-Belt–Purcell basement. Two-mica granite (adamellite) of Mount Spokane, Washington, and an adjacent foliated augen gneiss, the Newman Lake Gneiss, give identical, somewhat discordant Cretaceous U-Pb dates (94-143 million years), and they have similar initial $^{87}Sr/^{86}Sr$ ratios (~0.710). Single-sample whole-rock Rb-Sr dates on Rb-enriched pegmatite in the Mount Spokane pluton are Jurassic to Cretaceous—older dates may be real, or due to radiogenic initial Sr mobilized from basement rocks; Rb-Sr biotite and muscovite dates are Eocene. Both the Mount Spokane and Newman Lake bodies are of Cretaceous age and were variably deformed before Eocene resetting of mica dates.

The Silver Point Quartz Monzonite and associated aplite dikes give a whole-rock Rb-Sr isochron date of 39 ± 4 million years, which agrees with published K-Ar dates indicating an Eocene age. This gives a maximum age for major movement on the Newport fault.

In areas where published K-Ar dates for plutonic rocks are Eocene, the Rb-Sr dates for biotite and muscovite are also Eocene (average value 46 million years); this requires rapid cooling of the region at that time—probably due to uplift associated with tectonic denudation.

INTRODUCTION

Several interests provided the motive for this geochronometric study and the related structural reconnaissance of plutonic rocks in northeastern Washington and northern Idaho. Knowledge of the age and distribution of Precambrian basement in the hinterland of the Cordilleran fold-and-thrust belt is essential for reconstruction of the pre-orogenic configuration of North America and for disassembly of the Cordilleran tectonic collage. Over the past two decades considerable attention has been directed toward core complexes and their tectonic significance (Coney, 1979; Crittenden and others, 1980; Armstrong, 1982a). The Priest River metamorphic complex has characteristics common to many core complexes—a structural culmination exposing Precambrian basement, Mesozoic and Cenozoic granitic rocks, pervasive Cenozoic resetting of K-Ar and Rb-Sr mineral dates, regionally extensive mylonitic fabric, and Cenozoic denudation (detachment) faults.

This paper will focus on geochronometry. A companion paper (Rehrig and others, this volume) presents geologic and tectonic observations and inferences. Recent contributions to the regional interpretation of northeastern Washington and nearby areas in Idaho include papers by Cheney (1980) and Rhodes and Hyndman (1984). All are indebted to P. L. Weis, A. E. Weissenborn, A. B. Griggs, and F. K. Miller of the U.S. Geological Survey and to numerous graduate theses for geologic maps (cited in Rehrig and others, this volume) on which our new studies are based.

Our work began with reconnaissance structural studies in 1976 and was later enlarged with Rb-Sr and U-Pb dating of critical or problematic units. Preliminary reports were given by Rehrig and Reynolds (1981), Reynolds and others (1981), and Rehrig and others (1982). The localities from which samples were collected are shown on a generalized geologic map (Fig. 1); sample descriptions and analytical results are presented in Appendix 1 and Appendix 2. The laboratory techniques in use at the University of British Columbia are outlined in Appendix 3.

THE AGE OF BELT–PURCELL ROCKS AND PREVIOUS DATING OF PRE-BELT–PURCELL BASEMENT

The age of Belt–Purcell Supergroup rocks has been examined in a number of studies and reviewed by Harrison (1972) and McMechan and Price (1982); one recent definitive date has been provided by Zartman and others (1982). The current view is that most Belt–Purcell strata are 1,450 to 1,300 m.y. old. Lead isotopes in stratiform and remobilized ores have been interpreted to indicate an age of 1,400 to 1,430 m.y. (Long and others, 1960; Zartman and Stacey, 1971; LeCouteur, 1979). Strontium isotopic analyses of sedimentary rocks suggested a range in age from 1,300 to 850 m.y. (Obradovich and Peterman, 1968) with large hiatuses, but the younger dates must be viewed skeptically. Crosscutting veins near Coeur d'Alene contain uraninite dated at 1,190 m.y. old by U-Pb (Eckelmann and Kulp, 1957), and a muscovite-granite intrusive into Belt–Purcell rocks in British Columbia has been dated at 1,319 ± 25 m.y. (minimum age) by Rb-Sr on muscovite (Ryan and Blenkinsop, 1971) and 1,340 ± 4 m.y. by U-Pb dating of zircon (J. K. Mortensen, Univ. of British Columbia, unpublished data). This date is comparable to the oldest K-Ar dates reported for Purcell sills by Hunt (1962). Detrital zircons separated from Belt–Purcell sedimentary rocks in Idaho have given ages "greater than 1413 Ma" (Grauert and Hofmann, 1973b), approximately 1,467 m.y. (Reid and others, 1973), and 1,800 to 2,100 m.y. (Grauert and Hofmann, 1973a). Paleomagnetism suggests that all Belt–Purcell strata are more than 1,200 m.y. old (Evans and others, 1975; Elston and Bressler, 1980).

U-Pb dates reported for orthogneisses by Reid and others (1970, 1973) contradict some results already cited. Augen gneiss near Elk City, Idaho, intrusive into metasedimentary rocks correlated with Belt–Purcell rocks, is 1,501 m.y. old, and younger and older orthogneiss intruding Belt–Purcell metasedimentary rocks in the St. Joe area, Idaho, both contain zircon older than 1,760 m.y. (Reid and others, 1973, 1981). Inherited zircon was suspected of giving the old dates near St. Joe. Taken at face value, the Elk City date would dictate that the oldest Belt–Purcell deposits are at least 50 m.y. older than the generally accepted age of 1,450 m.y.

Evans and Fischer (1986) interpreted discordant U-Pb zircon dates from the same Elk City augen gneiss to indicate an age of only 1,370 ± 10 m.y., indicative of an episode of plutonism affecting Belt–Purcell strata after their deposition. These authors correlated the augen gneiss with similar rocks in the Salmon River Mountains that have similar U-Pb discordance (Evans, 1981). Their conclusions are based on a particular discordance model, and other interpretations, involving multiple episodes of Pb loss from pre-Belt–Purcell zircons, are possible.

Higher grade metamorphic rocks in northeastern Washington and northern Idaho have been variously interpreted as metamorphosed Belt–Purcell rocks or pre-Belt–Purcell basement. Weissenborn and Weis (1976) considered high-grade rocks around Spokane

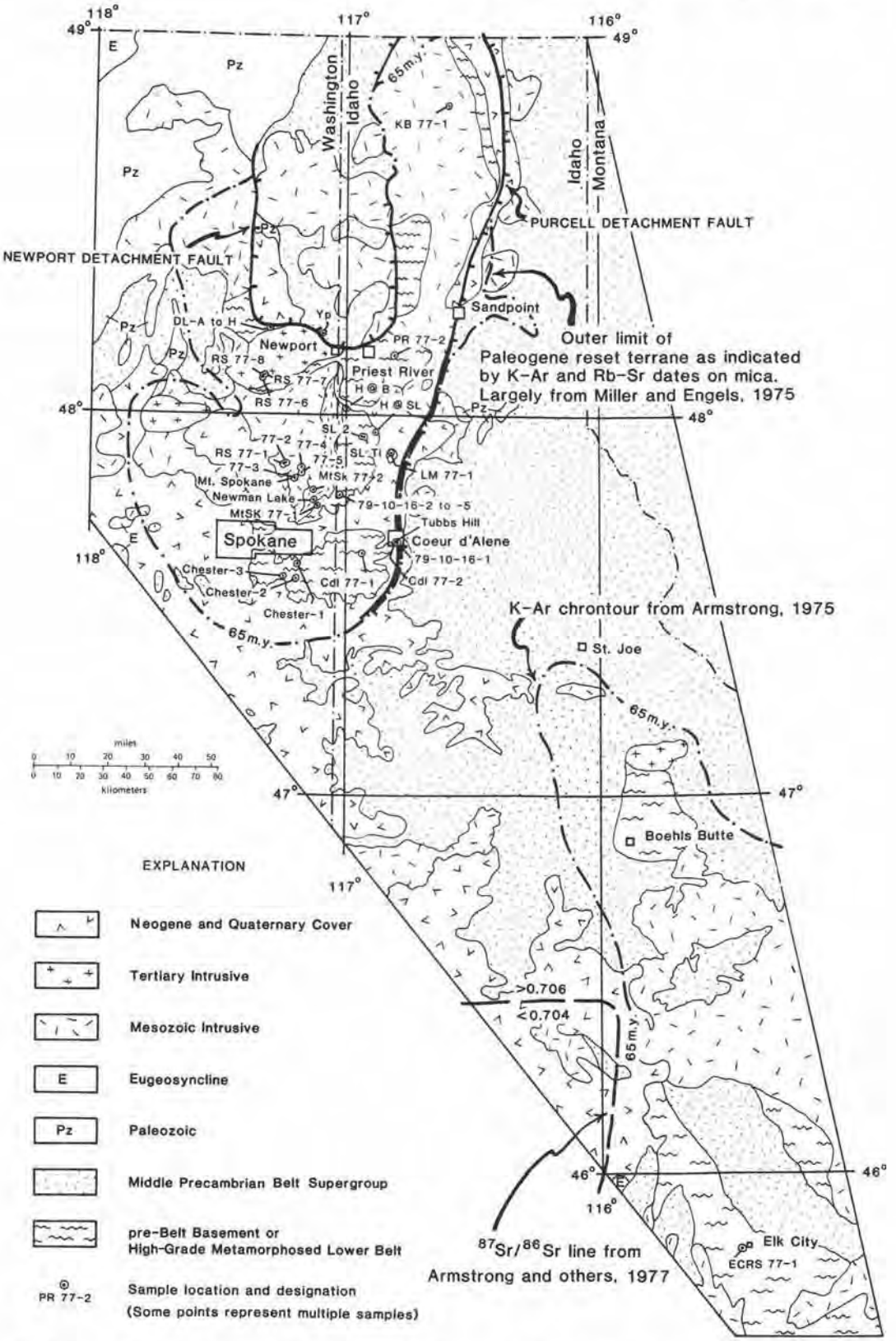

Figure 1.–Generalized geologic map of northeastern Washington and northern Idaho, showing locations for the analyzed samples. (After Griggs, 1973; Miller and Engels, 1975; Bond and others, 1978; Hietanen, 1984; and Rehrig and others, this volume.)

(Hauser Lake Gneiss of Weis, 1968) to be pre-Belt–Purcell or metamorphosed Belt–Purcell, whereas Griggs (1973) showed them as Prichard Formation (lowermost Belt–Purcell Supergroup) undivided. At the same time, Clark (1973) described as pre-Belt–Purcell an orthogneiss near Priest River, Idaho, that gave a U-Pb zircon date (estimated upper intercept) of 1,521 m.y.; that orthogneiss intrudes the same rocks studied by Weissenborn, Weis, and Griggs. All these metamorphic rocks were shown as probable pre-Belt–Purcell by Armstrong (1975); Harrison and others (1980) found that assignment acceptable. Evans and Fischer (1986) reported a U-Pb zircon upper-intercept age of 1,576 ± 13 m.y. for the same augen gneiss studied by Clark (1973).

High-grade metamorphic rocks of the Boehls Butte area, northwest of the Idaho batholith, were mapped by Hietanen (1963) as schist of the Prichard Formation, but Reid and others (1973) thought them to be pre-Belt–Purcell, because of greater deformation and zircon U-Pb dates of more than 1,635 m.y. and more than 1,600 m.y. These rocks were shown by Armstrong (1975) as probable pre-Belt–Purcell, and that interpretation was accepted by Hietanen (1984).

A single zircon analysis for pre-Belt–Purcell Marble Creek Gneiss of the St. Joe area, reported by Reid and others (1973, 1981) as indicating an age of 2,500 m.y., cannot be used because of its extreme lead loss (>90%) and strange discordance (either recent Pb loss or radiogenic Pb impossibly rich in ^{207}Pb). This date is likely an artifact of a troublesome analysis or incorrect choice of blank or common Pb.

Farther east, near Missoula, Montana, Chase and others (1978) and Bickford and others (1981) observed a xenocryst zircon component in Idaho batholith rocks that indicates basement or detrital Belt–Purcell zircons 1,700 to 2,350 m.y. old. North of latitude 49°, the older zircon component in Kaniksu batholith rocks and nearby gneisses is about 1,580 m.y. old (Archibald and others, 1984). With the exception of the Elk City and St. Joe results of Reid and others (1973, 1981), the reviewed data point to a pre-Belt–Purcell basement age of 1,521 to 1,635 m.y. in northeastern Washington and northern Idaho, and 1,700 to 2,350 m.y. in northern Montana. This is compatible with the already cited age of 1,450 m.y. for the beginning of Belt–Purcell deposition.

For comparison with the Rb-Sr results of our study, a large number of Belt–Purcell Sr isotopic analyses are plotted in Figure 2. A reference isochron for 1,450 m.y., the approximate time of initiation of Belt–Purcell deposition, is shown on Figure 2 and on all subsequent Rb-Sr isochron diagrams. On Figure 2, more highly metamorphosed samples are distinguished from those described as sedimentary rocks.

The pattern observed in Figure 2 is one expected for a suite of rocks that were coeval, with homogeneous initial ^{87}Sr/^{86}Sr ratios, and later subject to redistribution of Rb and Sr. In general, the collection of points is rotated clockwise about its centroid, and the scatter about an average regression line is increased. No samples with high Rb/Sr ratios lie above the reference line; they show a general loss of radiogenic Sr, or Rb gain. Some samples with low Rb/Sr ratios do shift to the left of the reference line by gain of radiogenic Sr or Rb loss—complementary to the changes in high-Rb/Sr samples. Metamorphosed rocks are most affected, and data points from these rocks may lie far to the right of the reference isochron. We have produced exactly these patterns on computer models of Rb-Sr isochron resetting processes.

Belt–Purcell rocks are distinctive in their relatively low Sr contents (10-80 ppm), moderately high Rb contents (100-300 ppm), and high and widely ranging Rb/Sr ratios (Obradovich and Peterman, 1968). The one Prichard Formation sample we analyzed plots near the center of rotation of the amassed Belt–Purcell data (Fig. 2) and is fairly typical. On the basis of Rb and Sr concentrations alone, a metamorphosed-Belt–Purcell-sediment origin for the schist and gneiss samples that we have analyzed is highly unlikely. Contrasts in metamorphic grade and isotopic age further support the view that we are dealing with pre-Belt–Purcell basement.

The average isochron calculated for Belt–Purcell rocks today is on the order of 1,100 m.y., somewhat younger than the likely age of deposition (1,300-1,450 m.y.), but not greatly out of line with that age. The result demonstrated by the abundant data of Figure 2 is probably typical of what can be expected in Rb-Sr dating of older sedimentary or metasedimentary rock suites.

Hauser Lake Gneiss

Sillimanite-grade migmatitic gneiss and schist northeast of Spokane were collected at several localities for Rb-Sr analyses, and two samples, one a migmatitic schist and the other a strongly lineated and foliated concordant leucogneiss layer, were processed for zircon. The Rb-Sr data are presented in Appendix 1 and Figure 3. The analyzed rocks are consistently more radiogenic than Belt–Purcell rocks. A 13-sample-composite model date (Cameron and others, 1981) of 2,053 m.y. (Table 1) is good evidence of a pre-Belt–Purcell age. Further evidence is provided by several zircon fractions from a composite sample of migma-

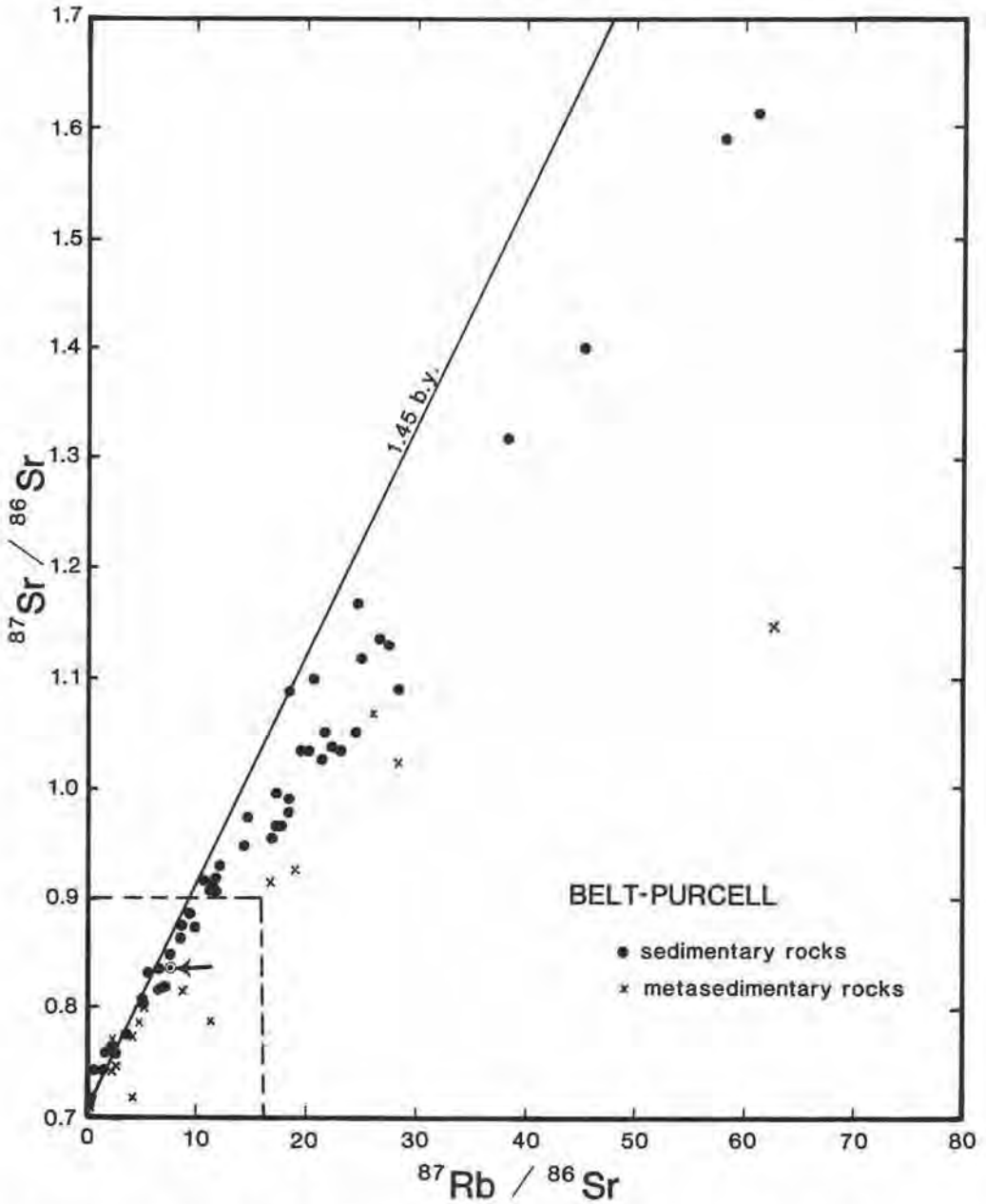

Figure 2.–Rb-Sr analyses for the Belt–Purcell Supergroup from a variety of published and unpublished sources (Obradovich and Peterman, 1968; Ryan, 1973; Fleck and others [listed in Armstrong, 1976]; Otto, 1978; Ryan and Blenkinsop, 1971). The one Belt–Purcell sample analyzed in this study, from the Prichard Formation, is shown by a small dot and indicated by an arrow. The 1.45-b.y. reference isochron is a consensus value for the time Belt–Purcell deposition began. The dashed rectangle is the area shown on all other Rb-Sr diagrams in this paper.

titic schist collected at Hauser Lake (Appendix 2, Fig. 4). Good analyses of three finer size fractions and a poorer analysis of a coarse, hand-picked fraction of a very heterogeneous zircon population define a discordia line with concordia intercepts of 1,668 ± 32 and 70 ± 17 m.y. Omission of the poorer and most discordant analysis does not change that result. The approximate 1,670-m.y. age indicated by these discordant zircons is a minimum for their creation because of multiple and protracted Mesozoic-Cenozoic lead-loss episodes. The result is similar to the previously cited zircon dates from the Boehls Butte area.

Augen gneiss that intrudes migmatitic Hauser Lake Gneiss near Priest River has a single-sample whole-rock Rb-Sr date of 1,440 m.y. (Appendix 1, Table 1), which is somewhat younger than the zircon

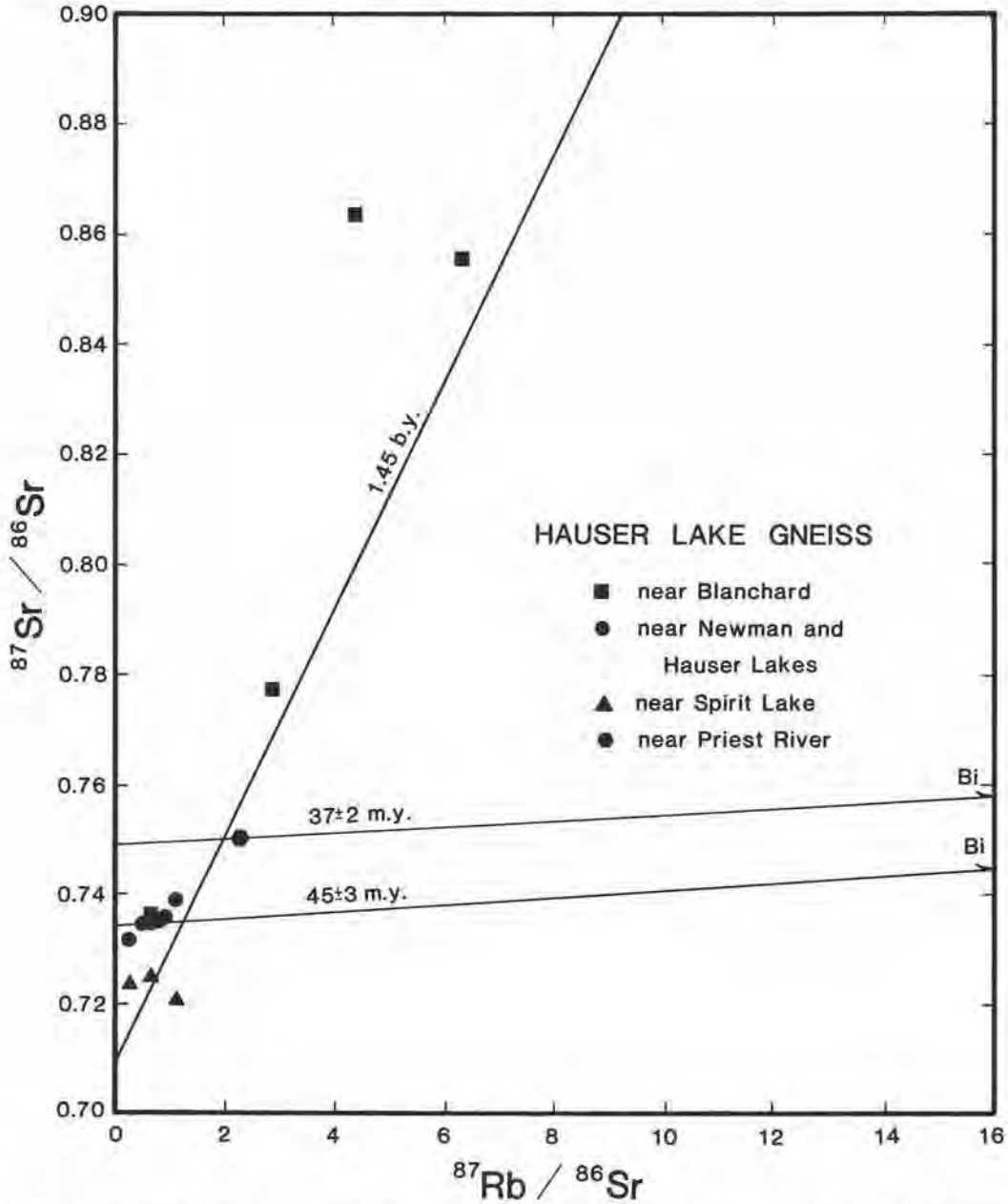

Figure 3.–Rb-Sr isochron plot for samples of Hauser Lake Gneiss. For the two whole-rock biotite + isochrons the biotite lies outside the diagram. Note the change in scale from Figure 2. Whole-rock symbols differ according to sample area.

dates of 1,521 m.y. reported by Clark (1973) and 1,576 ± 13 m.y. reported by Evans and Fischer (1986). In this discussion, both the migmatitic gneiss and the younger augen gneiss are included in the pre-Belt–Purcell Hauser Lake Gneiss.

If the migmatitic Hauser Lake Gneiss zircon contains a detrital component, then its upper-intercept age might be greater than the age of the protolith of the gneiss. Only the very radiogenic Sr and the augen-gneiss igneous-zircon dates make the case for a pre-Belt–Purcell age convincing. An ancient age for the migmatitic banding is indicated by the large layer-to-layer variability in Sr isotopic composition. A major metamorphic event to create that banding may be dated by the 1,670 m.y. zircon upper intercept.

Zircon from a strongly deformed leucogneiss sheet in Hauser Lake Gneiss near Blanchard (Table 1, Appendix 2, Fig. 4) gives younger Pb/Pb (~1200 m.y.) and upper intercept dates (1,387 +465/-371 m.y.). The calculated lower intercept is 100 +156/-200 m.y. A range of upper-intercept ages from 1,250 to 1,600 m.y. is possible if various lower intercepts are assumed.

Table 1.–Rb-Sr and U-Pb dates

Sample and/or suite	Type of Calculation	Date (m.y.±1σ)	$^{87}Sr/^{86}Sr$ initial ratio
	Precambrian rocks		
Hauser Lake Gneiss	13-whole rock, bulk-Earth initial	2,053	0.7021
MtSk 77-1, Hauser Lake Gneiss	Whole-rock + biotite	45±3	0.7340
79-10-16, Hauser Lake migmatitic gneiss	U-Pb, zircon, upper intercept	1,668±32	
	U-Pb, zircon, lower intercept	70±17	
H@B Gn, Hauser Lake leucogneiss	U-Pb, zircon, upper intercept	1,387+465 -371	
	U-Pb, zircon, lower intercept	100+156 -200	
PR 77-2, augen gneiss near Priest River	Single whole-rock, bulk-Earth initial	1,440	0.7029
PR 77-2, augen gneiss near Priest River	Whole-rock + biotite	37±2	0.749
DL-C, gneiss at Davis Lake	Single whole-rock, bulk-Earth initial	1,450	0.7029
Chester-3E, gneiss SE of Spokane	Single whole-rock, bulk-Earth initial	1,600	0.7028
Chester-2 PEG, pegmatite SE of Spokane	Whole-rock + muscovite	54±3	0.7385
Cdl 77-1, gneiss SE of Spokane	Whole-rock + biotite	43±2	0.7150
ECRS 77-1, gneiss near Elk City	Single whole-rock bulk-Earth initial	1,540	0.7028
Cdl 77-2, Tubbs Hill pegmatite	Single whole-rock bulk-Earth initial	2,280	0.7019
Cdl 77-2, Tubbs Hill pegmatite	Whole-rock + muscovite	487±50	0.767

Sample and/or suite	Type of Calculation	Date (m.y.±1σ)	$^{87}Sr/^{86}Sr$ initial ratio
	Mesozoic and Cenozoic Granitic Rocks		
Tubbs Hill, granite	2-whole-rock isochron	135±15	0.7458 ±0.0006
Tubbs Hill, granite	U-Pb zircon dates	93, 102	
RS 77-1, granite Mt. Spokane pluton	Whole-rock + biotite	51±2	0.7113
RS 77-1,2,3, granite + pegmatite, Mt. Spokane pluton	4-whole-rock isochron	84±4	0.7108 ±0.0001
RS 77-2, pegmatite Mt. Spokane pluton	Whole-rock + muscovite	47±16	0.722
RS 77-4, pegmatitic granite, Mt. Spokane pluton	Whole-rock assumed initial	159±3	0.711
RS 77-5, pegmatite, Mt. Spokane pluton	Whole-rock, assumed initial	113±2	0.711
RS 77-5, pegmatite, Mt. Spokane pluton	Whole-rock + muscovite	40±6	0.726
Mount Spokane	U-Pb zircon dates	94 to 143	
MtSk 77-2, Newman Lake Gneiss	Whole-rock + biotite	45±2	0.7119
Newman Lake Gneiss	U-Pb zircon dates	97, 139	
RS 77-6,7,8, Silver Point granite	3-whole-rock isochron	39±4	0.7063 ±0.0002
RS 77-6, Silver Point granite	Whole-rock + biotite	46±2	0.7064
KB 77-1, Kaniksu batholith	Whole-rock + biotite	49±2	0.7092

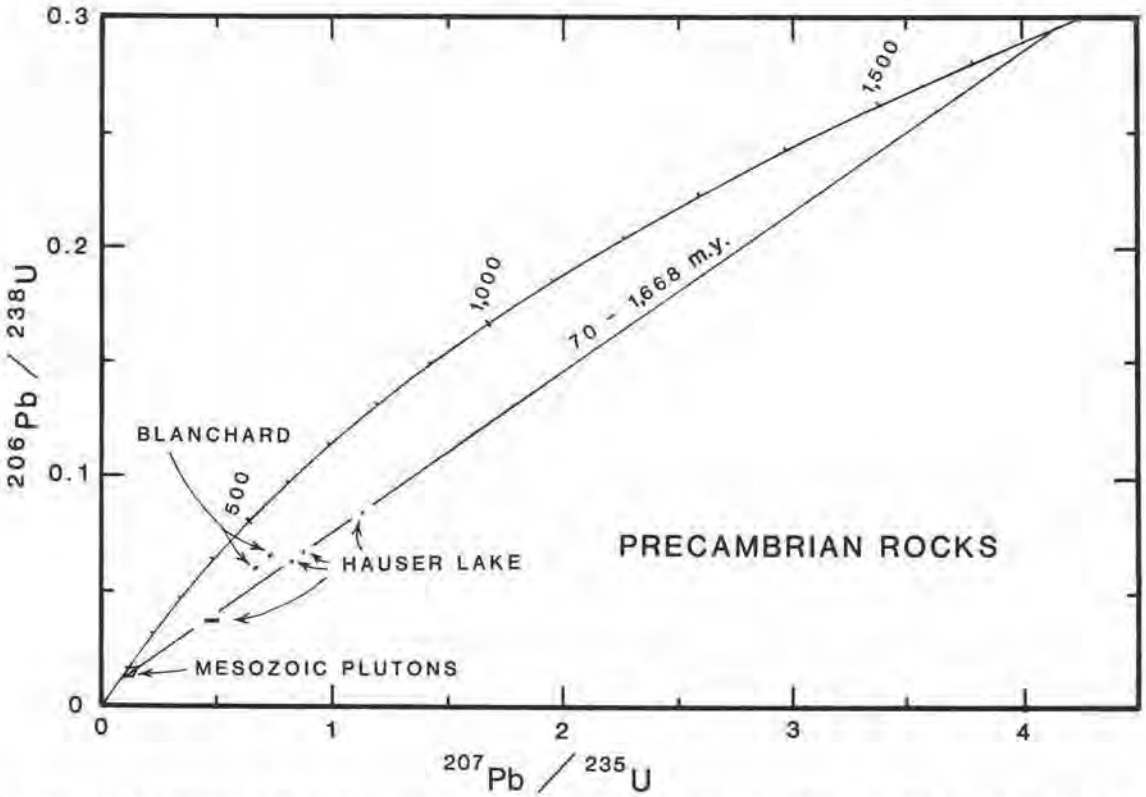

Figure 4.–U-Pb concordia plot for zircons of Hauser Lake Gneiss. A calculated discordia line (1,668 ± 32 m.y. to 70 ± 17 m.y.) for the composite migmatitic Hauser Lake Gneiss is shown. Because of protracted and multi-episodic Pb loss, the upper intercept is probably only a minimum age for a surviving older zircon component in these rocks, whose Rb-Sr data suggest an age of at least 2,000 m.y. The zircon analyses for a leucogneiss layer in the Hauser Lake Gneiss near Blanchard are shown, but they do not define a precise discordia line and are too discordant to be of use in inferring a true age. Their host rock is probably Precambrian, pre-Belt–Purcell in age, but xenocrystic zircon in a Mesozoic anatectic melt might give much the same appearance in this plot.

Sr of the leucogneiss layer is much less radiogenic than Sr in enclosing schists; values for the analysis lie above the 1,450-m.y. reference isochron on Figure 3, but the sample has too low a Rb/Sr ratio for a meaningful model age to be calculated. For this sample, a post-Belt–Purcell age, even Cretaceous, is possible. We treat it here as part of the Hauser Lake Gneiss. If Cretaceous, it would be of anatectic origin. Its age must remain an open question.

Gneiss at Davis Lake

High-grade mylonitic gneiss and deformed pegmatite sampled in a large roadcut near Davis Lake, north-northwest of Spokane, shows very scattered Sr isotopic compositions (Appendix 1, Fig. 5). These gneisses are presumably hybrid, composed of gneiss about 1,450 m. y. old (Table 1) injected by Mesozoic or younger pegmatitic material. A maximum age for the younger pegmatitic component is indicated by the 215-m.y. isochron drawn on Figure 5. Any initial ratio higher than 0.706 would give a younger age for the pegmatitic samples, and this is likely the case. Taking an initial ratio similar to that of nearby Mesozoic granites, 0.710, gives an age of 157 m.y.; a somewhat higher but still reasonable initial ratio could bring the pegmatite age down to 100 m.y. The older component is, on the basis of a Rb-Sr date of about 1,450 m.y., not necessarily pre-Belt–Purcell basement, but we prefer that interpretation because of the contrast of these rocks with much less metamorphosed Belt–Purcell rocks nearby. We would expect equally metamorphosed Belt–Purcell rocks to give consistently younger model whole-rock ages than many of the points shown in Figure 5. In Figure 2, none of the analyses that plot above the centroid of the Belt–Purcell data set lie significantly above the 1,450 m.y. reference isochron, whereas in Figure 5 the most radiogenic samples of each suite plot close to or above the same reference isochron.

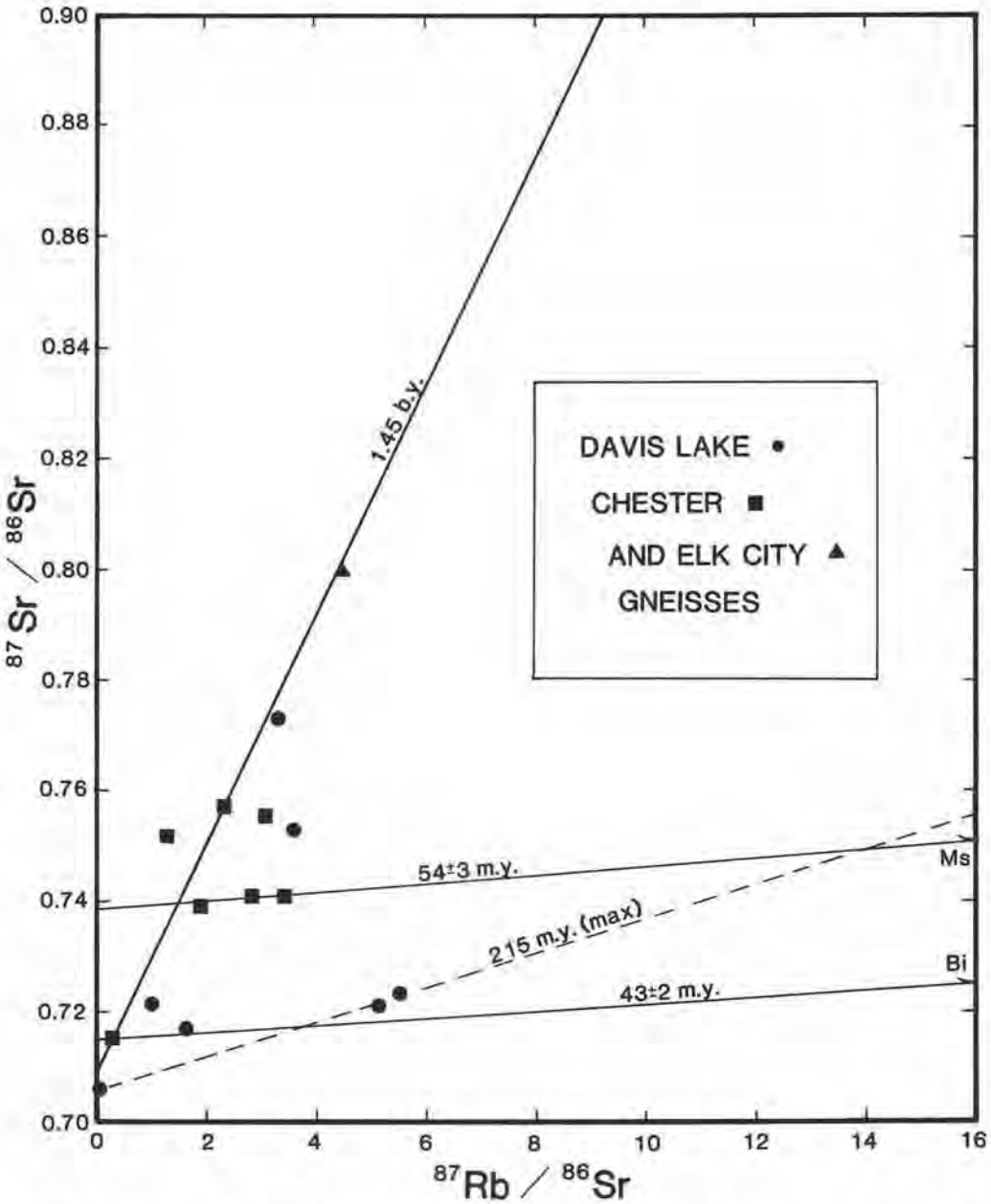

Figure 5.–Rb-Sr isochron plot for gneiss and pegmatite samples from Davis Lake roadcut and areas southeast of Spokane, Washington. Mineral analyses for the whole-rock + mica isochrons lie outside the area shown. The scattered whole-rock points indicate open-system behavior and the hybrid nature of the analyzed samples. Two muscovite-bearing pegmatite samples from Davis Lake which plot farthest to the right of the 1.45-b.y. isochron must be largely Mesozoic. They cannot be older than the 215-m.y. reference isochron shown with a dashed line.

Gneisses Southeast of Spokane

A large area of metamorphic rocks lies south of the Spokane River between Spokane, Washington, and Coeur d'Alene, Idaho. The rocks are separated from Hauser Lake Gneiss only by the Spokane River valley and are likely equivalent, but are discussed separately here. Weis (1968) reported Pb-alpha dates of 1,150 and 1,120 m.y. for these rocks, and if those zircons were anywhere near as discordant as the ones we have analyzed isotopically, then their discordia/concordia upper-intercept dates must lie well above 1,500 m.y. Our few isotopic analyses for these rocks scatter about the 1,450-m.y. reference isochron. One feldspar-rich layer, which has probably been a ^{87}Sr sink, gives an unrealistically old model age to which no special significance should be attached. The most radiogenic typical gneiss sample gives a model date of 1,600 m.y.

Pre-Belt–Purcell basement with later scrambled and reset Sr is our preferred interpretation for this suite, but the local isotopic evidence is equivocal.

Augen gneiss from the Elk City area gave one whole-rock Rb-Sr date of 1,540 m.y. (Table 1), in good agreement with the zircon date of 1,501 m.y. reported for those rocks (Reid and others, 1970), but older than the 1,370 m.y. zircon date of Evans and Fischer (1986). Our experience in several studies is that massive augen gneiss and orthogneiss in general tend to preserve Rb-Sr ages of crystallization under the same conditions where paragneiss Rb-Sr data become reset, discordant, and scrambled, and where U-Pb dates are reduced by loss of radiogenic Pb.

BASEMENT SUMMARY

Our new data, together with those of previous studies, reinforce the hypothesis that the large area of higher grade metamorphic rocks near Spokane is pre-Belt–Purcell basement, and not metamorphosed Belt–Purcell rock. The area where basement is exposed is a major structural culmination on the west side of the Belt–Purcell basin. The occurrence of Precambrian X (2,500-1,600 m.y.) rocks in this region is in accord with basement-age patterns of the North American craton (Armstrong, 1975; Daniel and Berg, 1981) and restricts Phanerozoic continental accretion to a fraction of the width of Washington State at 47° to 49° N. latitude.

TUBBS HILL

An isolated hill in the town of Coeur d'Alene is composed of variably foliated and lineated garnet-biotite-muscovite leucogranite containing deformed bodies of coarse biotite-muscovite pegmatite. Nearby exposures of Belt–Purcell argillite and siltite (Prichard Formation) are only weakly metamorphosed (small biotite porphyroblasts, but no secondary fabric other than slaty cleavage) and lack injections of leucogranite that would allow inference of an age relationship. The Coeur d'Alene fault (Griggs, 1973), thought to be an Eocene detachment structure (Rehrig and Reynolds, 1981; Rhodes and Hyndman, 1984; Rehrig and others, this volume), separates the plutonic rocks of Tubbs Hill from Belt–Purcell rocks, so that an age range from pre-Belt–Purcell to Cenozoic is allowed by field observations. The Tubbs Hill pluton lies on the east boundary of the metamorphic terrane where mineral dates were reset during the Eocene (Fig. 1). Our first Rb-Sr analyses of deformed pegmatite and muscovite hinted at an old age. The whole-rock date, assuming a bulk-Earth initial $^{87}Sr/^{86}Sr$ ratio, is 2,280 m.y. (Table 1); the whole-rock + muscovite date (487 ± 50 m.y.; Table 1) clearly indicated a pre-Mesozoic age. A premature conclusion was that the entire leucogranite body was basement, somewhat reset during Mesozoic-Cenozoic time. Two whole-rock analyses of the leucogranite itself only added confusion by failing to line up into an isochron. Rather, the three available analyses outline a triangle (Fig. 6) with two sides of negative slope. The line of positive slope between the two leucogranite samples gave a Mesozoic date, with large uncertainty because of the small difference in Rb/Sr ratios and the expectable isotopic heterogeneity of such radiogenic and obviously discordant samples.

Zircon yield from about 40 kg of leucogneiss was disappointing, less than 1 mg, but we ventured to analyze a hand-picked fraction of clear euhedral grains and were successful. The U-Pb dates (Appendix 2, Fig. 7) of 93 and 102 m.y. indicate a mid-Cretaceous age, with only a trace of older zircon present. The age is equal to that of many other plutons in the region (Miller and Engels, 1975), including the Kaniksu batholith, which was dated as 94 m.y. old by U-Pb at a locality just north of the international boundary by Archibald and others (1984). A Cretaceous age is compatible with the slope defined by the two whole-rock Sr analyses of leucogranite, but is obviously discordant with the earlier results on deformed pegmatite. Our best rationalization is that the pegmatite is a xenolith of basement which is the country rock of the Tubbs Hill pluton caught up in the much younger leucogranite magma.

Noteworthy is the very radiogenic initial $^{87}Sr/^{86}Sr$ ratio of the leucogranite, approximately 0.748 at 100 m.y. ago, a characteristic shared with other mid-Cretaceous granites of the Purcell anticlinorium (Armstrong, 1983). This initial ratio indicates that magma incorporated large amounts of old basement rock.

MOUNT SPOKANE PLUTON AND NEWMAN LAKE GNEISS

The mountains northeast of Spokane are underlain by granitic rocks, mostly medium-grained two-mica granite, that are massive to weakly foliated and discordantly cut by pegmatite and aplite dikes. K-Ar dates for mica are all Eocene, but thought to be reset after Cretaceous emplacement of the pluton (Miller and Engels, 1975). Between these intrusive rocks and the high-grade Hauser Lake Gneiss is a foliated, tabular body of more mafic augen gneiss called the Newman Lake Gneiss (Weis, 1968). On Griggs' map (1973), this gneiss is shown as Prichard Formation porphyroblastic gneiss. The texture is mylonitic, with thin shear zones parallel to foliation. The large porphyroblasts are distinctive and, together with the foliated mylonitic

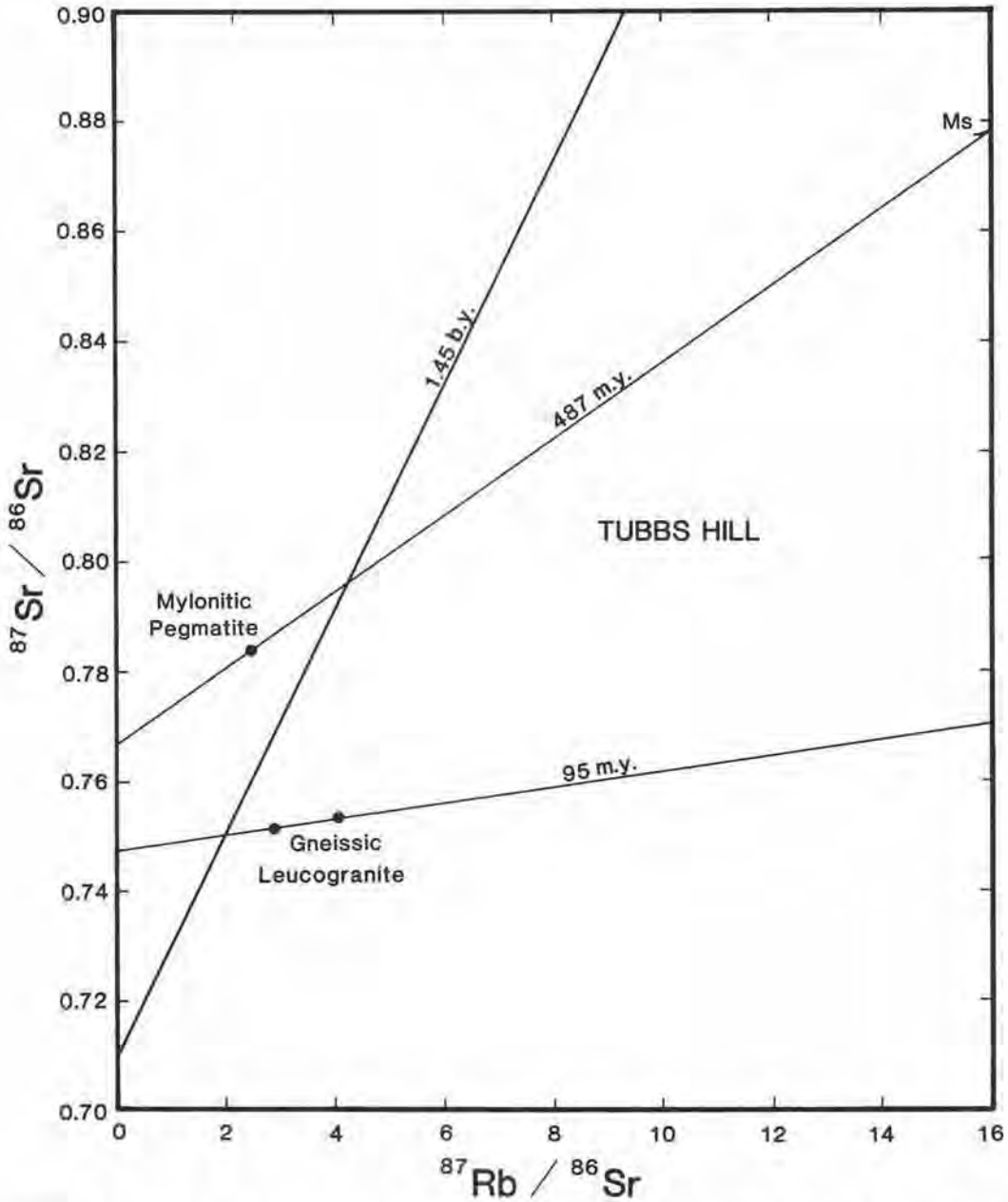

Figure 6.–Rb-Sr isochron plot for leucogranite and deformed pegmatite, Tubbs Hill, Idaho. Muscovite from the pegmatite lies on the 487-m.y. isochron outside the area shown. The 95-m.y. reference isochron is based on the ages of several nearby plutons dated by K-Ar and U-Pb (Miller and Engels, 1975; Archibald and others, 1984). The actual line connecting the two points is somewhat steeper, but the difference is not significant.

texture, serve to distinguish this orthogneiss from surrounding units. By tradition, a Precambrian age has been assigned, but, because the Newman Lake Gneiss is isotopically similar to the Mount Spokane pluton, we will discuss both together.

Sr analyses for granite and pegmatite of the Mount Spokane pluton do not lie on an isochron, thus creating some uncertainty in the age interpretation. Three granites and one pegmatite with highest Rb/Sr ratios give a maximum age of 84 ± 4 m.y. with initial $^{87}Sr/^{86}Sr$ ratio of 0.7108 (Table 1), which is in accord with geologic relationships and K-Ar dates for similar plutons outside the area of Eocene resetting of micas. The other two whole-rock samples contain more radiogenic Sr, so that calculated Rb-Sr dates are 113 ± 2 and 159 ± 3 m.y. (Table 1). This may indicate either a considerable time span for pluton genesis or local presence of more radiogenic initial Sr in late, discordant intrusive phases. In the Kootenay arc, major batholith emplacement began approximately 180 m.y.

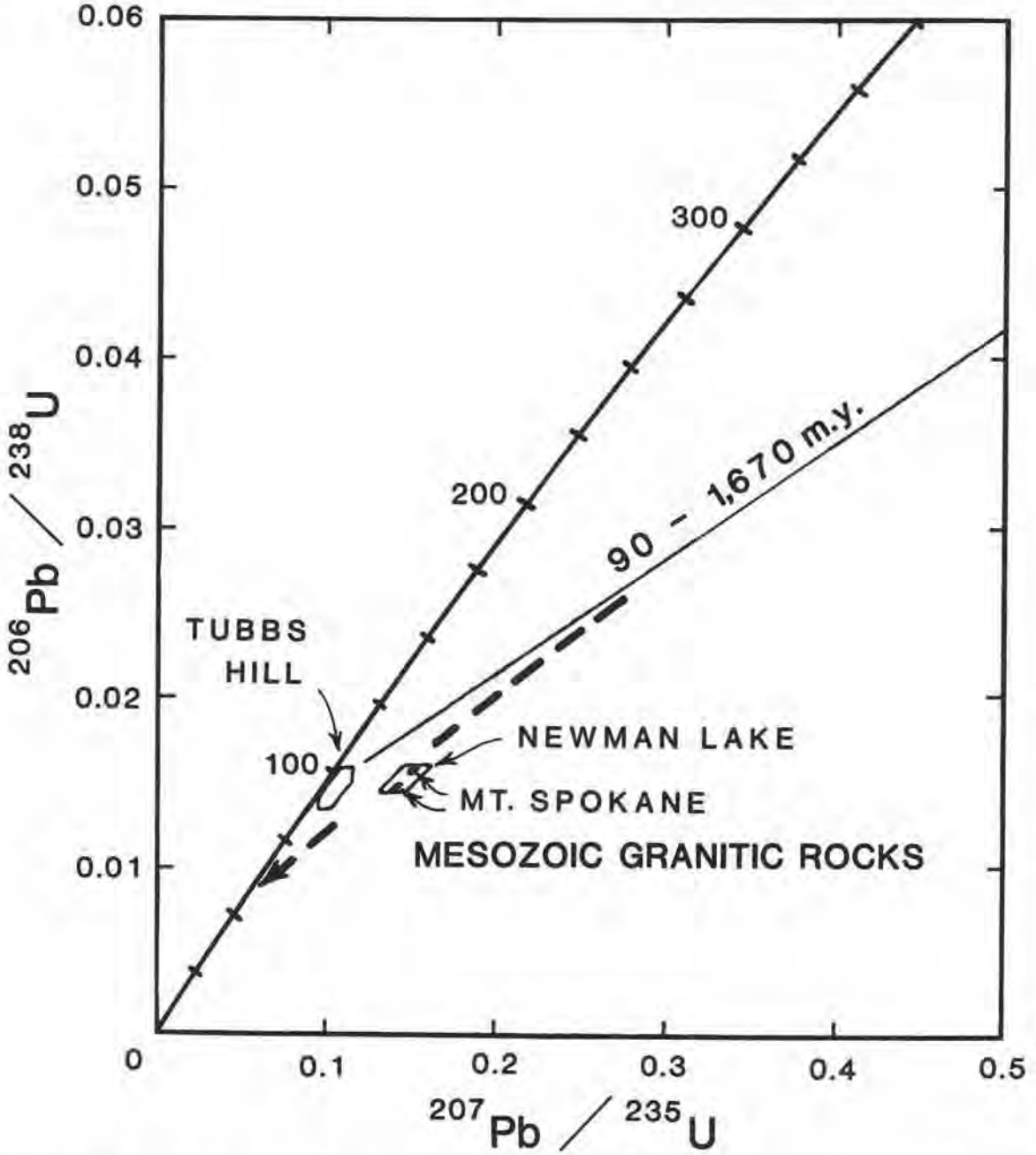

Figure 7.–U-Pb concordia plot for Mesozoic granitic rocks. The dashed line is a hypothetical multistage discordia trajectory for a Cenozoic Pb loss superimposed on 90-m.y.-old Mesozoic zircons with a trace of inherited 1,670-m.y.-old zircon. The 90-1,670-m.y. discordia line is the upper boundary of the triangular area within which these three-stage zircons would plot; it is not the same as the discordia line of Figure 4. The dashed line is a possible evolution trajectory for Mount Spokane zircons.

ago, and there were two Mesozoic culminations of igneous activity, at 173 to 157 m.y. ago and 110 to 90 m.y. ago (Gabrielse and Reesor, 1974; Armstrong, 1982b; Archibald and others, 1983, 1984), so that either a Jurassic or a mid-Cretaceous age is plausible. However, we have also found examples elsewhere where the second possibility—anomalous radiogenic Sr in pegmatites—is proven (Armstrong and Runkle, 1979; Woodsworth and others, 1983).

Two zircon fractions from two-mica granite of the Mount Spokane pluton were analyzed. The results are in reasonable agreement and somewhat discordant (Appendix 2, Fig. 7). The U-Pb dates of 94 to 143 m.y. can be interpreted to support the Rb-Sr isochron

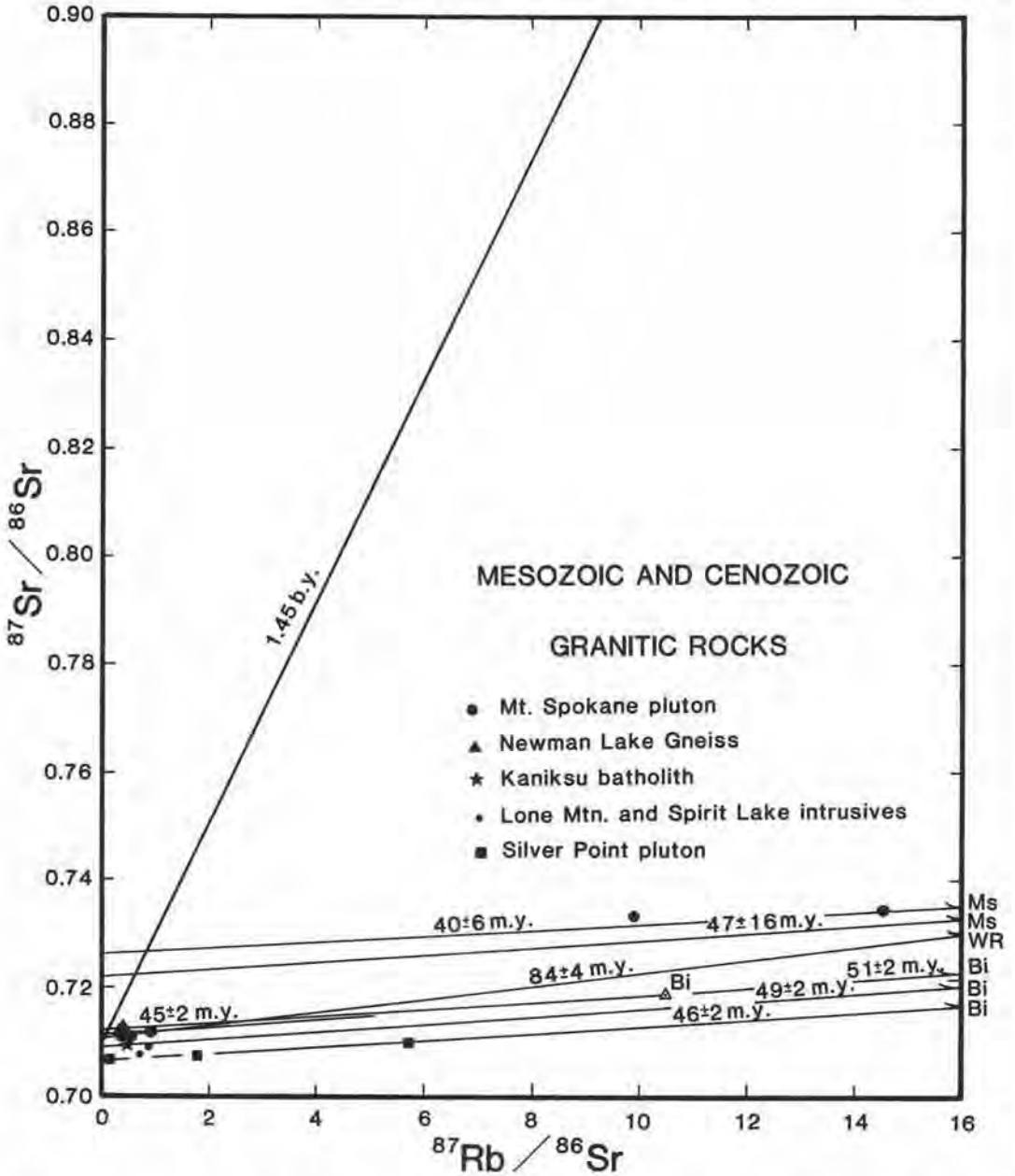

Figure 8.–Rb-Sr isochron plot for Mesozoic and Cenozoic granitic rocks. The scale is the same as that of preceding diagrams, to show the contrast in isotopic composition of these rocks with the Tubbs Hill granite and Precambrian rocks previously discussed. One whole-rock pegmatite and all minerals except one biotite (open symbol) plot outside the area shown. Whole-rock samples are solid symbols. The three-whole-rock isochron for the Silver Point pluton (39 ± 4 m.y.; Table 1) is virtually coincident with the whole-rock + biotite isochron labeled 46 ± 2 m.y. in the diagram.

date, and thus a mid- to Late Cretaceous age. The discordance is a familiar one, where small amounts of inherited zircon are present in a much younger granitic rock. The lower intercept, derived from regression of the Mount Spokane zircons with zircons likely to be the contaminant, is 60 to 75 m.y. This is lower than would be expected from a pristine Mesozoic pluton, but can be explained as the result of Eocene metamorphism causing Pb loss from already discordant zircons. The inferred history is thus three-stage, and not directly amenable to analysis on a concordia plot (Fig. 7). Three-stage zircons will plot in a triangular region with apexes at the three event points on concordia. A possible evolution trajectory for the Mount Spokane zircons is shown on Figure 7.

Two whole-rock samples of Newman Lake Gneiss have Rb/Sr and $^{87}Sr/^{86}Sr$ ratios very similar to those of the Mount Spokane pluton (Appendix 1, Fig. 8), insufficient spread in Rb/Sr to calculate an isochron age, and Rb/Sr ratios too low to distinguish Precambrian from Mesozoic using an assumed initial ratio.

Only a small amount of zircon was obtained from the Newman Lake Gneiss. Less than 1 mg of hand-picked, clear, euhedral zircon was analyzed. The obtained result is inaccurate because of large blank and common-Pb corrections, but there is no doubt that the zircon from this rock is essentially identical with that from the Mount Spokane pluton. This surprising result indicates that the gneiss is Mesozoic, not part of the Precambrian rock suite. It can be viewed as a curious border (bottom?) phase of the Mount Spokane pluton or a separate intrusive of comparable age that has been intruded or tectonically juxtaposed between the Mount Spokane and Hauser Lake units.

Of course, these ages indicate that major ductile and brittle deformation of these rocks is of Late Cretaceous or early Cenozoic age.

SILVER POINT PLUTON

Miller and Engels (1975) were confident that the Silver Point Pluton was Eocene (45-51 m.y. old), on the basis of concordant biotite and hornblende K-Ar dates. This age is important because some, and perhaps all, movement on the Newport fault (Miller, 1971) postdates that pluton (Harms, 1982). A suite of Silver Point granite and aplite dikes gave a whole-rock Sr isochron date of 39 ± 4 m.y. with $^{87}Sr/^{86}Sr$ initial ratio of 0.7063 (Table 1, Fig. 8). While not extremely precise, this result eliminates any doubt as to the Eocene age of this pluton and the structural conclusions that follow.

The $^{87}Sr/^{86}Sr$ initial ratio for the Silver Point pluton (0.7063 ± 0.0002) is the lowest we found among rocks included in this study. A low initial ratio rules out the possibility that these are Mesozoic rocks reset by an intense Eocene thermal event, and furthermore shows that crustal involvement in magma genesis was less in Cenozoic time than during the mid-Cretaceous, a circumstance common to many areas in the hinterland of the Cordilleran fold-and-thrust belt (Armstrong, 1983).

OTHER PLUTONS

Samples of Kaniksu batholith granite and of strongly lineated granodiorite from Lone Mountain have somewhat less radiogenic Sr than the Mount Spokane-Newman Lake bodies and similar, fairly low Rb/Sr ratios. There is no doubt about the mid-Cretaceous age of the Kaniksu batholith (Miller and Engels, 1975; Archibald and others, 1984). The Lone Mountain rock could be either Cretaceous or Cenozoic and slightly more radiogenic than the Silver Point pluton. Four unpublished Rb-Sr and Sr isotopic analyses done by Lois Jones of Conoco on main-phase and aplite samples collected by W. A. Rehrig, combined with our single analysis, are scattered on a $^{87}Rb/^{86}Sr$ vs. $^{87}Sr/^{86}Sr$ plot and do not give an isochron. The three least radiogenic points give a maximum date of 93 ± 39 m.y., with a minimal $^{87}Sr/^{86}Sr$ initial ratio of 0.7070 ± 0.0004 for the Lone Mountain pluton, which is affected by mylonitic deformation near the Coeur d'Alene fault.

An unfoliated biotite granite collected at Spirit Lake (Rathdrum Mountain Granite of Rhodes and Hyndman, 1984) has a Sr isotopic ratio as low as that of the Silver Point pluton and is thus inferred to be Cenozoic.

INITIAL $^{87}Sr/^{86}Sr$ RATIOS

A steep gradient in $^{87}Sr/^{86}Sr$ initial ratios of Mesozoic granitic rocks and Cenozoic volcanic rocks from <0.704 on the west to >0.706 on the east has been traced northward to 46.5° N. latitude in western Idaho, where it turns sharply westward (Fig. 1; Armstrong and others, 1977). In southern British Columbia the transition from <0.704 to >0.706 is more diffuse, extending from the Okanogan Valley to the Kootenay arc (Armstrong, 1979, 1983). Data to control the location of the 0.704 and 0.706 lines through northern Washington have been lacking. Our data for the Spokane and Sandpoint 1° x 2° quadrangles show that they lie east of any 0.706 line and that the situation may be complicated by systematically lower Cenozoic initial ratios. The Tubbs Hill pluton at Coeur d'Alene is one of the exceptionally radiogenic Cretaceous plutons that are a unique feature of the hinterland of the Cordilleran fold-and-thrust belt (Armstrong, 1983).

Rb-Sr DATES FOR MICAS

Biotite from schist, gneiss, and granitic rock and muscovite from pegmatite were dated to supplement the large number of biotite and muscovite K-Ar dates for plutonic rocks in northeastern Washington and northern Idaho (Miller and Engels, 1975). These several isotopic chronometers provide information on cooling history over a wide range of closure temperatures and are a check on the occurrence of excess Ar in polymetamorphic rocks. In all cases our samples came from within the reset terrane (K-Ar dates <65 m.y.) outlined by Miller and Engels (1975) and shown on

Figure 1. The ten mica dates (omitting the pegmatite at Tubbs Hill; Table 1) show no systematic difference between biotite and muscovite, they show no geographic pattern, and they are concordant (average 46 m.y.) with concordant biotite and muscovite K-Ar dates (45-51 m.y.) reported by Miller and Engels. The direct implication is that cooling over a temperature interval greater than 200° C was rapid, on the order of 20° to 40° C/m.y., as observed in other Cordilleran terranes reset during the Eocene (Armstrong, 1974; Medford, 1975; Mathews, 1981). This in turn requires uplift and erosion on the order of 2 km/m.y. (0.2 cm/yr) or tectonic denudation at a similar rate. At the time this occurred the region would have been as spectacularly mountainous as the Swiss Alps (Wagner and others, 1977) and Coast Mountains of British Columbia (Harrison and others, 1979; Parrish, 1983). Since then the time-averaged rates of uplift and erosion have been lower by a factor of at least 4.

CONCLUSION

This geochronometric study has supported previous geologic inference and isotopic evidence for extensive exposures of pre-Belt–Purcell basement in the core of the Priest River complex with ages between ~1,500, and 2,300 m.y., depending on isotopic system and interpretation. The migmatitic character of this basement and some deformation fabrics (Rehrig and others, this volume) are almost certainly inherited from pre-Belt–Purcell time.

Syn- to post-plutonic fabrics in Cretaceous plutons, cut discordantly by Eocene stocks, indicate late Mesozoic deformation with linear fabric oriented S.50°-60°W. (Rehrig and others, this volume), but mylonitic fabrics superimposed on Eocene plutonic rocks and dramatic cooling of the complex in Eocene time must relate to an episode of extension and tectonic denudation about 50 m.y. ago which is common to this and nearby metamorphic core complexes (Rehrig and others, this volume; Ewing, 1980; Armstrong, 1982a). Rhodes and Hyndman (1984) also recognized both late Mesozoic and early Cenozoic stages of deformation within the Priest River complex and are undertaking further detailed U-Pb work there. Their results to date (Bickford and others, 1985) confirm the results reported in this paper.

ACKNOWLEDGMENTS

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APPENDIX 1.–Sample, locality, and Rb-Sr analytical data

Sample	Rock Type or Mineral	Locality	Latitude, Longitude	Sr (ppm)	Rb (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
		Pre-Belt Basement: Hauser Lake Gneiss - Northeast of Spokane					
H@B A	Migmatitic schist: migmatitic kyanite-garnet-sillimanite-biotite schist	Blanchard, small quarry near country club	48°01.1' 116°59.8'	157	154	2.86	0.7706
H@B B	Migmatitic schist: migmatitic kyanite-garnet-sillimanite-biotite schist with few rounded garnets up to 2 cm diameter	Blanchard, small quarry near country club	48°01.1' 116°59.8'	83.7	181	6.35	0.8555

Appendix 1.–Sample, locality, and Rb-Sr analytical data (continued)

Sample	Rock Type or Mineral	Locality	Latitude, Longitude	Sr (ppm)	Rb (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
H@B C	Gneiss: garnet-kyanite-quartz-feldspar gneiss, 8-cm-thick layer in schist; garnet and kyanite up to 1 cm across; strongly lineated and foliated	Blanchard, small quarry near country club	48°01.1' 116°59.8'	135	202	4.40	0.8637
H@B Gn	Leucogneiss: medium to coarse-grained flaser quartz-feldspar leucogneiss; concordant layer, 0.5 m thick, enclosed in schists; rust-stained spots and streaks, some micaceous streaks, strong cataclastic lineation	Blanchard, small quarry near country club	48°01.1' 116°59.8'	462	108	0.681	0.7360
H@SL	Schist-gneiss: mixed sillimanite-biotite schist and biotite-quartz-feldspar gneiss	Spirit Lake just S of town on Bricket Creek Rd.	47°57.5' 116°52.4'	560	50.9	0.263	0.7235
SL 2A	Gneiss: fine-grained, yellowish-brown quartz-feldspar granulite gneiss with graphite flakes	N side of Spirit Lake, along Bricket Creek Rd.	47°56.8' 116°55.1'	56.6	22.3	1.142	0.7209
SL 2B	Gneiss: fine-grained, grayish-orange quartz-feldspar granulite gneiss with trace of disseminated biotite	N side of Spirit Lake, along Bricket Creek Rd.	47°56.8' 116°55.1'	123	27.4	0.645	0.7248
M†Sk77-1	Schist: fine-grained, medium-gray, slabby biotite schist and flaser sillimanite-biotite schist	Roadcut along Newman Lake Ctr. sec. 10, T.26N., R.45E.	47°46.2' 117°05.9	328	82.4	0.729	0.7345
M†Sk77-1	Biotite from schist	Roadcut along Newman Lake Ctr. sec. 10, T.26N., R.45E.	47°46.2' 117°05.9'	14.1	551	115	0.8081
79-10-16 composite	Gneiss and schist: migmatitic biotite gneiss and sillimanite-biotite schist; composite of samples 79-10-16-2 through 79-10-16-5	Roadcut, N side of Hauser Lake	47°47' 117°01'	296	99.4	0.973	0.7358

Appendix 1.–Sample, locality, and Rb-Sr analytical data (continued)

Sample	Rock Type or Mineral	Locality	Latitude, Longitude	Sr (ppm)	Rb (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
79-10-16-2	Gneiss: mylonitic sillimanite-muscovite-biotite-quartz-feldspar gneiss	Roadcut, N side of Hauser Lake	47°47' 117°01'	521	48.6	0.271	0.7318
79-10-16-3	Gneiss: foliated and lineated, biotite-rich quartz-feldspar-mica gneiss with granitic pods	Roadcut, N side of Hauser Lake	47°47' 117°01'	357	97.0	0.789	0.7350
79-10-16-4	Gneiss: quartz-rich granitic gneiss interlayered with sillimanite-muscovite-biotite-feldspar-quartz schist	Roadcut, N side of Hauser Lake	47°47' 117°01'	338	64.2	0.551	0.7345
79-10-16-5	Gneiss: biotite-rich sillimanite-mica-quartz-feldspar gneiss with granitic pods	Roadcut, N side of Hauser Lake	47°47' 117°01'	184	86.8	1.369	0.7383
PR 77-2	Augen-gneiss: medium-grained, biotitic flaser-augen gneiss with feldspar augen up to 4 cm long; well foliated; discordant intrusive	Near Priest River, 3 km SW of Laclede, along Hwy. 2/195	48°10.5' 116°47.4'	245	193	2.30	0.7504
PR 77-2	Biotite from augen gneiss	Near Priest River, 3 km SW of Laclede, along Hwy. 2/195	48°10.5' 116°47.4'	29.8	830	81.3	0.7914
	Other Precambrian Rocks: Elk City Gneiss and Belt-Purcell sediments						
ECRS 77-1	Augen gneiss: coarse augen gneiss with nearly euhedral feldspar megacrysts up to 5 cm across in matrix of biotite-quartz-feldspar gneiss	Red River SW of Elk City	45°48.3' 115°28.6'	122	185	4.43	0.8007
Yp	Siltite: aphanitic, light olive-gray siltite with gray siltite with rust-stained fractures; Prichard Formation	4 km NNW of Newport, WA along Hwy. 31	48°12.8' 117°04.2'	67.0	174	7.60	0.8359

Appendix 1.–Sample, locality, and Rb-Sr analytical data (continued)

Sample	Rock Type or Mineral	Locality	Latitude, Longitude	Sr (ppm)	Rb (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
		Pre-Belt Basement and Mesozoic Hybrid Gneisses: Davis Lake roadcut and southeast of Spokane					
DL-A	Gneiss: medium-grained, lineated and seriate, light-brownish-gray quartz-feldspar gneiss with muscovite and feldspar porphyroclasts, muscovite and biotite present in matrix	Roadcut east of Davis Lake along Hwy. 331	48°13.6' 117°17.2'	374	132	1.019	0.7211
DL-B	Pegmatite: coarse-grained, mylonitic muscovite-quartz-feldspar pegmatite with muscovite and feldspar porphyroclasts and quartz ribbons several cm long	Roadcut east of Davis Lake along Hwy. 331	48°13.6' 117°17.2'	74.0	132	5.16	0.7208
DL-C	Gneiss: medium-grained, well foliated and lineated, brownish-gray mylonitic gneiss with muscovite and feldspar porphyroclasts	Roadcut east of Davis Lake along Hwy. 331	48°13.6' 117°17.2'	146	169	3.37	0.7730
DL-D	Mylonite: fine-grained, lineated, greenish-gray mylonite	Roadcut east of Davis Lake along Hwy. 331	48°13.6' 117°17.2'	1060	92.1	0.251	0.7059
DL-E	Gneiss: medium- to fine-grained, lineated, brownish-gray mylonitic gneiss with muscovite and feldspar porphyroclasts	Roadcut east of Davis Lake along Hwy. 331	48°13.6' 117°17.2'	113	140	3.61	0.7524
DL-F	Gneiss: fine- and medium-grained, variable lineated, slightly mylonitic, light-gray muscovite-biotite-quartz-feldspar gneiss	Roadcut east of Davis Lake along Hwy. 331	48°13.6' 117°17.2'	260	147	1.643	0.7167
DL-H	Pegmatitic gneiss: medium- to coarse-grained, strongly lineated, mylonitic flaser gneiss; almost pegmatitic; muscovite and feldspar porphyroclasts in quartz-feldspar matrix	Roadcut east of Davis Lake along Hwy. 331	48°13.6' 117°17.2'	85.0	163	5.55	0.7231

Appendix 1.–Sample, locality, and Rb-Sr analytical data (continued)

Sample	Rock Type or Mineral	Locality	Latitude, Longitude	Sr (ppm)	Rb (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
Chester -1	Gneiss: well-foliated, thinly laminated biotite-muscovite gneiss	2 km E of Chester on Hwy. 27	47°36.9' 117°13.3'	246	154	1.823	0.7386
Chester -2A	Quartzite: light-gray, feldspathic quartzite with biotite- and muscovite-rich partings	5 km SSE of Chester, 0.6 km W of Hwy. 27	47°34.4' 117°13.8'	69.6	73.8	3.08	0.7547
Chester -2 PEG	Pegmatite: Coarse-grained biotite-muscovite-quartz feldspar pegmatite	5 km SSE of Chester, 0.6 km W of Hwy. 27	47°34.4' 117°13.8'	113	111	2.84	0.7407
Chester -2 PEG	Muscovite from pegmatite	5 km SSE of Chester, 0.6 km W of Hwy. 27	47°34.4' 117°13.8'	17.0	520	89.4	0.807
Chester -3B	Gneiss: two-mica gneiss with wavy micaceous layers	5 km S of Chester, 0.8 km NW of Valleyford Junction	47°34.5' 117°14.5'	202	240	3.45	0.7404
Chester 3E PEG	Pegmatite layer: coarse quartz-feldspar pegmatite layer in Chester -3E GN gneiss	5 km S of Chester, 0.8 km W of Hwy. 27	47°34.5' 117°14.5'	260	117	1.309	0.7515
Chester 3E GN	Gneiss: medium- to fine-grained biotite gneiss	5 km S of Chester, 0.8 km NW of Valley-Ford Junction	47°34.5' 117°14.5'	118	95.1	2.33	0.7562
Cdl 77-1	Gneiss: fine-grained, yellowish-gray, biotite-rich quartz-feldspar gneiss; banded, almost migmatitic, strongly lineated	WSW of Coeur d'Alene, along Cougar Creek	47°38.1' 116°55.4'	455	56.6	0.361	0.7152
Cdl 77-1	Biotite from gneiss	WSW of Coeur d'Alene, along Cougar Creek	47°38.1' 116°55.4'	18.1	326	52.4	0.7469
	Tubbs Hill: Precambrian Pegmatite, Cretaceous Leucogranite						
Cdl 77-2	Pegmatite: coarse, mylonitic biotite-muscovite-quartz-feldspar pegmatite; muscovite books up to 2 cm across, bent and distorted, quartz-ribboned	NW flank of Tubbs Hill, Coeur d'Alene	47°40.15' 116°46.9'	129	110	2.49	0.7839

Appendix 1.–Sample, locality, and Rb-Sr analytical data (continued)

Sample	Rock Type or Mineral	Locality	Latitude, Longitude	Sr (ppm)	Rb (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
Cdl 77-2	Muscovite from pegmatite	NW flank of Tubbs Hill,	47°40.15' 116°46.9'	10.2	554	177	1.995
79-10-16-1	Leuco-granite: medium- to fine-grained, lineated weakly foliated garnet-biotite-muscovite leucogranite (adamellite)	Tubbs Hill, Coeur d'Alene	47°40.15' 116°46.9'	140	140	2.903	0.7514
Tubbs	Leuco-granite: medium- to fine-grained, lineated weakly foliated garnet-biotite-muscovite leucogranite (adamellite)	Tubbs Hill, Coeur d'Alene	47°40.15' 116°46.9'	130	181	4.05	0.7536
	Mesozoic Plutons: Mount Spokane, Newman Lake, Lone Mountain, and Kaniksu						
RS 77-1	Granite: medium-grained, unfoliated muscovite-biotite granite (adamellite); Mount Spokane pluton	Road to Mount Spokane along Deer Creek; SE 1/4 SE 1/4 sec. 34, T.28N., R.44E	47°52.7' 117°13.0'	563	190	0.997	0.7120
RS 77-1	Biotite from granite	Road to Mount Spokane along Deer Creek; SE 1/4 SE 1/4 sec. 34, T.28N., R.44E.	47°52.7' 117°13.0'	43.0	1094	74.2	0.7653
RS 77-2	Pegmatite: coarse, graphic, garnet-muscovite-quartz-muscovite-quartz-feldspar pegmatite; Mount Spokane pluton	Road to Mount Spokane along Deer Creek; SE 1/4 SE 1/4 sec. 34, T.28N., R.44E.	47°52.7' 117°13.0'	65.0	484	21.6	0.7365
RS 77-2	Muscovite from pegmatite	Road to Mount Spokane along Deer Creek; SE 1/4 SE 1/4 sec. 34, T.28N., R.44E.	47°52.7' 117°13.0'	3.0	1924	2116	2.138
RS-77-3	Granite: fine-grained leucogranite (adamellite) with patchy zones of disseminated biotite flakes; Mount Spokane pluton	Mt. Spokane Rd., 0.2 km E of Independent School; Ctr. sec. 13, T.27N., R.44E.	47°50.25' 117°11.0'	700	136	0.562	0.7112

Appendix 1.–Sample, locality, and Rb-Sr analytical data (continued)

Sample	Rock Type or Mineral	Locality	Latitude, Longitude	Sr (ppm)	Rb (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
RS 77-4	Granite: fine- to medium-grained, weakly foliated, seriate, almost pegmatitic muscovite granite (adamellite); Mount Spokane pluton	Road parallel to Mt. Spokane Rd., Ctr. sec. 6, T.27N., R.45E.	47°52.0' 117°09.7'	70.7	242	9.93	0.7334
RS 77-5	Pegmatite: coarse, partly graphic, muscovite-quartz-feldspar pegmatite; Mount Spokane pluton	Mt. Spokane Rd., 5 km N of Independent School; Ctr. sec. 7, T.27N., R.45E.	47°51.3' 117°09.8'	78.4	394	14.6	0.7345
RS 77-5	Muscovite from pegmatite	Mt. Spokane Rd., 5 km N of Independent School; Ctr. sec. 7, T.27N., R.45E.	47°51.3' 117°09.8'	6.6	1524	697	1.118
Mt. Spokane	Granite: medium-grained, unfoliated muscovite-biotite (adamellite); Mount Spokane pluton	Mt. Spokane Rd., 0.2 km E of Independent School	47°50.25' 117°11.0'	712	111	0.449	0.7115
MtSk 77-2	Augen gneiss: porphyritic-flaser augen gneiss with 2- to 3-cm feldspar megacrysts in moderately foliated and lineated biotite-granodiorite gneiss; Newman Lake Gneiss	Roadcut NW of Newman Lake; N 1/2 N 1/2 sec. 33, T.27N., R.45E.	47°48.0' 117°07.1'	768	101.5	0.383	0.7121
Mtsk 77-2	Biotite from augen gneiss	Roadcut NW of Newman Lake; N 1/2 N 1/2 sec. 33, T.27N., R.45E.	47°48.0' 117°07.1'	157	569	10.5	0.7185
Newman Lake	Augen gneiss, same as MtSk 77-2	Newman Lake, W side	47°46.7' 117°07.0'	717	102	0.413	0.7129
LM 77-1	Granodiorite: fine-grained, medium-light-gray, strongly lineated, weakly foliated biotite granodiorite	Lone Mtn., cliff on N side	47°53.9' 116°49.0'	378	116	0.888	0.7091

Appendix 1.–Sample, locality, and Rb-Sr analytical data (continued)

Sample	Rock Type or Mineral	Locality	Latitude, Longitude	Sr (ppm)	Rb (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
KB 77-1	Granite: medium-grained, unfoliated biotite adamellite; Kanisku batholith	Trout Creek Rd.; Ctr. sec. 15, T.63N., R.2W., Boundary Co., ID	48°48.6' 116°34.6'	599	98.7	0.477	0.7095
KB 77-1	Biotite from granite	Trout Creek Rd.; Ctr. sec. 15, T.63N., R.2W., Boundary Co., ID	48°48.6' 116°34.6'	22.2	734	96.3	0.7767
		Cenozoic Plutons: Silver Point and Spirit Lake					
RS 77-6	Granite: medium- to coarse-grained, seriate hornblende-biotite granite (adamellite), main phase of Silver Point pluton	NW 1/4 NW 1/4 sec. 24, T.30N., R.43E., on Hwy. 2/195, Pend Oreille Co., WA	48°05.35' 117°19.4'	1471	90.4	0.178	0.7065
RS 77-6	Biotite from granite	NW 1/4 NW 1/4 sec. 24, T.30N., R.43E., on Hwy. 2/195, Pend Oreille Co., WA	48°05.35' 117°19.4'	62.9	371	17.09	0.7176
RS 77-7	Aplite: fine-grained, light-gray aplite speckled with small biotite flakes; Silver Point pluton; 15-cm-wide dike trending N 40° W	Sec. 13, T.30N., R.43E., on Hwy. 2/195, Pend Oreille Co., WA	48°05.7' 117°19.0'	189	112	1.719	0.7071
RS 77-8	Aplite: fine-grained, very pale orange aplite speckled with small biotite flakes; Silver Point pluton; 7.5-cm-wide dike	SW 1/4 SW 1/4 sec. 13, T.30N., R.43E., on Hwy. 2/195, Pend Oreille Co., WA	48°05.45' 117°19.3'	98.8	194	5.69	0.7095
SL-Ti	Granite: fine-grained, equigranular biotite granite (adamellite)	N side of Spirit Lake, along Bricket Creek Rd., WA	47°56.8' 116°55.1'	465	112	0.700	0.7077

APPENDIX 2.–Isotopic analyses of zircons: U-Pb and Pb-Pb dates

Isotopic composition of blank: 6/4 = 17.75; 7/4 = 15.57; 8/4 = 37.00; $^{206}Pb/^{204}Pb$ = 17.75. Isotopic composition of common lead based on Stacey-Kramers growth curve: 206/204 = 11.152, 207/204 = 12.998. 208/204 = 31.23 at 3.7 b.y. with $^{238}U/^{204}Pb$ = 9.74, $^{232}Th/^{204}Pb$ = 137.88. 1.700 m.y. Pb used for 6 older zircons, 100 m.y. for 4 younger zircons. Errors (2σ) are derived from spike calibrations and fractionation uncertainty and uncertainties in mass-spectrometer within-run isotope-ratio measurement.

Sample	Split	U (ppm)	Pb (ppm)	Relative abundance ^{206}Pb = 100 ^{207}Pb	^{208}Pb	^{204}Pb	$^{206}Pb/^{204}Pb$ (measured)	Blank Pb (mole %)	Radiogenic Pb / Radiogenic + Common Pb	$^{206}Pb/^{238}U$ (m.y. ± 2σ)	$^{207}Pb/^{235}U$ (m.y. ± 2σ)	$^{207}Pb/^{206}Pb$ (m.y. ± 2σ)
79-10-16 Composite	0.15 mg +100 mesh HP	608	27.7	15.445	22.792	0.412	86.1	32.1	0.800	226 ± 5.1	99 ± 28	1,580 + 137 - 151
	9.45 mg 200-325 mesh	663	51.0	13.199	22.555	0.270	349	0.8	0.866	394 ± 2.3	09 ± 3.7	1,522 ± 13
	14.4 mg 200-325 mesh	634	49.9	12.455	20.634	0.21	414	1.1	0.892	414 ± 2.2	34 ± 3.6	1,527 ± 11
	22.9 mg 100-200 mesh	717	65.4	9.926	15.188	0.015	2,390	0.6	1.000	523 ± 2.7	68 ± 3.1	1,570 ± 5
HØB GN	29.5 mg -200 mesh	1,621	91.8	8.279	1.117	0.021	3,584	0.6	0.987	376 ± 2.0	15 ± 2.1	1,193 ± 3
	11.3 mg +70 mesh HP	2,799	171	8.296	0.568	0.016	4,348	0.6	0.990	406 ± 2.4	53 ± 2.6	1,214 ± 4
Tubbs Hill	0.1 mg -200 mesh HP	4,580	66.5	6.879	6.790	0.111	298	13.0	0.928	93 ± 4.4	02 ± 7.3	310 + 120 - 130
Newman Lake	0.6 mg 200-235 mesh HP	408	10.4	20.172	41.054	0.921	50.5	41.7	0.588	97 ± 5.2	39 ± 14	924 + 170 - 191
Mount Spokane	10.1 mg +200 mesh	1,077	17.7	9.851	15.149	0.199	311.5	7.7	0.883	93.7 ± 0.5	34.2 ± 0.9	927 ± 11
	43.8 mg -200 mesh	1,208	19.2	6.973	9.417	0.004	8,993	0.5	0.998	01.4 ± 0.5	43.0 ± 0.8	905 ± 6

APPENDIX 3.–Analytical details

U-Pb ANALYSES

Zircons were separated from finely crushed 20- to 40-kg rock samples, using wet shaking table, heavy liquids, and a magnetic separator. They were acid-washed in strong aqua regia, sized using nylon-mesh screens, and hand-picked as required. Chemical dissolution and mass spectrometry followed the procedures of Krogh (1973). We use a mixed ^{208}Pb-^{235}U spike and measured Pb and U on an automated Vacuum Generators Isomass 54R solid-source mass spectrometer. Automation and data reduction were done with a dedicated Hewlett-Packard HP-85 computer.

U-Pb date errors (2σ) were obtained by individually propagating all calibration and analytical uncertainties through the entire date calculation program and summing all the individual contributions to the total variance.

U decay constants and isotope ratio are:

$$^{238}U\lambda = 0.155125 \times 10^{-9}\ yr^{-1}$$
$$^{235}U\lambda = 0.98485 \times 10^{-9}\ yr^{-1}$$
$$^{238}U/^{235}U = 137.88$$

All published dates cited in the text have been recalculated, as necessary, to conform to these constants.

Rb-Sr ANALYSES

Rb and Sr concentrations were determined by replicate analysis of pressed powder pellets using X-ray fluorescence. U.S. Geological Survey rock standards were used for calibration; mass absorption coefficients were obtained from Mo Kα Compton scattering measurements. Rb/Sr ratios have a precision of 2 percent (1σ) and concentrations a precision of 5 percent (1σ). Sr isotopic composition was measured on unspiked samples prepared by standard ion-exchange techniques. The mass spectrometer, Vacuum Generators Isomass 54R, has data acquisition digitized and automated using a Hewlett-Packard HP-85 computer. Experimental data have been normalized to a $^{86}Sr/^{88}Sr$ ratio of 0.1194 and adjusted so that the NBS standard $SrCO_3$ (SRM987) gives a $^{87}Sr/^{86}Sr$ ratio of 0.71020 ± 2 and the Eimer and Amend Sr a ratio of 0.70800 ± 2. The precision of a single $^{87}Sr/^{86}Sr$ ratio is <0.00010 (1σ). Rb-Sr dates are based on a Rb decay constant of $1.42 \times 10^{-11}\ yr^{-1}$. The regressions were calculated according to the technique of York (1967).

Decay constants are those recommended by the IUGS Subcommission on Geochronology (Steiger and Jäger, 1977). Reported errors are for one standard deviation or the standard error of the mean, unless otherwise noted.

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RECENT STRUCTURAL ANALYSES OF THE KOOTENAY ARC IN NORTHEASTERN WASHINGTON

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ABSTRACT

During the late Precambrian and Cambrian, the passive, rifted margin of the North American continent appears to have been aligned on a north-south trend through northeastern Washington. The Sauk transgressive sequence of quartzites (Gypsy Quartzite), shales (Maitlen Phyllite), and limestones and dolomites (Metaline Limestone) is seen in the upper Proterozoic–Cambrian sequences, which are typical miogeosynclinal shelf assemblages. The Mississippian Antler orogeny that is expressed in the geologic record to the south in Nevada (and as the Cariboo orogeny to the north in British Columbia) has no obvious record in northeastern Washington. Structures of "Nevadan" polarity in the Inchelium region are present, along with local middle Devonian sedimentary rocks interpreted to be the result of uplift. However, until the metamorphism associated with the deformation is dated, evidence of Paleozoic orogeny remains equivocal.

By comparison with structures immediately to the north in British Columbia (Kootenay arc), the main tectonic events in Washington appear to have commenced at the margin with the collision of Quesnellia in approximately mid-Jurassic time, followed later by Wrangellia.

Recent detailed structural and strain analyses (mostly in the Maitlen units) give us a clearer picture of the structure of the margin and accreted terranes—a form of wedge delaminating of the margin rocks, plus a dextral(?) shear component, resulting in along-strike extension, followed by vertically emergent F_2 structures with thrust-belt-style tectonics (pop-up and triangular zones).

These Mesozoic compressional structures are preserved in an Eocene extensional structural low between the Colville batholith to the west and the Priest River complex (Spokane gneiss dome) to the east. Paleomagnetic and structural data show dextral rotation of structural blocks during the Eocene extension. COCORP data, corroborating the structural analysis presented here, suggest east-dipping extensional faults to the Moho that cut the previous, contractional, west-dipping thrusts.

INTRODUCTION: GENERAL KOOTENAY ARC GEOLOGY

The Kootenay arc is a 400-km-long, arcuate belt of repeatedly deformed metasedimentary rocks, exposed from central British Columbia south and southwestward toward Spokane, Washington. In Washington, the rocks range in age from late Proterozoic to Triassic and comprise, for the most part, rocks characteristic of a miogeosynclinal environment. Yates (1976) has grouped the arc rock units into two assemblages: a western assemblage of greenstones, argillites, limestones, cherts, conglomerates, and pillow lavas; and an eastern assemblage of quartz arenites, shales, and limestones. The eastern assemblage is representative of the transgressive Sauk sequence and occupies the late Precambrian to Cambrian continental shelf, in a setting perhaps analogous to the eastern seaboard of the United States today. The western assemblage is related to a eugeosynclinal environment (accretionary and/or slope sediments?), although few geological studies of these rocks have been completed.

The regional metamorphic grade of rocks exposed in the arc ranges from "amphibolite" facies in British Columbia (Crosby, 1968; Hoy, 1977) to the chlorite zone of the "greenschist" facies in the United States. Later contact-metamorphic overprinting by late Mesozoic acidic plutons is observed throughout the arc. In northern Washington, several porphyroblast phases were formed (for example, andalusite, tremolite-biotite). Pseudomorphs of two other porphyroblastic phases are also recognized in Washington: chlorite after actinolite in the Ordovician slates, and chlorite after biotite(?) in the Precambrian McHale Slate (Ellis, 1984).

The eastern boundary of the Kootenay arc is represented by the Belt Supergroup, a sequence of late

Proterozoic geosynclinal clastic rocks. The Purcell series in British Columbia is involved in a large anticlinorium which plunges gently toward the north at the northern end of the Kootenay arc. The differences between the Purcell and Kootenay arc structures are not as significant as once imagined; Price (1981) considered the arc to represent the westernmost "monoclinal" shoulder of the Purcell anticlinorium, although this view may not be completely justified (Ellis and Watkinson, 1984). The structural style of the Belt Supergroup adjacent to the arc in the United States is incompletely understood. The western boundary of the arc is marked by a clear metamorphic and structural break in Washington (Wilson, 1981), where higher grade rocks comprising the Kettle gneiss dome and associated "granites" are exposed.

The structural features give the arc its geographic form; two essentially coaxial fold systems having axes parallel to the trend of the arc dominate the style of deformation (Crosby, 1968; Fyles, 1962, 1967, 1970; Fyles and Hewlett, 1959; Hedley, 1955; Hoy, 1977; Ross, 1970; Mills and Nordstrom, 1973). In Stevens County, Washington, the first folds are tight to isoclinal and overturned to recumbent, facing both to the east and to the west. An associated penetrative axial planar cleavage, S_1, is seen in places to parallel low-angle reverse faults. A well-developed mineral lineation is observed parallel to the first fold axes and is contained within S_1. Second folds are nearly open and upright to overturned (both east and west), usually exhibiting a spaced cleavage. The style of the second folds is more varied than that of the first (Bressler, 1979; Conklin, 1981; Moser, 1978; Phillips, 1979; O'Keefe, 1980), varying from kink bands to rounded-hinge and chevron folds. A minor third system of folds is recognized throughout the arc as an orthogonal set, causing the earlier folds to plunge gently to the north and south.

The Kootenay arc in northern Washington between the Columbia and Pend Oreille Rivers has been mapped reconnaissance by Yates in the Northport area (1971) and Deep Creek area (1964), by Dings and Whitebread (1965) in the Metaline district, by Campbell and Loofbourow (1962) farther south near Chewelah, and most recently by Burmester and Miller (1983) in the area of Russian Creek and Abercrombie Mountain. Yates (1976) described the lithologies and stratigraphy in some detail and provided a discussion of the general tectonic setting. However, to date, there has been no structural analysis and synthesis of this northern section of the Kootenay arc. Recent work by Mills and others (1985) has provided structural information for the central part of the arc in the Kettle Falls area, and recent work by Snook and others (1981) provides more information about the more southerly part, near Inchelium (Fig. 1).

This paper presents a summary of detailed structural analyses that have been carried out in rocks of the Kootenay arc in northeastern Washington near the Canadian border. The major outcrops of the Maitlen Phyllite and field-associated lower Cambrian and upper Paleozoic rocks have been mapped northward along the Columbia River from China Bend (Bressler, 1979; Conklin, 1981; Moser, 1978; Phillips, 1979) to Squaw Creek (Beka, 1978, 1980) and Northport (Schriber, 1981; Mills and Nordstrom, 1973; Ellis, 1984). About 10 mi east of Northport along Deep Creek, Haviland (1983) mapped at Cedar Lake, and O'Keefe (1980) mapped on Red Top Mountain.

The distinct lithology of the Maitlen Phyllite (Table 1) and the ease with which carbonate units can be correlated throughout the region (in particular, the lowermost Reeves Limestone Member, which is recognized as the well-known Badshot Formation in British Columbia) have made possible detailed structural mapping. Furthermore, the alternating phyllites and limestones of the Maitlen have been sensitive to the deformational events that have produced the variety of ductile features now visible, particularly when compared the more massive and homogeneous units below (Gypsy Quartzite) and above (Metaline Limestone). The ductile-folding phases, the faulting, and structural relationships between the upper and lower Paleozoic rocks have been more clearly defined in theses areas. The structural sequence observed within the Maitlen serves as a model for the region, where poor exposure and more massive and monotonous sequences frequently make detailed analyses difficult. Very little detailed stratigraphic analysis of the major units, such as the Metaline Limestone, has yet been attempted in this area, in contrast to that done farther east in the vicinity of Metaline Falls (Fischer, 1976, 1981); this lack of stratigraphic analysis is in part due to the regional metamorphism and recrystallization.

STRATIGRAPHY

The mapped rocks range from the upper Gypsy Quartzite, through the Maitlen Phyllite, to lower and middle Metaline Limestone, and to upper Paleozoic limestones, phyllites, and argillites. The original details of both stratigraphy and sedimentary fabric have been significantly altered by recrystallization and formation of a penetrative tectonite fabric; the latter is associated with an almost ubiquitous bedding-normal shortening of 50 to 60 percent (Ellis, 1984) and comprises well-developed cleavage and stretching lineation. Although deformed, fossils are preserved (for example, conodonts in Devonian limestones; Beka, 1980), and origi-

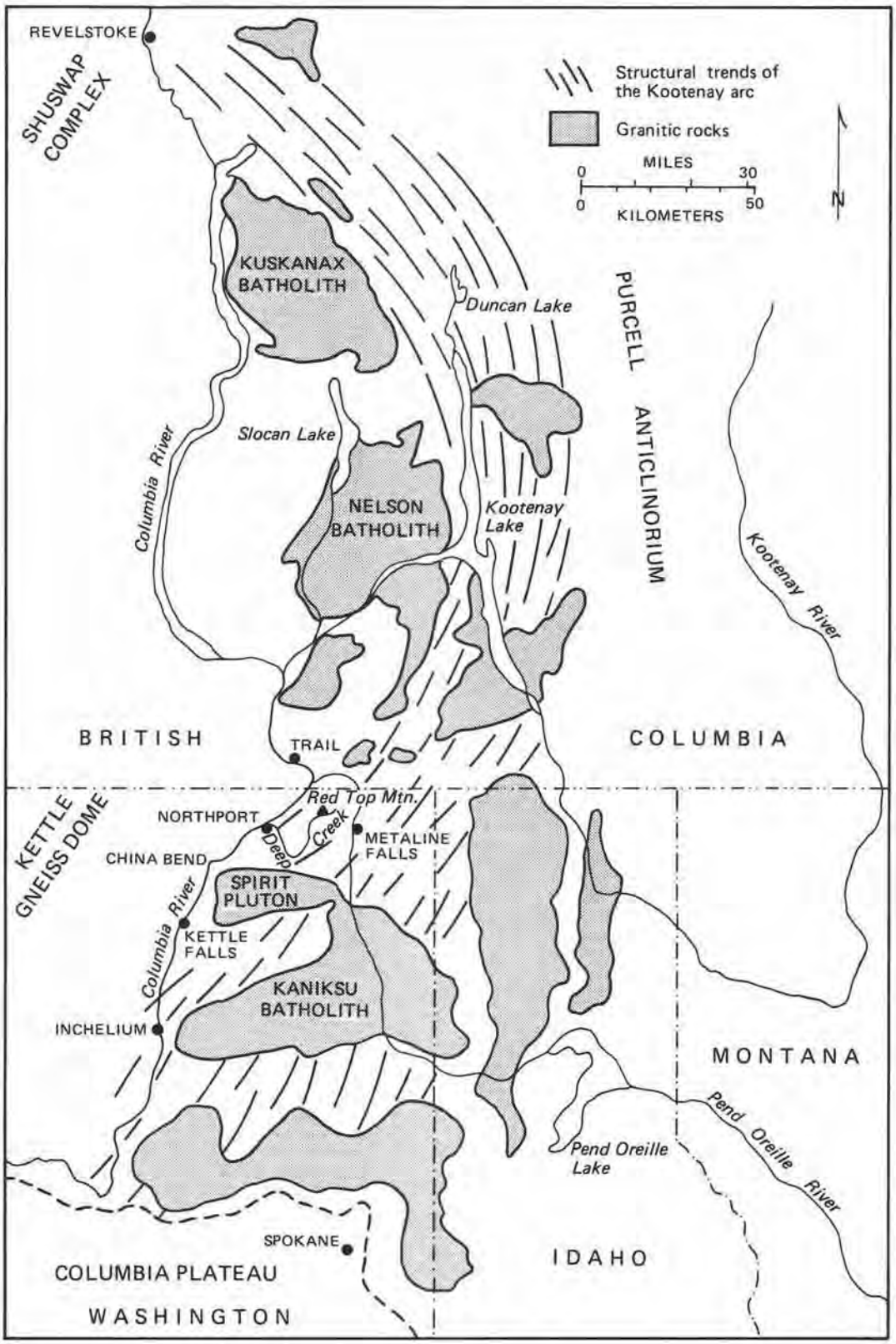

Figure 1.–Location map and general geology of the Kootenay arc (adapted from Fyles, 1970).

nal sedimentary structures are well preserved in, for example, the Gypsy quartzites.

The Maitlen Phyllite contains the first limestone (Reeves Limestone Member) above the Gypsy Quartzite and stratigraphically higher alternating phyllites and two more limestones, up to the massive carbonate sequence of the Metaline Limestone (Table 1). No angular unconformities have been observed, although the regional strain and recrystallization may have obscured any such feature. Because of the structural effects, bed thicknesses vary with position within the major folds and the amount of strain. The best stratigraphic section we have observed is that at the cliffs at China Bend (Bressler, 1979; Conklin, 1981); although faulted, the Maitlen sequence on the upper limb of a major F_1 recumbent fold there is well exposed. Another complete section from the Gypsy through the Maitlen to the Metaline is on the upper limb of another F_1 overturned fold on Gorge Mountain, 2.5 km south of Northport (Phillips, 1979). Archaeocyathids have been found in the Reeves Limestone Member at Colville, giving an early Cambrian age limit for at least the lower part of the Maitlen Phyllite. The Maitlen Phyllite is clearly a transgressive sequence linking the Gypsy Quartzite and Metaline Limestone, suggesting a continental-margin origin and consequently an autochthonous or parautochthonous origin for at least the presumed lower Paleozoic sequence in this Northport area.

STRUCTURES

The major ductile structures in the study areas are the large overturned to recumbent F_1 folds. All the folds have a penetrative axial planar cleavage, and all the anticlines have cores of Gypsy Quartzite. Only a synthesis of the structures will be given here, since detailed structural analyses have been presented in cited theses.

The China Bend area (Fig. 2A) is the upright limb of a major recumbent anticline, facing northwest, with the overturned limb partially exposed on the west. The hinge is disrupted by thrust faults which are essentially parallel to the axial planar cleavage.

South of Northport, the Lehigh quarry area exhibits a well-preserved northwest-facing F_1 anticline complicated by F_2 folding and, again, possible faulting in the hinge zone. Just to the west and northwest of Northport, a fault-disrupted Maitlen section is faulted against the Metaline Limestone, and all these rocks appear to be folded into an overturned F_1 syncline facing northwest.

In the valley of Deep Creek, west of Cedar Lake, a major F_1 anticline closes northwestward, whereas immediately on the east side of the valley the axial plane is significantly steeper (Fig. 2B). Immediately to the east, on Red Top Mountain (Fig. 3), the major F_1 fold faces southeast; this fold is similarly disrupted through the hinge area by a low-angle fault, essentially parallel to the axial plane, and, again, the fault appears either to be folded by the F_2 folds or to cause the F_2 folds (Fig. 3).

All these areas show refolded structures; these range from microscopic structures with associated crenulation cleavage, to mesoscopic structures of variable style (depending on lithology; Conklin, 1981), to large-scale structures. At China Bend and Northport, the F_2 folds verge toward the Columbia River from both sides, suggesting that the river in part follows the hinge zone of a major, steeply inclined F_2 fold.

The F_1 and F_2 folds are essentially coaxial with a regional southwest-northeast trend. F_2 is frequently 10° to 15° more westerly than F_1, and significant deviations from coaxiality between F_1 and F_2 have been observed in impressive fold-interference patterns at Red Top Mountain (O'Keefe, 1980) and at Cedar Lake (Thiessen and Haviland, 1986). However, the dominant structural fabric of the region is nearly coaxial. F_1 axial lineations, along with a penetrative strain extension lineation (involving at least 120 percent elongation), formed within the S_1 surface and parallel to the F_1 fold axes.

F_1 formed within the chlorite zone of the greenschist facies. (Chlorite is boudinaged within S_1.) In places, minor amounts of biotite formed, either because Fe- and Mg-rich phases were present or because higher grade conditions were reached in certain zones. In general, F_2 folds deform the older metamorphic phases, and only minor signs of retrogression have been observed. Quartz-fiber fabrics show that the principal extension during F_2 was vertical.

Localized, minor kink folds, crenulations, and broad culminations are generally at high angles to the southwest-northeast trend of the earlier folds described above. These later F_3 folds may account for the change in plunge of F_1 and F_2 structures from a shallow plunge to the southwest to a shallow plunge to the northeast.

As Yates' maps (1964, 1971) show, the region contains many fault-bounded blocks. However, most are poorly constrained, and we have few data from which net displacements can be obtained. Recent structural analysis has shown that most of the F_1 ductile structures have associated low-angle faults parallel or subparallel to the F_1 axial plane. These faults are folded by F_2 events, which constrains the time of faulting to pre-F_2 or early syn-F_2. The low-angle faults are contractional, placing Cambrian units over upper Paleozoic rocks. It is difficult to show the direction of

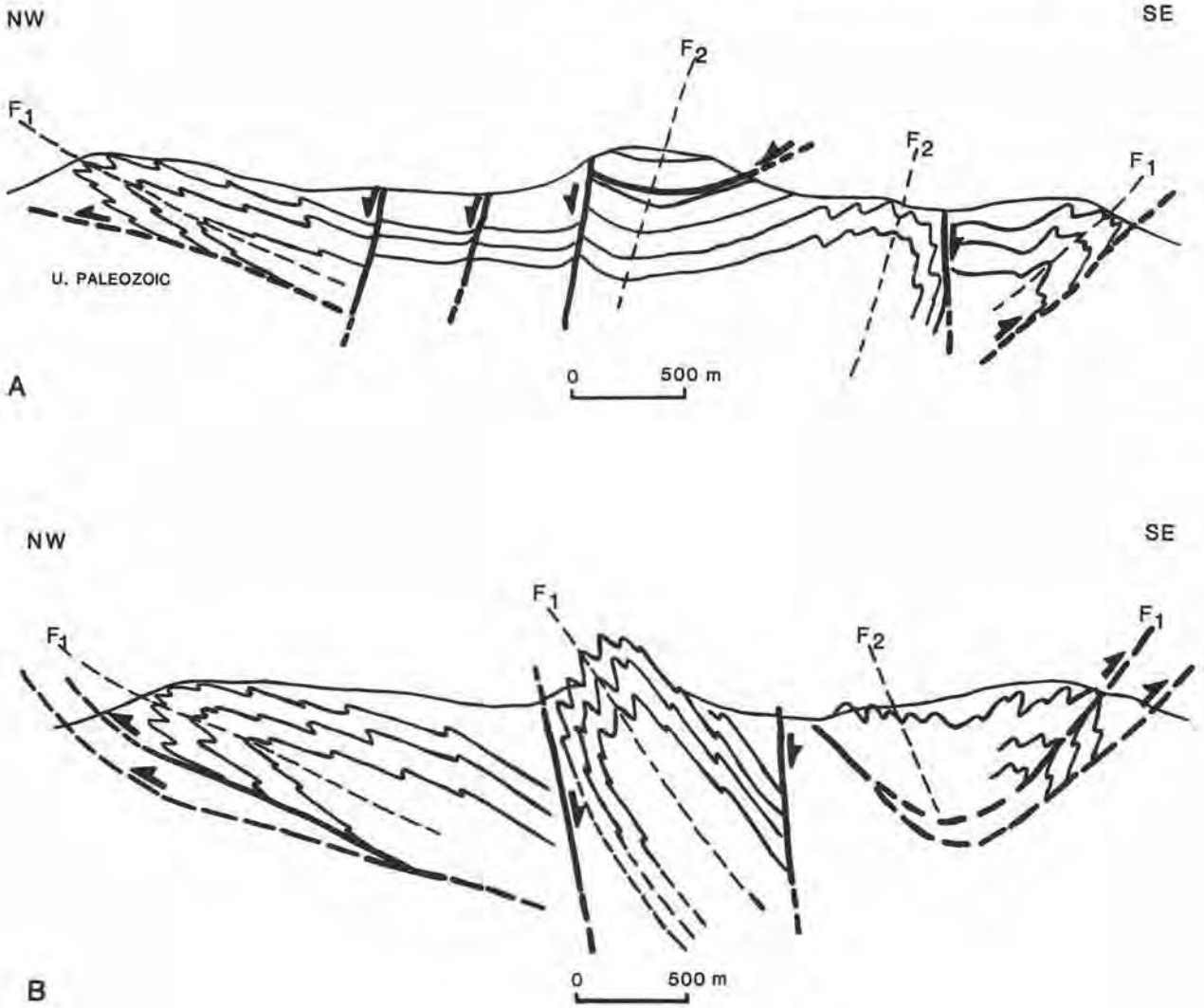

Figure 2.–A. A sketch structural cross section through China Bend, synthesized from detailed cross sections by Bressler (1979), Moser (1978), Phillips (1979), and Conklin (1981). F_1 folds, facing both northwest and southeast, have been refolded by steep F_2 folds and cut by later normal faults. B. A sketch structural cross section through the Cedar Lake area and Red Top Mountain, 16 km east of Northport along Deep Creek, synthesized from detailed cross sections by Haviland (1983) and O'Keefe (1980). Oppositely facing F_1 folds are cut by later, steep, normal(?) faults.

thrust transport in all locations, although at Red Top Mountain the thrusts can be seen to cut up-section in the facing direction (southeast) of the F_1 folds. We assume that all observed thrusts displace material in the facing direction of the host fold.

Two low-angle faults place younger rocks over older rocks, at Red Top Mountain and China Bend. The low-angle fault at China Bend (Moser, 1978; Phillips, 1979) places Metaline over Maitlen and is associated with a double-facing F_1 fold structure (Fig. 2); in this context, that fault is likely extensional, particularly if the section at China Bend is to be balanced. The low-angle fault at Red Top Mountain (O'Keefe, 1980) may also be extensional, but it may be a contractional fault through a sequence that was previously folded isoclinally, making the anomalous juxtaposition possible.

The dominant trend of later steep faults is from NNW to NNE; many of the faults are associated with Eocene lamprophyre dikes. Most of these faults appear to be extensional and clearly cut the main metamorphic fabric. The China Bend block, for example, is in part a horst, with down-dropped Metaline on the east side and a series of well-exposed normal faults on the west side of the cliff section by the river.

The steep faults may have a significant rotational component: some of the fault-bounded blocks show a pervasive tectonite fabric which trends more westerly

Table 1.–Correlation of units within the Maitlen Phyllite (after Conklin, 1981). The descriptions are field terms only. The total thickness of the Maitlen Phyllite is estimated by Park and Cannon (1943) to be 1,500 m in the Metaline quadrangle, east of these areas

China Bend (Bressler, 1979)	Red Top Mountain (O'Keefe, 1980)	Northport (Nordstrom, 1972)	Deep Creek (Yates, 1976)	Flat Creek (Conklin, 1981)	Gorge Mtn. (Phillips, 1979)	Crown Creek-Bowen Lake (Phillips, 1979)	Bowen Lake (Moser, 1978)
Red phyllite 60 m	Phyllitic quartzite >35 m		Phyllitic schist 60 m	Upper phyllite >80 m	Upper phyllite 101 m	Upper phyllite 85 m	Quartzite 15-33 m
Laminated quartzite >50 m	Gray limestone with lenticular pods of calcite 28 m		Unnamed limestone 30-150 m	Gray limestone with lenticular pods of calcite	Gray limestone with lenticular pods of calcite	Gray limestone with lenticular pods of calcite	Gray limestone with lenticular pods of calcite
Limestone with blue-gray top stratum 2.7-11 m	Limestone 45-160 m	Limestone 17 m		Micaceous quartzite 60 m	Micaceous quartzite 56 m	Micaceous quartzite 20 m	Upper phyllitic quartzite 12-24 m
Green phyllite 4-58 m	Quartzite-phyllite 25-150 m	Quartzite-phyllite 50 m	Phyllitic schist 30 m	Limestone with blue-gray top stratum 15-20 m	Limestone with blue-gray top stratum 14 m	Limestone with blue-gray top stratum 50 m	Limestone with blue-gray top stratum 15-43 m
Reeves Limestone Member 8-60 m	Reeves Limestone Member 30-90 m	Reeves Limestone Member 13 m	Reeves Limestone Member 15-30 m	Lower phyllite 72 m	Lower phyllite 65 m	Lower phyllite 115-136 m	Lower phyllite 12-24 m
TOTAL: 125-239 m	TOTAL: 179-479 m	TOTAL: 130 m	TOTAL: 135-270 m	TOTAL: 289-294 m	TOTAL: 295 m	TOTAL: 300-321 m	TOTAL: 108-228 m

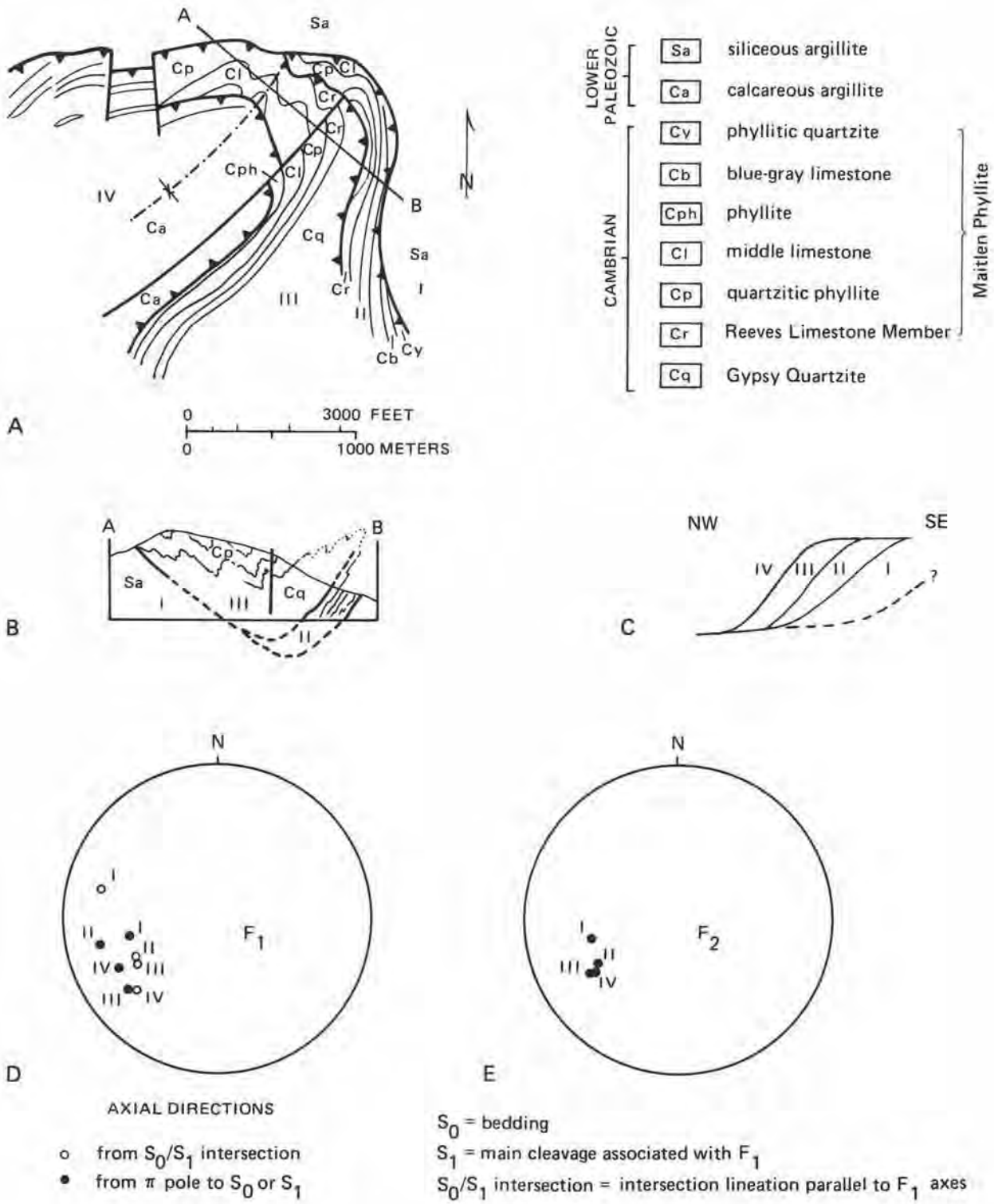

Figure 3.–Structural geology of Red Top Mountain (after O'Keefe, 1980). A. Geologic map, showing thrusts cutting up-section toward the southeast; B. Cross section (present-day); C. Diagrammatic cross section (pre-F_2); D. Orientation of F_1 features showing a general clockwise change from domain IV to domain I; E, Orientation of F_2 features.

Table 2.–Summary comparison of three areas along the Kootenay arc

Inchelium	Columbia River	Kootenay and Duncan Lakes
Typical eugeosynclinal sediments lower Paleozoic	(not found)	(not found)
F_1 folds and associated thrusts face east[1]	F_1 folds and associated thrusts face both east and west	F_1 folds face west[2,3]
F_2 folds upright[1]	F_2 folds steeply inclined toward the east and west	F_2 folds isoclinal and overturned toward the east[2]
No direct evidence of mid-Paleozoic orogeny	No evidence of mid-Paleozoic highland	Some mid-Paleozoic conglomerates (significance equivocal)[4,5]
Pre-F_1/S_1 porphyroblasts in Precambrian units	Pre-F_1/S_1 biotite porphyroblasts in lower Paleozoic (date unknown)	Small Devonian pluton in Shuswap[6] complex

Sources: [1]Snook and others (1981) [2]Hoy (1977) [3]Fyles (1962, 1967, 1970) [4]Wheeler (1967) [5]Reesor and Moore (1971) [6]Okulitch and others (1975)

than the regional trend. For example, the fault block on the west side of Flat Creek at China Bend has F_1 and F_2 lineations trending more westerly than those within adjacent blocks to the east. One interpretation is that some of the fault-bounded blocks may have rotated 15° to 20° clockwise with respect to the regional Kootenay trend. The rotated fault blocks mapped to date do not show any systematic change in orientation. Lamprophyre dikes are associated with some of these steep faults; the general chemical affinity of these dikes to the Eocene Coryell batholith in southern British Columbia (Conklin, 1981) suggests that extensional movement and possible rotation took place during the Eocene. This idea is supported by the general Eocene clockwise rotation across major faults in northeastern Washington and southeastern British Columbia recognized by Price (1979) and Fox and Beck (1985). However, the same dikes are also seen to use earlier structural grains (for example, S_1 surfaces and a-c joints); thus the significance of their association with the steep faults described above is somewhat equivocal.

A systematic rotation of F_1 axial lineations is, however, recognized within the imbricated sequence at Red Top Mountain (Fig. 3), where younger imbricates show more westerly orientations (Fig. 3D). F_2 orientations are constant in the top three thrust slices, whereas in the lowest, most easterly (youngest?) slice F_2 is clockwise of earlier F_2 orientations (Fig. 3E). This systematic westerly change may indicate that the arcuate form of the Kootenay arc developed during the formation of these structures and is not due to later bending (Ellis and Watkinson, 1984).

In summary, the structural sequence is that of major F_1 folds and associated thrust faults coaxially refolded by upright (to slightly overturned) F_2 folds. Minor localized folds refolded all earlier structures. High-angle extensional faulting followed, and clockwise rotation may have taken place in Eocene time.

REGIONAL TECTONIC SETTING

We attempt here to place this northern Washington section of the Kootenay arc in a tectonic setting compatible with that of its neighboring regions. Table 2 summarizes the important differences and similarities between this section of the arc and those farther north (near Kootenay and Duncan Lakes) and south (near Inchelium).

Debate surrounds the existence and extent of the so-called Antler orogeny. Snook and others (1981) suggested that the east-facing folds and associated thrusts near Inchelium are "Nevada" in style (that is, of Antler polarity). However, the age of these structures is poorly constrained; they can only be said to have developed sometime after the early Paleozoic. Greenman and others (1977) have described a sequence of Silurian(?) and Devonian siltstones, conglomerate, and fossiliferous limestones from northern Pend Oreille County which is believed to reflect local uplift and rapid sedimentation. Although there is no structural evidence of mid-Paleozoic orogeny in the

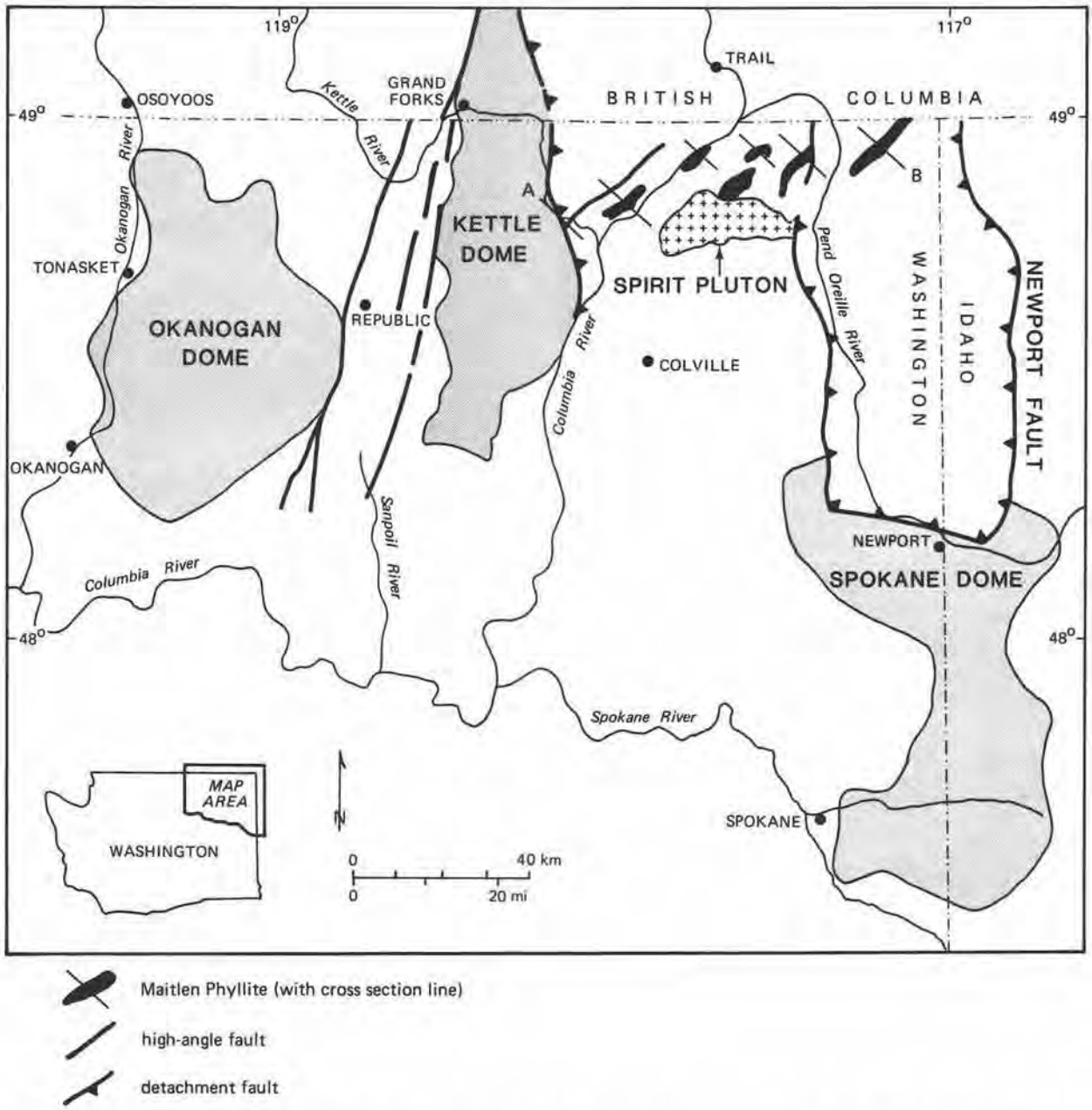

Figure 4.–Generalized geologic map of northeastern Washington, showing outcrops of the Maitlen Phyllite south of the U.S.–Canadian border and north of the Spirit pluton (Fig. 1). Cross sections through the Maitlen outcrops (between A and B) are shown in Figure 5.

northern part of the arc, two geological features are suggestive of an orogeny of Antler age: first, a Devonian pluton within the Shuswap complex (Okulitch and others, 1975); and second, a mid-Paleozoic horizon of conglomerates, some of which are reported to contain clasts of previously metamorphosed and deformed rock types, and others suggest at best only a gentle (disconformity-producing) tilting of the shelf (Wheeler, 1967; Reesor and Moore, 1971).

The region described in this paper has yielded no conclusive evidence of an Antler event. The structural sequence appears to involve rocks as young as Triassic in the Kettle Falls area (J. W. Mills, personal commun., 1985), albeit the orientations of F_1 isoclinal folds tend to be less consistent (southeast, north, and northwest). There is some evidence of a thermal event prior to the full development of F_1 and S_1 structures in the Columbia River region. Large porphyroblasts of

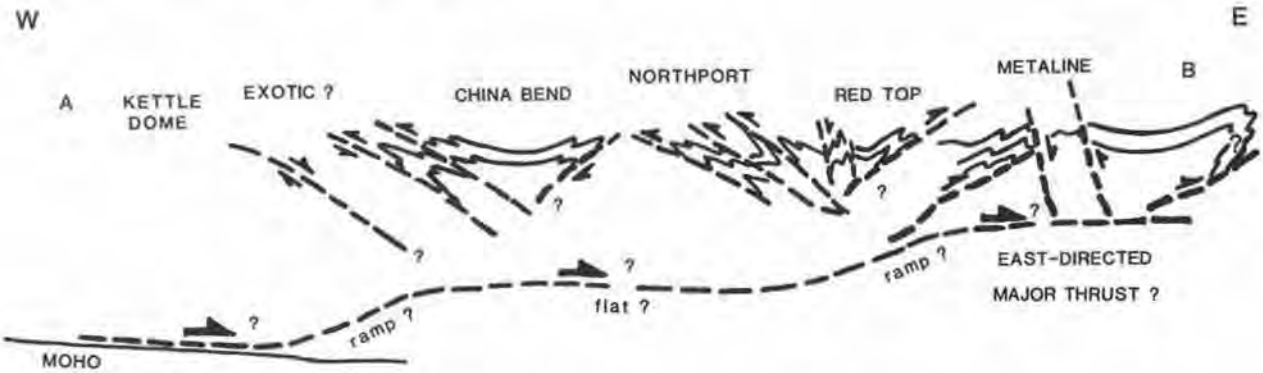

Figure 5.–Composite, schematic structural cross section through the Maitlen Phyllite outcrops in the Kootenay arc south of the U.S.–Canadian border (lines of sections shown on Fig. 4). All deep structure is hypothetical and inferred from surface data.

biotite clearly pre-date the main S_1 fabric in the China Bend area; younger biotites which define S_1 are easily distinguished by their texture and size (Ellis, 1984). Chlorite pseudomorphs after biotite? porphyroblasts, also pre- or early-S_1, occur in Precambrian slates farther south, near Chewelah. In addition, two lamprophyre dikes near Northport are boudinaged within S_1 (Schriber, 1981). However, none of these features is dated, and it is unclear whether they represent immediate precursors of the probable mid-Jurassic deformation and metamorphism or significantly older events.

The generalized regional geologic map of northeastern Washington (Fig. 4) shows elongate Maitlen outcrops trending NNE-SSW; these outcrops are generally intercalated between younger rocks, suggesting that the Maitlen outcrops are structurally higher than in intervening areas. This map pattern, combined with the observed facing directions of the major F_1 structures in the Maitlen, is the basis of speculative regional tectonic models—speculative due to poor exposure and the complex overprinting of Eocene extensional events. Obviously, subsurface data would help to constrain the models further.

A model of thrust-style tectonics, with formation of triangular and "pop-up" zones (Fig. 5) possibly rising from a regional east-facing lower thrust zone, would explain the structural highs and the facing direction of the F_1 folds. Analogous structures are seen in a segment of the Himalayan thrust-belt cross section constructed by Coward and Butler (1985). Such pop-ups and triangular zones occur typically in thrust-belt regions (for example, in the foothills of Alberta; Price, 1981). Because all of the discussed folds are cored by Gypsy Quartzite, we presume the thrusts in this region have flats in the quartzite.

This model is basically one of plane strain, implying no major strain in the third dimension (or out of section), parallel to the strike of the arc. However, strain measurements show appreciable elongation parallel to the F_1 lineation, and the model should accommodate this significant observation. A further constraint is the implication of accretion and of northward translation of accreted terranes (Irving, 1979). A model of oblique underthrusting (Fig. 6A), that is, a component of northward-directed shear plus a component of east-west shortening, could produce both the oppositely facing F_1 structures and the stretching along the F_1 axial lineation. The doubly vergent F_1 folds could have been either produced above cryptic dextral shear faults trending NNE (positive flower structures [Lowell, 1985, p. 8]; Fig. 6B) or, more likely, formed in pervasive ductile shear zones involving both the cratonic margin and accreted rocks (Ellis and Watkinson, unpublished data). The F_2 structures (with their vertical-stretch fabric; Ellis, 1984) then would have been produced as emergent structures—emergent after the F_1 metamorphic peak on the late- or post-F_1 faults of the thrust-belt type shown in Figure 5.

Finally, there is the component of Eocene extension which overprints the penetrative earlier deformation. The Mesozoic compressional structures are preserved in an Eocene structural low. The Kettle dome detachment fault bounds the western margin of the arc, and the Newport detachment fault and the Priest River complex bound the eastern margin. Between, the deformation is characterized by the north-striking, steep normal faults with associated dikes, such as those at China Bend. The horst block at China Bend is an example of another component in creating the structural highs which expose the Maitlen.

The change from contractional to extensional faulting from mid-Jurassic(?) to Eocene means probable reactivation of older fault systems, as well as creation of new faults, and presumably some thinning of a previously thickened crust. As yet we have been unable to document any specific structural examples of

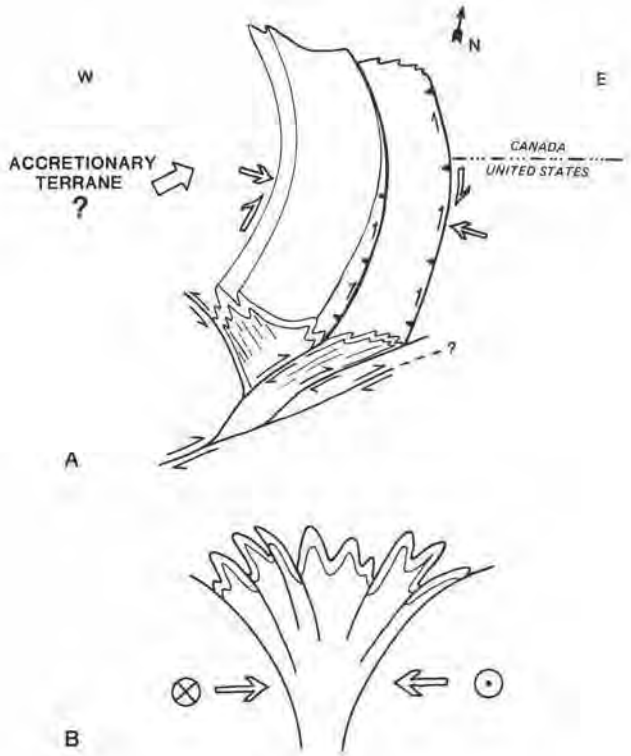

Figure 6.–A. Sketch of a tectonic model showing possible oblique underthrusting across the pre-accretion margin of the northwestern U.S. and southwestern Canada. B. A positive flower structure. Cross section of a dextral shear with shortening across the fault zone, illustrating how oppositely facing folds could form above a cryptic fault zone.

reactivation. Preliminary geophysical modeling (R. L. Thiessen, personal commun., 1986) suggests a crust 33 to 35 km thick. Preliminary COCORP data (Potter and others, 1985) suggest east-dipping extensional faults to the Moho, which cut the previous contractional, west-dipping thrust faults.

The nature of the boundaries between the Shuswap complex, Kettle dome, and the Kootenay arc is also equivocal. The boundary changes from an east-facing to a west-facing thrust along the western tongue of Kootenay Lake in Canada. In Washington, the western boundary of the rocks of the arc with the higher grade rocks of the Kettle gneiss dome is still not well understood. Clearly, a major fault (Kettle River fault) separates rocks of different metamorphic grades from Kettle Falls northward to the confluence of the Columbia and Kettle Rivers; Wilson (1981) shows a change in both dominant tectonite lineation direction and metamorphic grade across this fault. Farther north, the Kettle River fault is exposed along the Kettle River and is interpreted by Rhodes and Cheney (1981) to be a low-angle normal fault of Eocene age. Its low angle is confirmed by seismic evidence (Hurich and others, 1985). South of Kettle Falls, a fault zone—in places at a low angle with an associated mylonite zone, and in other places at a high angle—separates the predominantly Lower Ordovician Covada Group from the gneisses (K. F. Fox, Jr., personal commun., 1985). More dating is needed to determine how much, if any, of the Kootenay arc deformation is retained in the metamorphic assemblages west of the Columbia River. The Kettle dome gneisses and associated Colville batholith rocks are dominated by Eocene extension, thermal events, and mylonite formation (Fox and others, 1977; Atwater and others, 1984).

As far as we can tell, deformation in the Kootenay arc rocks of northeastern Washington is continuous and compatible with deformation associated with the possible docking of terranes in Canada (Monger and others, 1982) in the mid-Mesozoic. Clearly the structural style changes in detail along the western margin of the arc. How much of this change is due to local control of structural styles and how much is due to differences in tectonic events along the margin remains a topic of further research. Eocene deformation is observed as faulting; what other tectonic events occurred between the presumed Jurassic deformation and metamorphism and the Eocene faulting and thermal events—apart from the intrusion of Cretaceous plutons such as the Spirit pluton—is another important question yet to be answered.

ACKNOWLEDGMENTS

The contributions of the series of Master of Science theses from Washington State University has been invaluable for this synopsis. We thank Dr. Joe Mills for sharing his enthusiasm and detailed knowledge of the Kootenay arc in Washington with us. We also appreciate the annual discussion with Dr. Ken Fox, who is always stimulating and always makes us think twice—in addition to giving us the benefit of his experience and knowledge of the geology of the region.

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KETTLE AND OKANOGAN DOMES, NORTHEASTERN WASHINGTON AND SOUTHERN BRITISH COLUMBIA

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ABSTRACT

Kettle and Okanogan domes are structural culminations of predominantly upper amphibolite facies gneisses and plutonic rocks in the southern part of the Omineca crystalline belt in northeastern Washington and southern British Columbia. They are morphologically similar to metamorphic core complexes elsewhere along the axis of the North American Cordillera.

The age or ages of the rocks in the cores of Kettle and Okanogan domes are poorly known. Tentative lithologic correlations suggest that the paragneisses may, in part, be Proterozoic; scanty radiometric age data suggest late Mesozoic through Eocene ages for the plutonic rocks. K-Ar determinations throughout the crystalline terrane yield reset Eocene ages as young as 45 to 50 million years.

The gneisses were metamorphosed and highly deformed during the Jurassic. Metamorphism, folding, and, locally, mylonitization resulted from the mid-Jurassic obduction of the allochthonous, eugeoclinal rocks of Quesnellia over the continentally affiliated rocks of the domes.

Within Quesnellia the Chesaw thrust is a regional tectonic feature that places Triassic greenstone over late Paleozoic pelitic rocks.

Eocene extension produced the deeply rooted, oppositely verging, low-angle normal faults in the Kettle and Okanogan River valleys; these faults define a composite Okanogan-Kettle lens. Gently dipping, 1- to 2-kilometer-thick mylonitic zones capped by coplanar zones of chloritic breccia mark these faults. Eocene extension also generated a series of chloritic, brecciated, normal faults without coplanar mylonitization or other evidence of major (more than 10 kilometers) displacement.

The Eocene faults form most of the present boundaries of the domes, separating the crystalline rocks from the adjacent greenschist facies rocks and unmetamorphosed volcanic and sedimentary rocks of the Challis suite of Eocene age. Because the three major units of the Challis suite are bounded by unconformities and occur in structural lows adjacent to the domes, they probably were originally more regionally extensive and may predate the present structural relief of the domes.

A cause for this Eocene extension and the contemporaneous clockwise rotation evident in the Challis suite is speculative. One hypothesis relates them to dextral transcurrent faulting induced by northward translation of the Pacific plate with respect to North America.

INTRODUCTION

Kettle and Okanogan domes (Fox and others, 1976, 1977; Cheney, 1980; Cheney and others, 1982a; Rhodes and Cheney, 1981) are structural culminations of predominantly upper amphibolite facies metamorphic rocks and associated intrusive rocks in the Okanogan[1] highlands of northeastern Washington and southern British Columbia. Together with the Spokane dome (Cheney, 1980) or Priest River complex (Rehrig and others, this volume; Rhodes and Hyndman, 1984) and the recently identified Lincoln dome (Atwater and others, 1984), the Kettle and Okanogan domes form the southern extension of the Omineca crystalline belt into Washington (Fig. 1). The domes are composed of highly deformed sillimanite-grade metamorphic rocks and associated, variously foliated and unfoliated intrusive igneous rocks. The domes show apparent structural relief of 2 to 3 km relative to the adjacent lower grade and unmetamorphosed strata.

Structural features morphologically similar to Kettle and Okanogan domes, termed metamorphic

[1] The Canadian spelling is Okanagan; the American spelling is used throughout this paper.

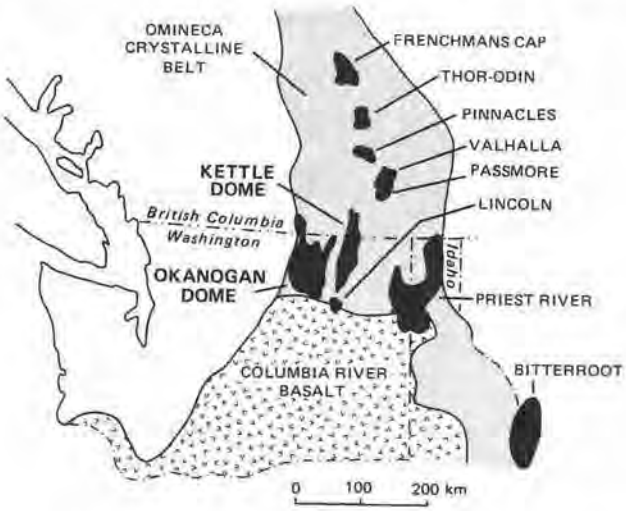


Figure 1.–Location of metamorphic core complexes in the northwestern United States and southern British Columbia.

core complexes (Crittenden and others, 1980), occur in a discontinuous chain along the axis of the North American Cordillera from Mexico to the Yukon Territory. Typically, these core complexes consist of a crystalline core of plutonic and metamorphic rocks of various ages and structural histories. A gently dipping, 1- to 4-km-thick, mylonitic zone commonly exists at the structural top of the crystalline core, although in some complexes the mylonitic front is not coincident with the core margin. The most diagnostic feature of the core complexes is the abrupt structural discontinuity separating the crystalline, ductilely deformed core from a brittlely extended cover sequence. This discontinuity is a low-angle, extensional or detachment fault, marked by a relatively thin (100-200 m) zone of intense brecciation and chloritic alteration.

Despite the broad morphologic similarities among core complexes, significantly diverse models for their formation have been proposed. Particular points of contention include: (1) the timing of mylonitization and its possible genetic link to detachment faulting, (2) the relative importance of extension versus compression in the evolution of the complexes, especially with respect to the mylonitic deformation, and (3) the process and timing of doming or uplift of the complexes relative to detachment faulting. Geographically, Kettle and Okanogan domes lie between the dome-like features of the Shuswap terrane, described as Mesozoic compressional features (Brown and Read, 1983), and the core complexes of the southwestern United States, thought to have evolved during major Tertiary extension (for example, Davis, 1983).

In this paper we describe the structural features of Kettle and Okanogan domes and summarize their setting within the regional geology of northeastern Washington. This paper supplements earlier publications of Cheney and his co-workers in that it presents a compilation map of the northern parts of the Kettle and Okanogan domes (north of lat. 48°30′ N.). The southern parts of the domes, as well as the Lincoln dome, are shown on a preliminary 1:100,000-scale map of the Colville Reservation (Atwater and others, 1984). This paper also updates Cheney (1980) and Cheney and others (1982b) by including recent work of Orr (1982, 1985). Detailed mapping is still largely lacking in this area; most of the mapping has been at scales of 1:40,000 or smaller. We do not discuss in detail Tertiary sedimentary and volcanic rocks adjacent to the domes. Little work has been done on them following the important regional synthesis of Pearson and Obradovich (1977). Finally, we integrate our ideas on the geologic history of Kettle and Okanogan domes with the evolving concepts of the tectonic evolution of the North American Cordillera.

ROCK UNITS

Three generalized packages of rocks are present in the region of the Okanogan and Kettle domes (Figs. 2 and 3): (1) pre-Jurassic (Proterozoic?) metamorphic cores and associated Mesozoic and Tertiary batholiths of the domes, (2) Carboniferous to Triassic heterogeneous, eugeoclinal rocks of the allochthonous terrane Quesnellia, and (3) Eocene volcanic and volcaniclastic strata of the Challis suite.

Rocks Within Core Complexes

The Tenas Mary Creek sequence of Kettle dome and the northeastern part of the Okanogan dome (Fig. 2) is a regionally continuous suite of quartzite, marble, and pelitic gneisses interlayered with two thick stratiform units of granodioritic orthogneiss. Fibrous sillimanite is widespread as a stable phase in the pelitic rocks. The potassic phase is potassium feldspar; muscovite is absent except as a postkinematic retrograde mineral (Rhodes, 1980; Wilson, 1981b), indicating that metamorphism was within the upper sillimanite zone of the amphibolite facies. Within Kettle dome many different local names have been applied to these metamorphic rocks (summarized by Wilson, 1981b, p. 9-10); however, Cheney (1980) showed that a coherent stratigraphy of orthogneisses and paragneisses is continuous throughout the dome and that the rocks of Kettle dome are correlative with the Tenas Mary Creek sequence described by Parker and Calkins (1964) in the northeastern arm of Okanogan dome.

Lithologies of the units in the Tenas Mary Creek sequence were described in detail by Parker and Calkins (1964) and Preto (1970). Figure 2 differs from earlier maps published by Cheney (1980; Cheney and others, 1982b) in that we have adopted the reinterpretation of Orr (1982) for the structure of the northern part of Kettle dome and have correlated the rocks of Kettle dome with the equivalent rocks of the Grand Forks Group, mapped by Preto (1970) in British Columbia. Specific correlations are listed in Table 1.

A useful stratigraphic marker unit in the Tenas Mary Creek sequence is a massive feldspathic quartzite (unit q of Fig. 2). Correlation of this unit with Unit II of Preto (1970) is based on several criteria (Orr, 1982). Both are massive, rusty-weathering, medium-grained quartzites that include 5 to 10 percent equant, milky-white potassium-feldspar grains. This quartzite lacks the micaceous partings characteristic of other quartzites in the Kettle dome (specifically, unit eq and, to a lesser extent, quartzite layers in units bu and bl of Fig. 2). In the Togo Mountain quadrangle just south of the International Boundary (Pearson, 1977) and in the Grand Forks area (Preto, 1970), the top of the pelitic unit that underlies the feldspathic quartzite is dominated by marble. Similarly, in the northeast arm of the Okanogan dome, Parker and Calkins (1964) mapped a 240-m-thick marble unit underlying the quartzite. On the other hand, the feldspathic quartzite in the Grand Forks area is generally thinner than it is in the United States, and the units are not directly continuous across the border.

Another previously indefinite lithologic correlation across the 49th Parallel was that of the Cascade granodioritic gneiss in British Columbia with one of the two granodioritic orthogneisses in the Tenas Mary Creek sequence (units g and gppg of Fig. 2). Daly (1912) first mapped moderately foliated granodioritic rocks near Grand Forks as his Cascade gneissic batholith. Bowman (1950) named similar rocks south of the International Boundary the Cascade granodiorite. Both Daly and Bowman felt that these rocks were younger than and intrusive into the adjacent gneisses. Later workers (Preto, 1970; Rhodes, 1980; Donnelly, 1978; Wilson, 1981b; Cheney, 1980) have shown that the Cascade granodiorite is an orthogneiss having the same deformational history as the surrounding gneisses.

Along the eastern margin of the Kettle dome, near Cascade, B. C., the Cascade gneiss and unit g are continuous across the International Boundary (Orr, 1982). In addition, all outcrops of Cascade gneiss that we have examined in the Grand Forks area closely resemble unit g in mineralogy and texture. All are moderately well foliated, biotitic, granodioritic gneisses, completely lacking the pegmatitic zones and feldspar megacrysts characteristic of unit gppg.

The northern part of Okanogan dome consists primarily of two suites of rocks (Fig. 2). The Tonasket gneiss in the west has been described alternatively as layered paragneiss (Snook, 1965; Fox and others, 1976; Goodge and Hansen, 1983) or as orthogneiss derived from a compositionally zoned pluton (Cheney, 1980). Weakly to variably foliated biotitic quartz monzonitic to granodioritic plutons intrude the gneisses in the eastern part of the dome. The western contacts of the three major plutons shown in Figure 2 dip gently eastward, and the major phases of each pluton are texturally and mineralogically similar. Thus, prior to erosion, these plutons may have been a single, crudely tabular body (Cheney, 1980, and unpub. reconnaissance mapping).

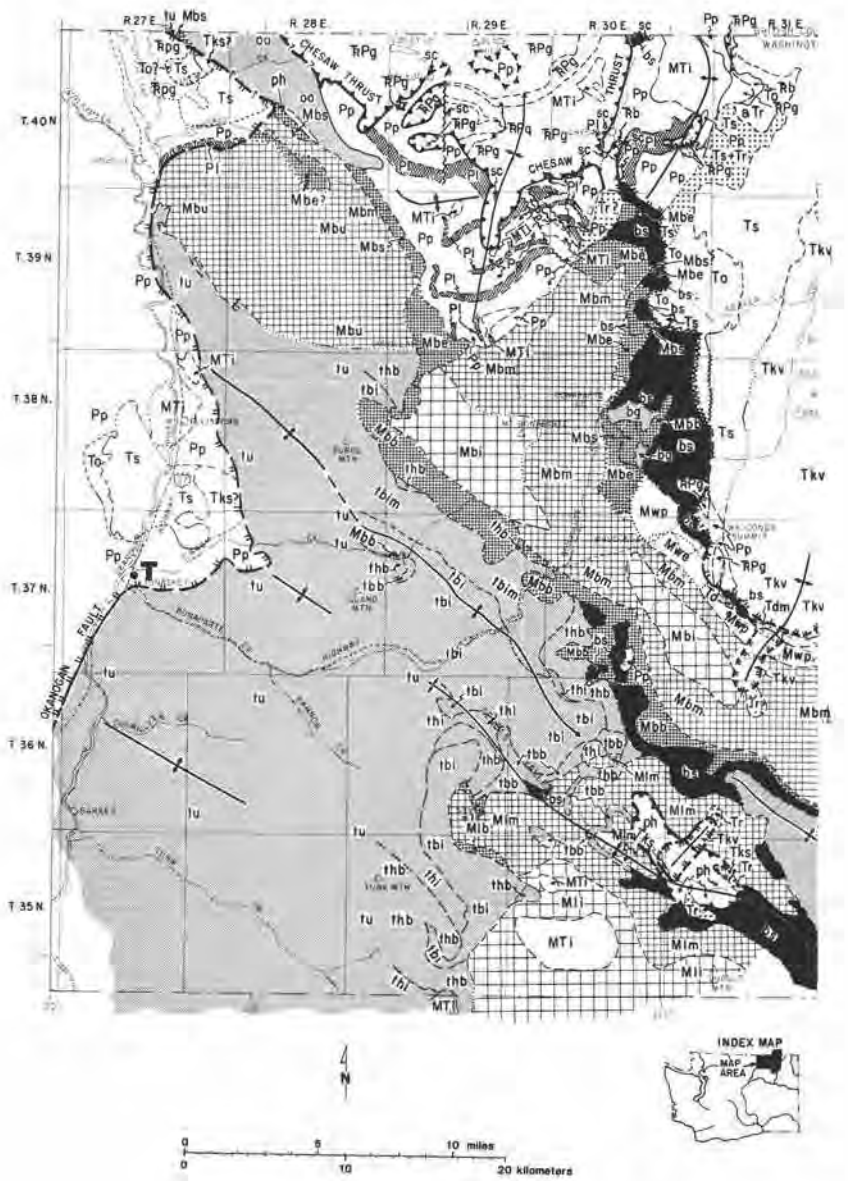
Throughout most of Okanogan and Kettle domes, the Tonasket gneiss and rocks of the Tenas Mary Creek sequence are within the upper amphibolite facies, and high metamorphic grade has been implicitly used in defining the domes. In the Tenas Mary Creek area, however, Orr (1985) confirmed the gradual transition to lower metamorphic grade in the Tenas Mary Creek sequence originally mapped by Parker and Calkins (1964). No apparent postmetamorphic tectonic discontinuity disrupts this sequence. Lower amphibolite facies schist and phyllite also occur in the Beaver Creek area (Pearson, 1967; Orr, 1985), in the Swan Lake area southwest of Republic, and in the Togo Mountain area south of Grand Forks (Pearson, 1977; Orr, 1985). Whether all of these lower grade rocks are correlative is indeterminate. In some areas they are apparently separated from high-grade rocks by low-angle faults, presumably postmetamorphic, but pre-Eocene in age (Orr, 1985).

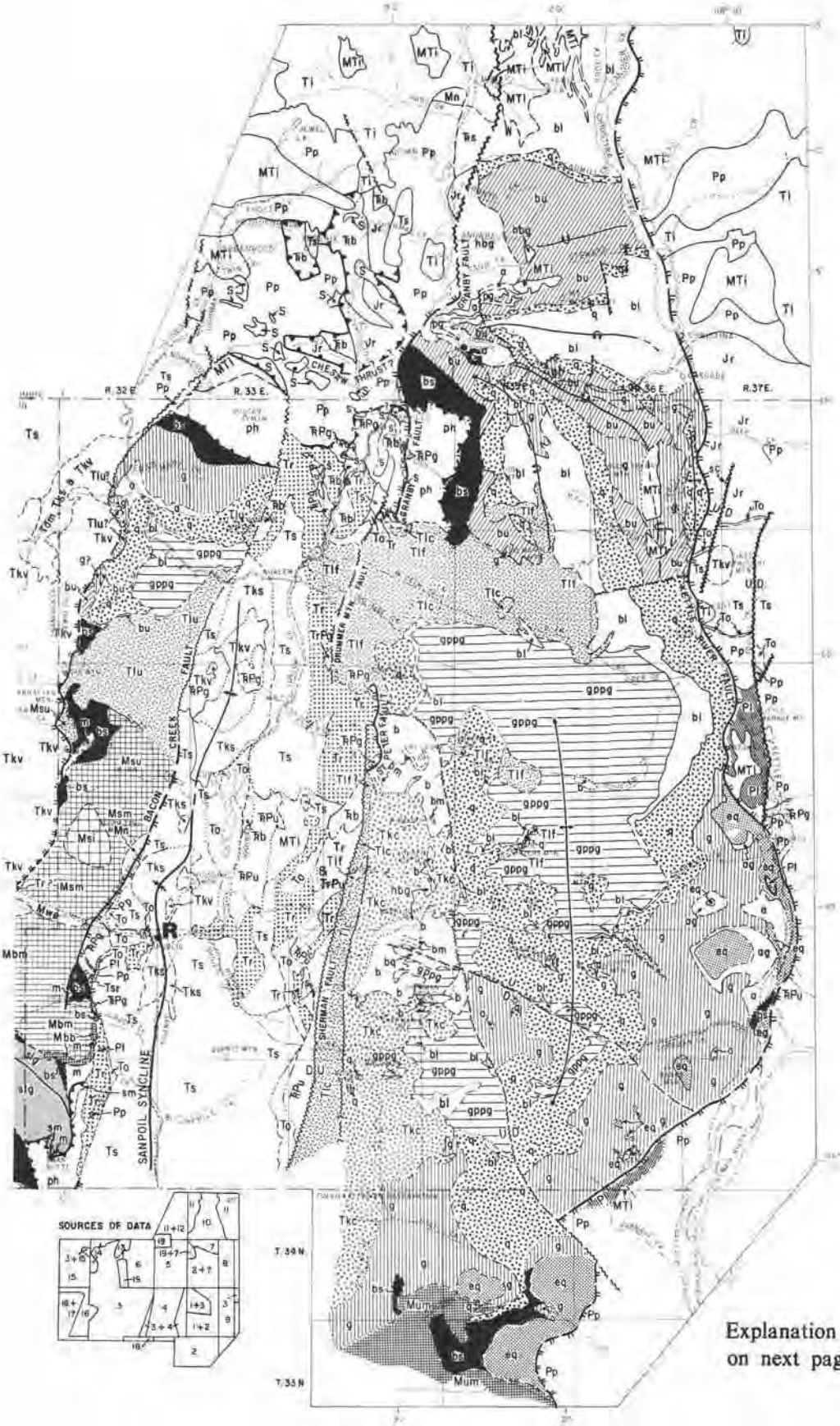
K-Ar age determinations throughout the Okanogan and Kettle domes yield slightly discordant Eocene ages. Clearly, for the metamorphic rocks these are not primary ages. Other than K-Ar analyses, only very preliminary dating has been done for the Tenas Mary Creek sequence. Rb-Sr and U-Pb (zircon) analyses of rocks from several localities in the orthogneissic sheets yield probable late Mesozoic magmatic ages with a component of older (Precambrian) inherited zircon (R. L. Armstrong, personal commun., 1985). The ggpg unit of Figure 2 is texturally similar to the megacrystic granodioritic Newman Lake orthogneiss of the Spokane dome of the Priest River complex; radiometric dating of the Newman Lake orthogneiss indicates that it is Cretaceous (Armstrong and others, this volume; Bickford and others, 1985).

A similar paucity of radiometric ages shrouds the age of the Tonasket gneiss of the Okanogan dome; Fox

Figure 2.–Preliminary geologic map of the northern part of the Okanogan and Kettle metamorphic core complexes.

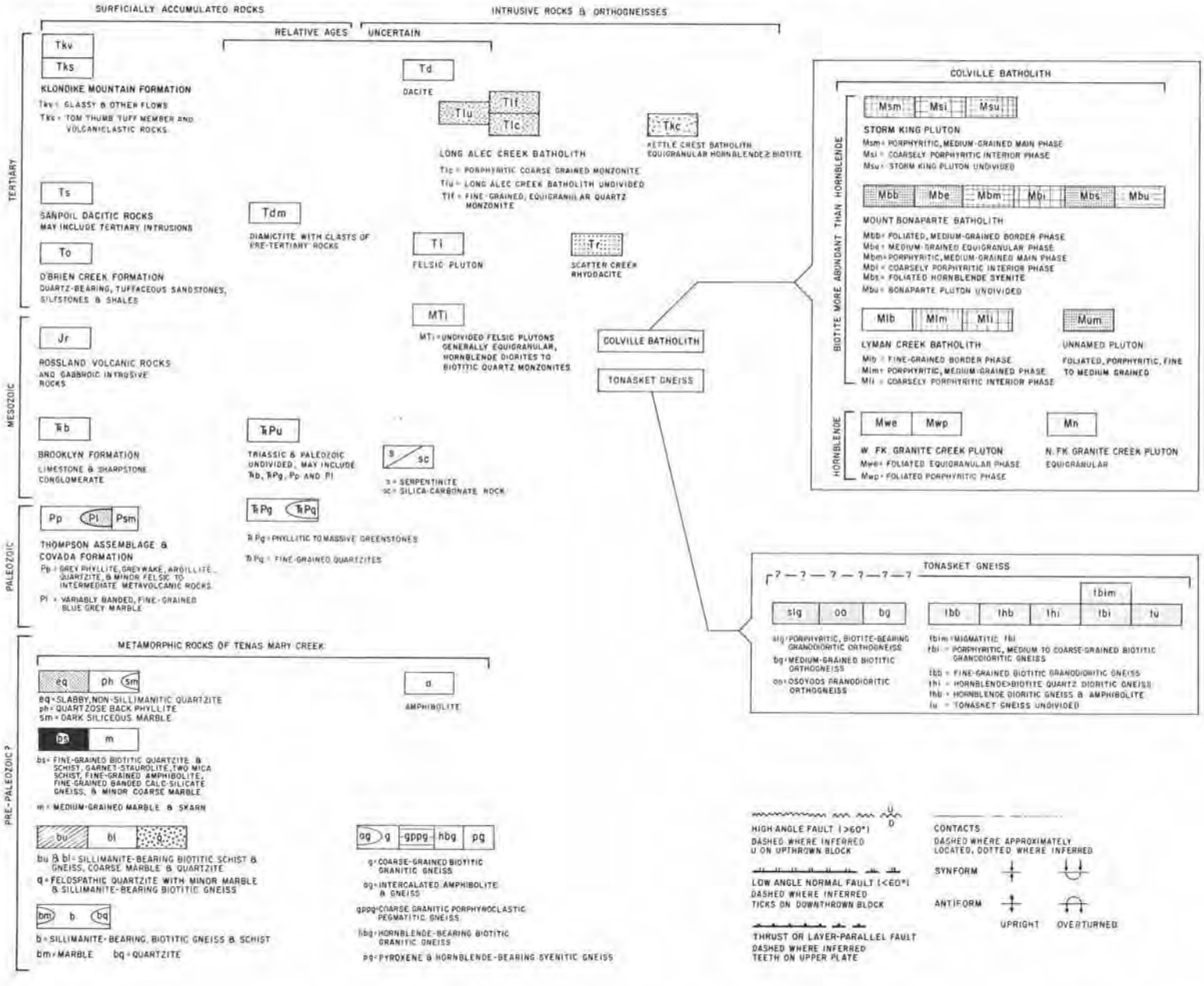
Sources of data: (1) Campbell and Thorsen, 1975; (2) Cheney, 1980; (3) Cheney, unpublished mapping; (4) Muessig, 1967; (5) Parker and Calkins, 1964; (6) Pearson, 1967; (7) Pearson, 1977; (8) Rhodes, 1980; (9) Wilson, 1980, 1981a; (10) Preto, 1970; (11) A. V. Okulitch and G. W. Woodsworth, unpublished data; (12) Little, 1983; (13) McMillen, 1979; (14) Fox, 1978; (15) Fox, 1970; (16) Fox and Rinehart, 1972; (17) Snook, 1965; (18) Staatz, 1964; (19) Orr, 1985.





Explanation for Figure 2 on next page.

Explanation for Figure 2 (See map, previous facing pages)



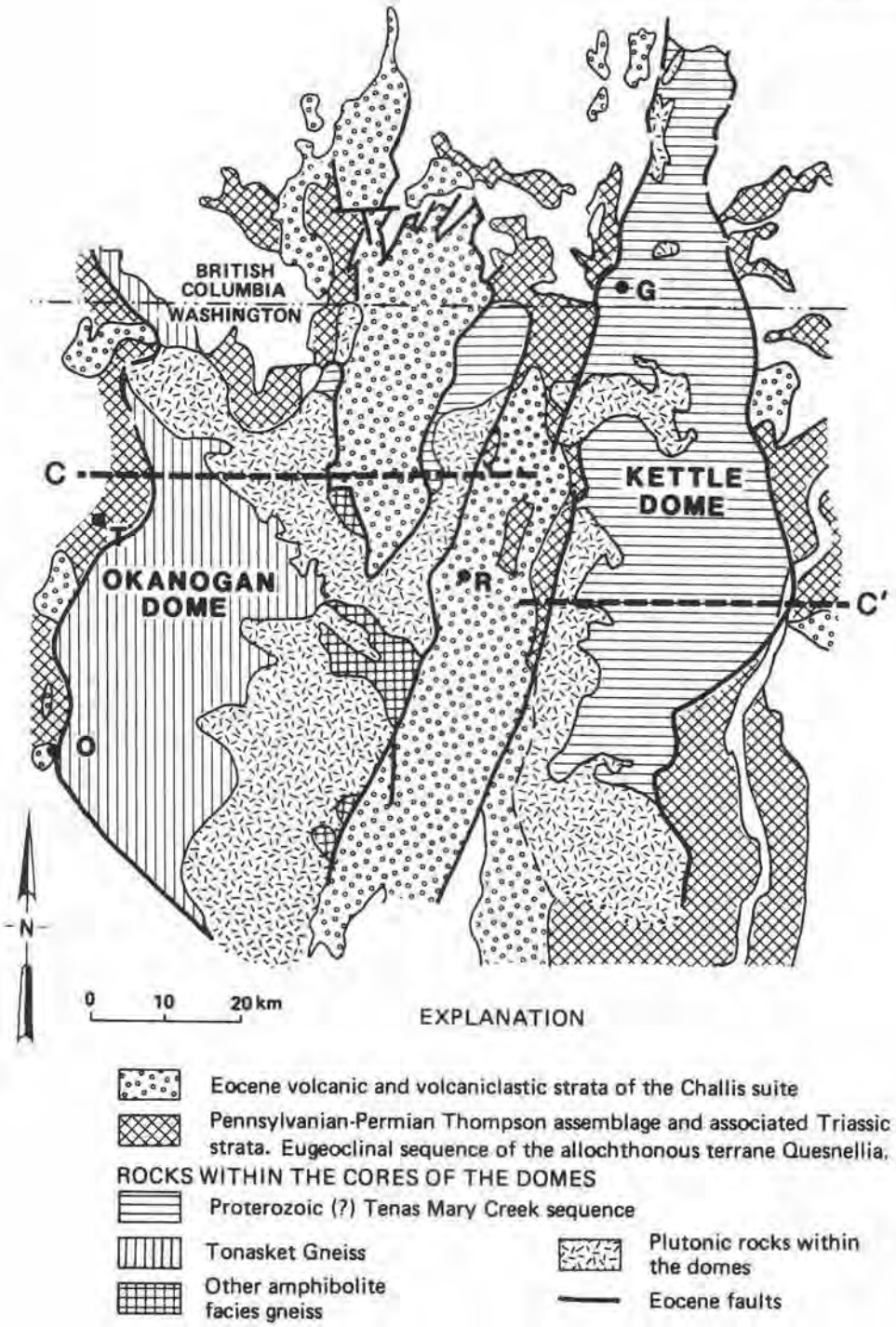


Figure 3.– Major lithologic assemblages of Okanogan and Kettle domes. C - C′, location of cross section in Figure 5. T, Tonasket; O, Omak; R, Republic; G, Grand Forks.

and others (1976) reported a Cretaceous age based on U-Pb (87 and 100 m.y.) and Th-Pb (94 m.y.) analyses on zircons in one sample from a locality 16 km northeast of Tonasket. On Figure 2, the Tonasket gneiss is tentatively shown as Mesozoic. Parkinson (1985) obtained a U-Pb age of 210 m.y. from zircon for the Osoyoos orthogneiss in the Okanogan dome near the International Boundary.

The regional continuity and quartzose nature of the metasedimentary rocks within Okanogan and Kettle domes suggest an affinity with the North American craton—either as part of the craton itself or as part of

Table 1.–Regional correlation of metamorphic rocks of the Tenas Mary Creek sequence (from Orr, 1982)

This Report (modified from Cheney, 1980)	Tenas Mary Creek (Parker and Calkins, 1964)	Grand Forks Group (Preto, 1970)	Laurier and Orient quadrangles (Rhodes, 1980; Rhodes and Cheney, 1981)	Togo Mountain quadrangle (Pearson, 1977)
eq	absent	absent	absent	absent
pg	absent	VIII	absent	absent
hbg	absent	IX and X	absent	absent
bs, m	sch	V	absent	sch
a	hs	IV	am	am
g	qgn	VII "Cascade gneiss"	ggn	part of ggn
bu	part of qu	III	bgn III, ma III	part of bgn
q	qu	II	qt III	qt and qb
bl	ma, part of qu	part of I	qt I, ma II, qt II, bgn I	parts of bgn and ma
gppg	ggn	absent	pggn	part of ggn
b, bm, bq	absent	part of I	absent	absent

the continentally derived miogeocline. This inference is supported by the data of Armstrong and others (1977), which indicate that the margin of the pre-Paleozoic North American craton (as indicated by initial $^{87}Sr/^{86}Sr$ ratios of $\geqq 0.074$ for Mesozoic and Cenozoic intrusions) lies to the west of the Okanogan dome. Specific correlation of the paragneissic rocks with sequences elsewhere in the Cordillera is difficult because the rocks in the domes are highly metamorphosed and deformed and cannot be traced out of the core into lower grade equivalents.

The thick quartzite in the Tenas Mary Creek sequence provides a starting point for speculative correlations. Potential correlatives of this quartzite include (1) quartzite described by Weis (1968) in an area south of Spokane (in an area termed the Priest River complex by Rehrig and others [this volume]), (2) quartzite described by Read (1979, 1980) in the mantling gneiss sequence in Thor–Odin and Frenchman's Cap domes in the Monashee terrane in British Columbia—possibly correlative with part of the Proterozoic Belt–Purcell Supergroup, and (3) the Lower Cambrian Gypsy Quartzite (Yates, 1971). The Gypsy is equivalent to the Hamill Formation in British Columbia.

Cheney (1980) proposed a correlation between the Tenas Mary Creek sequence and metamorphic rocks of the Priest River complex. Radiometric dates of more than 1,450 m.y. indicate that the paragneisses of the Priest River complex are Proterozoic (Armstrong and others, this volume). Orr (1985) favored a correlation with the Monashee mantling gneisses based on the overall heterogeneity of both sequences and a general correspondence of lithologies. Either alternative implies a Proterozoic protolith for the paragneisses of the Tenas Mary Creek sequence.

Rocks Adjacent to the Core Complexes

Thompson Assemblage

Thompson assemblage is a general term proposed by A. V. Okulitch (*in* Monger, 1977) for patchily exposed late Paleozoic through early Mesozoic, predominantly pelitic, eugeoclinal rocks in the Okanogan region. Locally mapped and named, their detailed stratigraphy is unknown, and specific correlations between localities are uncertain (Monger, 1977; Peatfield, 1978). In Washington these rocks have been variously called the Anarchist Group, Covada Group, Churchill

Mountain, and Mount Roberts Formation(s). The Thompson assemblage is part of the large allochthonous terrane, Quesnellia, thought to have been accreted to North America in mid-Jurassic time (Monger and others, 1982).

Paleontological age control for the Thompson assemblage is meager in the Kettle-Okanogan area. Localities reported by Parker and Calkins (1964), Fox and others (1977), McMillen (1979), and Little (1983) yield ages from middle Pennsylvanian to Triassic. The Covada Group southeast of Kettle dome has yielded Ordovician fossils (Snook and others, 1981) and is, therefore, no longer considered part of the Thompson assemblage (Silberling and others, 1984). Nevertheless, because all of these low-grade pelitic rocks are indistinguishable in the field, they are lumped as unit Pp on Figure 2.

The Thompson assemblage is overlain by the Permian and Triassic greenstones and sharpstone conglomerates of the Brooklyn Formation (Peatfield, 1978). A distinctive clastic limestone unit differentiates the Brooklyn Formation from the underlying Paleozoic sequence. Locally, the Jurassic Rossland greenstones and volcaniclastic rocks unconformably overlie the Brooklyn and equivalent rocks.

Read and Okulitch (1977) described the top of the Paleozoic sequence as a major regional unconformity. North of Okanogan dome, Cheney (unpub. data, 1975) and McMillen (1979) mapped the contact as the Chesaw thrust—a tectonic zone marked by ultramylonites and sheared lenses of serpentinite and magnesite. Ross (1981) also described the contact as tectonic near Kamloops, B.C. In the Conconully quadrangle, Rinehart and Fox (1976) mapped ultramafic rocks along the upper contact of the Paleozoic sequence; however, they interpreted the contact as an unconformity. West of Kettle dome near the International Boundary, the contact has not been studied in any detail; however, lenses of serpentinite are associated with Paleozoic phyllites and greenstones (Parker and Calkins, 1964; Little, 1983), suggesting that the Chesaw thrust may be a regional tectonic feature (as shown on Fig. 2). Timing of movement on the Chesaw thrust is broadly constrained as post-late Triassic and pre-Cretaceous (McMillen, 1979).

Challis Suite

The Challis suite of Eocene age is a regionally extensive, tripartite sequence cropping out adjacent to the domes. In the United States, the parts of the suite were defined by Muessig (1967). Tuffaceous sandstone and quartz latitic pyroclastic deposits of the O'Brien Creek Formation are overlain by thick rhyodacitic flows (Sanpoil Formation). A heterogeneous unit composed of tuffs, flows, and volcaniclastic beds caps the sequence (Klondike Mountain Formation). Equivalent rocks in British Columbia are the Kettle River, Marron, White Lake, and Skaha Formations (Church, 1973; Monger, 1968). In the United States the maximum thicknesses and apparent ages of the formations are: O'Brien Creek, 1,300 m and 54.3 m.y.; Sanpoil, 2,400 m and 52.3 m.y.; and Klondike Mountain, 1,200 m and 47 m.y. (Pearson and Obradovich, 1977; Fox and Beck, 1985; ages corrected using the constants of Dalrymple, 1979).

The eruptive and depositional history of the Challis suite is not well understood. Unconformities occur between and within formations. Coarse conglomerates such as those along Washington State Route 20 in the valley of the West Fork of Granite Creek suggest that some deposition was fault controlled. Nevertheless, the presence of some or all of the tripartite package in several areas in northeastern Washington and adjacent British Columbia (Pearson and Obradovich, 1977) suggests that the present areas of outcrop may be preserved parts of formerly more regionally extensive units, rather than areas of local deposition (Cheney, 1980).

STRUCTURAL EVOLUTION OF THE CORE COMPLEXES

The rocks in the cores of Kettle and Okanogan domes display three macroscopic structural features (Fig. 4). The rocks have undergone complex deformation that accompanied metamorphism (Preto, 1970; Donnelly, 1978; Orr, 1982; Snook, 1965; Goodge and Hansen, 1983). Secondly, a penetrative, gently dipping, mylonitic fabric marks some margins of the domes. Brittlely deformed, chloritic faults of Eocene age (some coplanar with zones of underlying mylonite) form most of the present margins of the domes.

Regional Metamorphism

Despite the high degree of mesoscopic deformation, the metamorphic foliation parallels compositional layering in the gneisses and outlines the macroscopic structure in the domes. The thick feldspathic quartzite unit is a useful indicator of the internal structure in Kettle dome (Fig. 4). South of lat. 48°50′ N., the metamorphic rocks of Kettle dome outline a broad asymmetrical antiform that plunges gently to the south. Dips on the eastern limb (20°-40°) and the strike of the foliation roughly parallel the margin of the dome.

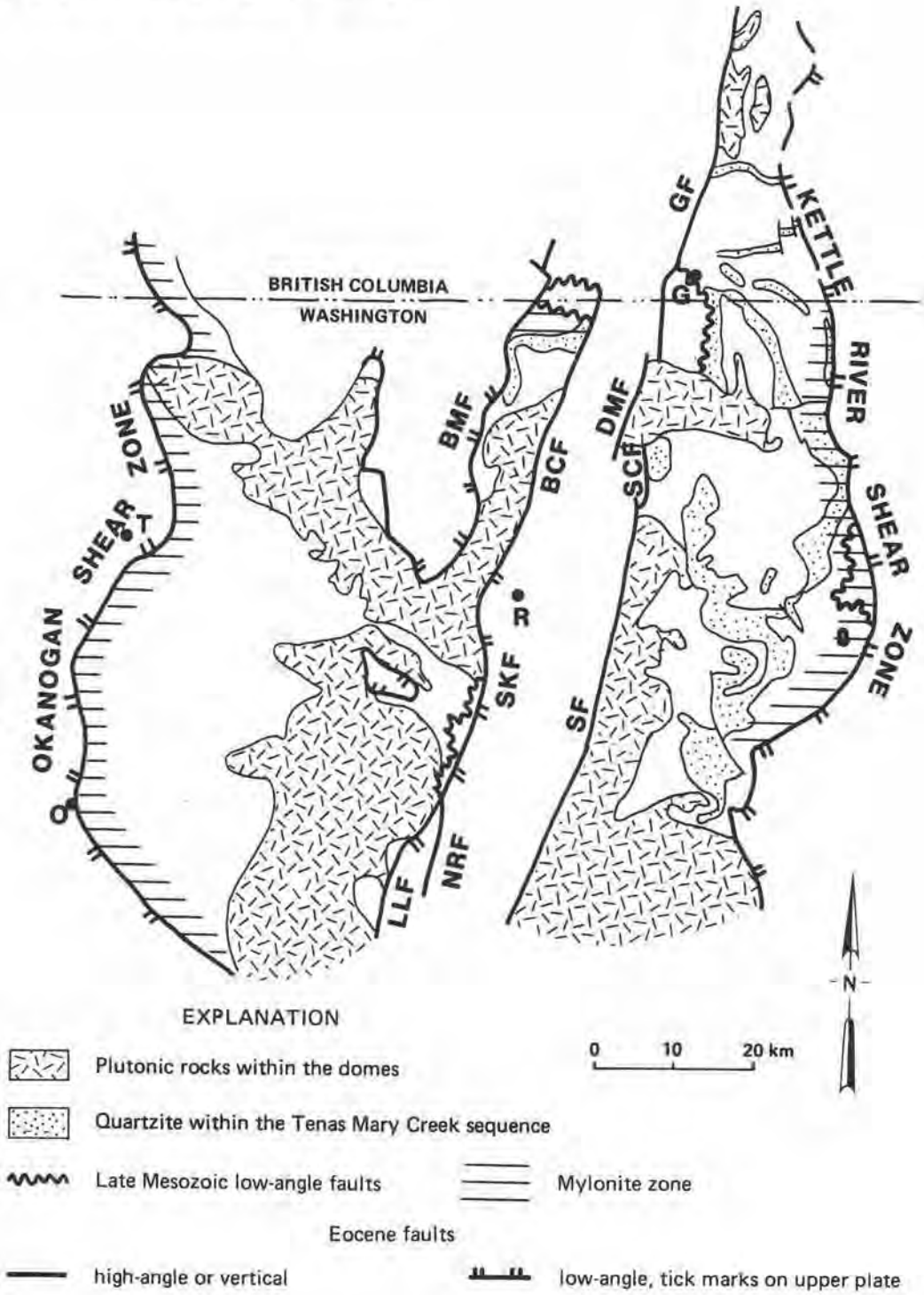


Figure 4.–Structural features within Okanogan and Kettle domes. BMF, Bodie Mountain fault (Pearson, 1967; Little, 1983; Orr, 1985); BCF, Bacon Creek fault (Parker and Calkins, 1964; Muessig, 1967; Staatz, 1964; Cheney and others, 1982b); DMF, Drummer Mountain fault (Parker and Calkins, 1964); GF, Granby fault (Preto, 1970; Pearson, 1977; Orr, 1985); LLF, Long Lake fault (Staatz, 1964); NRF, Nespelem River fault (Staatz, 1964); SCF, St. Peters Creek fault (Parker and Calkins, 1964); SKF, Scatter Creek fault (Muessig, 1967); SF, Sherman fault (Muessig, 1967; Cheney and others, 1982b); Okanogan shear zone (Snook, 1965; Atwater and others, 1984; Templeman-Kluit and Parkinson, 1986); Kettle River shear zone (Rhodes and Cheney, 1981; Wilson, 1981a; Atwater and others, 1984); T, Tonasket; O, Omak; R, Republic; G, Grand Forks.

Near the International Boundary and in Canada, the internal structure of Kettle dome is more complex. One of the orthogneisses (gppg) does not crop out, and the Cascade gneiss (g) is present only as isolated patches. The foliation in both the orthogneisses and paragneisses outlines several large overturned folds that have axes transverse to the main north-south trend and plunge of Kettle dome (Preto, 1970; Orr, 1982). Similarly, in the Tenas Mary Creek area, south of the International Boundary in the northeast part of Okanogan dome, and in the Swan Lake area, southwest of Republic along the east margin of Okanogan dome, the strikes of foliation in the metamorphic rocks are highly discordant to the margins of the domes. This discordance implies that metamorphism and deformation predate and are unrelated to the present structural relief of the Kettle and Okanogan domes.

Mylonitization

Distribution

In the Tenas Mary Creek area along the northeast arm of Okanogan dome, Orr (1985) described mylonitization associated with the tectonic suture between the continentally affiliated rocks of the Tenas Mary Creek sequence and the overlying allochthonous rocks of Quesnellia (Thompson assemblage). Mylonitization was pre- to synmetamorphic. The dynamics of the accretion of Quesnellia to North America and the attendant deformation and metamorphism have been well studied elsewhere in the Canadian Cordillera (Templeman-Kluit, 1979; Rees, 1981; Struik, 1981; Snook and others, 1981), and the suture is dated as mid-Jurassic.

Prominent 1- to 2-km-thick zones of mylonitized rock also are present along the east flank of Kettle dome (Donnelly, 1978; Rhodes, 1980; Rhodes and Cheney, 1981; Wilson, 1981b) and along the west flank of Okanogan dome (Snook, 1965; Goodge and Hansen, 1983). Mylonitic deformation in the Kettle River and Okanogan shear zones (Fig. 4) is texturally distinct from that in the Tenas Mary Creek area. In the Kettle and Okanogan River zones, mylonitic deformation is penetrative on an outcrop scale, whereas in the Tenas Mary Creek area, mylonitization is restricted to zones of varied width (1-50 cm) and spacing (1-10 m). Also, a well-developed stretching lineation is more prominent in the Kettle and Okanogan River zones.

Work by Rhodes and Hyndman (1984) in the Priest River complex also indicated the occurrence of two distinct mylonitic events. A 4-km-thick, top-to-the-east, Mesozoic mylonitic zone outlining the Spokane dome contrasts with the post-46-m.y. mylonitic zone marking the Eocene, scoop-shaped, extensional Newport fault (Harms and Price, 1983).

Kinematic indicators within the Okanogan shear zone are presently under study (Bardoux, 1985). Preliminary results from Kelowna, B.C., suggest a complex movement history with possible early top-to-the-east movement (Jurassic?) overprinted by normal top-to-the-west. In Washington, near Omak Lake, movement on mylonites in the Okanogan shear zone was top-to-the-west (V. L. Hansen, personal commun., 1985). In mylonites of the Kettle River zone, however, Donnelly (1978) reported top-to-the-east displacement. Thus the Okanogan and Kettle River shear zones appear to be independent, oppositely dipping and verging, low-angle normal faults.

Age

In contrast to many of the authors working in the metamorphic core complexes in the southwestern United States, Brown and Read (1983) emphasized that the low-angle mylonitic Monashee decollement in the Shuswap terrane is a Jurassic thrust unrelated to Tertiary crustal extension. Similarly, a 4-km-thick mylonite spatially unrelated to detachment faulting in the Priest River complex has been dated between 55 and 100 m.y. (Rhodes and Hyndman, 1984; Bickford and others, 1985; Armstrong and others, this volume). Thus, a consensus had been growing that, in the southern Omineca crystalline belt, ductile deformation (mylonitization) was Mesozoic and compressional and that Tertiary deformation was restricted to brittle deformation.

Radiometric dating by Parrish and Ryan (1983) in the Valhalla dome, however, showed that mylonitization also affected rocks as young as Paleocene and Eocene. Parkinson (Templeman-Kluit and Parkinson, 1986) produced an Eocene magmatic age based on U-Pb analysis of zircon from a mylonitized dike within the Okanogan shear zone. In addition, Rhodes and Hyndman (1984) described Tertiary mylonites that cross-cut the earlier Mesozoic mylonitic zone in the Priest River complex. Clearly then, ductile deformation has occurred in Tertiary time in the Okanogan-Kettle region.

Brittle Deformation

Detachment Faults

Faults, marked by brittle brecciation and chloritic alteration, generally separate the crystalline cores of Okanogan and Kettle domes from the lower grade rocks of Quesnellia and the Challis suite. These fault zones are typically less than 200 m thick and poorly exposed, but, where present, they resemble the zones of

chloritic breccia described along detachment faults in the southwestern United States. Localities where these faults zones are reported include: the Kettle River fault 1.5 km north of Barstow (Rhodes and Cheney, 1981); the Granby fault along the Granby River, north of Grand Forks, B.C. (Preto, 1970; Orr, 1982); the Bacon Creek fault in outcrop and drill holes north of Republic (Cheney and others, 1982b); and the Okanogan fault 3 km southwest of Tonasket and at Omak Lake (Snook, 1965; Goodge and Hansen, 1983).

In the Okanogan and Kettle River valleys this chloritic breccia is superimposed on the gently dipping mylonitic zones discussed above. Deep seismic reflection profiling shows the continuity of these major shear zones at depth. Potter and others (1986) noted the westerly dipping Okanogan shear zone as reflection F at a depth of 4 km about 10 km west of its trace on the west flank of Okanogan dome. Likewise, they inferred a nearly flat fault (their reflection H) at a depth of 5 km near Toroda Creek, which suggests that the west-dipping Bodie Mountain fault may extend to that depth (Fig. 5). Hurich and others (1985) identified the east-dipping mylonitized stratigraphy of the Tenas Mary Creek sequence at a depth of 3 km about 10 km east of Kettle dome. Elsewhere, the chloritic faults that bound the domes are not affiliated with mylonitic deformation and, with the exception of the southern part of the Bodie Mountain fault (Pearson, 1967), appear to be steeply dipping (Orr, 1985).

Reset Ages of Core Rocks

K-Ar age determinations in the metamorphic complexes of northeastern Washington yield discordant ages as young as 45 to 50 m.y. (Miller and Engels, 1975; Engels and others, 1976; Fox and others, 1977; Parkinson, 1985; Armstrong and others, this volume). Clearly, these are not primary ages.

Our preferred hypothesis for the extensive resetting of the K-Ar ages in the southern part of the Omineca belt is that they reflect quenching during rapid uplift. Fox and others (1977) support this hypothesis with the observation that, overall, the reset ages in the southern part of the Omineca crystalline belt are remarkably uniform and their extent is bounded by abrupt discontinuities coincident with the tectonic margins of the crystalline complex. In the northern part of the Okanogan complex, Mathews (1981) reported ages that are progressively younger away from the contact of the crystalline rocks with adjacent Tertiary strata. He speculated that this was due to the progressive uplift of structurally lower rocks through the K-Ar blocking temperature. Although this evidence collectively appears to support regional uplift as the cause of resetting, it does not preclude local thermal or hydrothermal effects.

STRUCTURAL SYNTHESIS

The complex deformation and regional metamorphism in Okanogan and Kettle domes attest to the occurrence of Mesozoic compressional orogenesis. Obduction and thrusting continued from Jurassic through earliest Cenozoic time and resulted in mylonitization. Normal movement on the Eocene ductile-brittle Okanogan and Kettle River shear zones and on other nonmylonitic Eocene faults indicates that extensional tectonism dominated in Eocene time.

Templeman-Kluit and Parkinson (1986) suggested that the Okanogan shear zone is a crustal-scale, west-dipping, extensional fault similar to the model proposed by Wernicke (1981). This model accounts for the superposition of brittle on ductile fabrics in a zone of progressive simple shear. In this model, lower plate rocks are initially deformed at depth in the ductile regime; as extension along the shear zone continues, these ductilely deformed rocks rise to structurally higher positions, eventually undergoing brittle deformation. Displacement on the fault must be large, at least enough to bring the brittle-ductile transition to the surface. If the Kettle River shear zone is an analogous east-dipping structure, Kettle and Okanogan domes are a composite, fault-bounded, crustal-scale lens. The multiple low-angle reflectors identified by Potter and others (personal commun., 1986) suggest the story may be more complex and that the entire southern part of the Omineca belt may be a set of many crustal lenses separated by deeply rooted, low-angle faults.

Faults other than the Kettle and Okanogan shear zones that bound the crystalline complexes do not show mylonitic deformation. In contrast to the Okanogan and Kettle River zones, they are generally high-angle normal faults (Orr, 1985). Displacements are difficult to constrain, but are probably on the order of 5 to 10 km. Latest movement on these faults was presumably synchronous with that of the ductile-brittle shear zones; all cut the Sanpoil or younger formations. Figure 5 shows synoptic cross sections illustrating the postulated movement histories of these faults. Estimates of total regional Eocene extension are dominated by estimates of displacement on the Okanogan and Kettle River shear zones. The high-angle faults shown on Figure 5 create roughly 7 to 10 km of horizontal extension; the Kettle River and Okanogan shear zones account for 50 km. Discounting any possible dilation by intrusions, this implies roughly 60 percent regional extension. Based on a vector analysis of

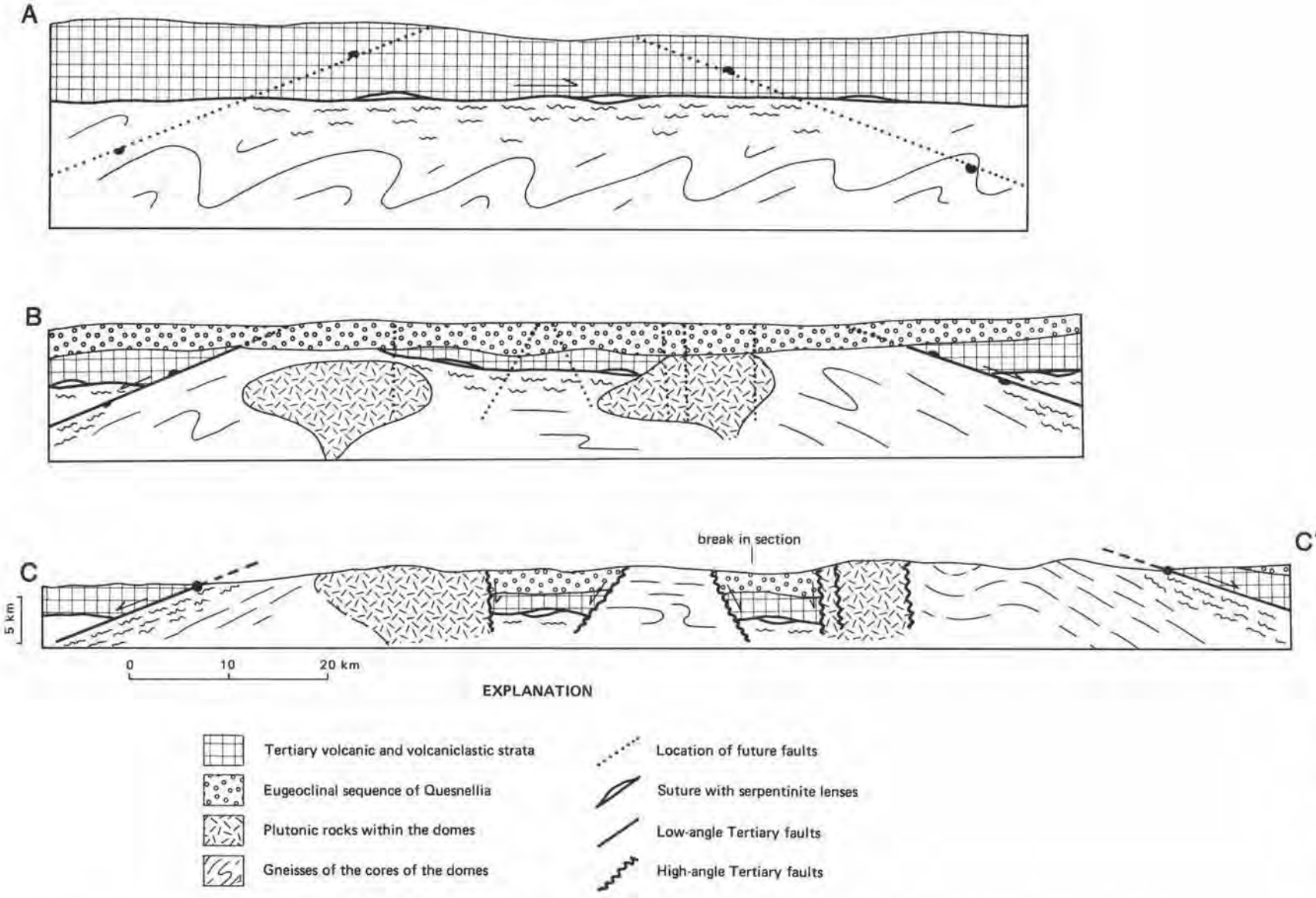


Figure 5.–Synoptic cross sections of the tectonic evolution of Okanogan and Kettle domes. Cross section C - C′ is to scale. Location is shown on Figure 3. A. Mid-Jurassic. Accretion of Quesnellia over North American Rocks. Regional metamorphism and deformation with concentrated mylonitic deformation near the suture. B. Mid-Eocene. Initial movement on Okanogan and Kettle River shear zones. Erosion and deposition of Eocene volcanic and volcaniclastic strata. C. Present. Final movement on Okanogan and Kettle River shear zones and on high-angle graben-forming faults. Regional erosion.

Eocene movement on major strike-slip faults in the southern part of the Omineca crystalline belt, Ewing (1980) also suggested about 50 percent extension should be expected in this area.

Fox and Beck (1985) showed that the paleomagnetic directions of the Challis rocks in the area of Kettle and Okanogan domes indicate that they have been rotated approximately 25° clockwise. Similarly the northwest strike of mylonitic lineations in Okanogan dome contrasted with the east-west strike in Kettle dome suggests that, if these mylonitic lineations formed in a single tectonic event, Okanogan dome has subsequently rotated clockwise with respect to Kettle dome. Fox and Beck (1985) attributed rotation to east-directed thrusting, but they noted that the latest Rocky Mountain thrusts (Paleocene) predate the rotated Eocene strata.

Price (1979) cited right-lateral strike-slip faulting induced by the overall northward migration of the Pacific plate with respect to North America as the cause for Eocene extension and rotation. He described the Okanogan area as a wrench basin between the dextral Straight Creek–Fraser fault system and the northern Rocky Mountain trench fault. Confirmed middle Eocene dextral movement on the Straight Creek fault (Vance, 1985) and the Mount Vernon fault in the Puget Lowland (Cheney, this volume) lend credence to this hypothesis.

Considerable regional uplift and denudation are required by the Eocene quenching ages in Kettle and Okanogan domes. Movement on the Okanogan and Kettle River shear zones may account for part of this uplift and denudation. Clearly, deep crustal gneisses and mylonites have been transported to shallow crustal levels along the Okanogan and Kettle River fault zones. The absence of domal reflectors in the data of Potter and others (1986) renders broad crustal arching or doming unlikely as a mechanism for uplift. One major problem arises, however. In the interior parts of the Okanogan-Kettle composite dome, away from the Okanogan and Kettle River shear zones, no evidence exists for low-angle Tertiary denudational faulting, yet the entire area gives anomalous Eocene ages.

Whether some of the denudation of the gneisses was erosional remains an open question. Erosional denudation would require unusually high erosion rates. Johnson (1984) proposed the Okanogan region as a potential source for part of the Chuckanut Formation of Eocene age in northwestern Washington. The deposition of the Chuckanut (>49 m.y.), however, at least partially predates resetting of the K-Ar ages, some of which occurred as recently as 45 m.y. The younger Roslyn, Chumstick, and Naches Formations and the Puget Group are part of a voluminous sequence within the Challis rocks that may, in part, be derived from the rising core complexes. They are appropriately arkosic, and their ages are ≤45 m.y. (Tabor and others, 1982a, 1982b; Frizzell and others, 1984; Johnson, 1984).

WESTERN EXTENT OF THE CORE COMPLEXES

The western extent of the metamorphic core complexes in northeastern Washington is still poorly known. The very well foliated Osoyoos orthogneiss on the northwest margin of the main body of the Okanogan dome appears to link the U.S. part of the dome with the high-grade gneisses in the Okanogan Valley of British Columbia (Tipper and others, 1981; Ross, 1981; Templeman-Kluit and Parkinson, 1986).

Relations on the southwest side of the Okanogan dome are less clear. Goodge and Hansen (1983) regarded the mylonitic, megacrystic gneiss at Omak Lake (about 25 km south of the southwest corner of Fig. 2) as the southwest margin of the Okanogan dome. These rocks are bounded by a zone of chloritic breccia. Similar megacrystic granitic rocks without a mylonitic foliation are present on the southwest side of Omak Lake, and 4 km southwest of Omak Lake these rocks are cut by a second zone of chloritic breccia. Multiple detachment faults have been described in southeastern California (John, 1982); this may be a similar occurrence. Because the southward extent of the granitic and high-grade metamorphic rocks is obscured by overlying Columbia River basalts (Fox and others, 1977; Atwater and others, 1984), the southern extent of the crystalline terrane is unknown.

The Chewak-Pasayten fault appears to be a major boundary between two geologic provinces. The province east of the fault is characterized by crystalline metamorphic complexes tectonically overlain by North American miogeoclinal rocks and the allochthonous rocks of Quesnellia. Southwest of the fault are the allochthonous Jurassic and Cretaceous, predominantly marine strata of the Methow-Tyaughton trough, the Permian to Jurassic Hozomeen Group, and the Mesozoic crystalline rocks of the Cascade Range. North of lat. 48° N., few, if any, Challis volcanic rocks are present west of the fault. The pre-Tertiary rocks west of the fault are cut by several major, high-angle, northwest-trending faults (Tipper and others, 1981; Monger and others, 1982), whereas no major northwest-trending faults occur east of the Chewak-Pasayten fault.

CONCLUSIONS

The age of the rocks within the cores of Okanogan and Kettle domes remains poorly known. Tentative lithologic correlations and the general continental affiliation of the metasedimentary Tenas Mary Creek sequence suggest that it is Proterozoic. Metamorphism throughout the Omineca crystalline belt was Jurassic. In the southern part of Kettle dome the metamorphic stratigraphy and foliation roughly define a dome. In the north, however, the structure is considerably more complex, and the main structural trend is transverse to the north-south axis of Kettle dome, implying that, genetically, Kettle dome is not a dome. Similar discordant features occur in the eastern part of Okanogan dome.

Two stages of mylonitization exist. The first was Mesozoic and compressional, the second Eocene and extensional. In the Tenas Mary Creek area, the first mylonitization resulted from the obduction of the allochthonous terrane Quesnellia. In the Canadian Cordillera this event is dated as mid-Jurassic; compressional mylonitization in the Spokane dome is younger, between 55 and 100 m.y. (Bickford and others, 1985; Armstrong and others, this volume).

By 50 m.y. ago the tectonic regime was extensional. Considerable regional uplift and denudation set K-Ar ages in the Okanogan and Kettle core complexes at 45 to 50 m.y. The Okanogan and Kettle River shear zones bound a composite lens-shaped Okanogan-Kettle dome. Within this lens, moderate to high-angle faults with relatively minor displacement (<10 km) separate the basement gneisses from structural lows in which Tertiary volcanic and volcaniclastic strata are preserved.

The Tertiary rocks of the Okanogan highlands still are poorly known. The possibilities of regional correlation need to be evaluated, as does the evidence for local source areas for both the sedimentary and volcanic rocks. Identification of source areas and of structures in the Tertiary rocks could provide considerable insight into the tectonic evolution of the domes.

The cause of Tertiary crustal extension is speculative. Coney and Harms (1984) note that the age of Tertiary detachment faulting in the Cordillera varies with the age of Tertiary volcanism. They suggested that pre-Tertiary obduction and thrusting generated early mylonites and over-thickened the crust. Tertiary plutonism and volcanism heated this crust, rendering it susceptible to attenuation. In northeastern Washington this attenuation may have been transtensional, related to the northward movement of the Pacific plate with respect to North America. The presumed cause of dextral faulting on the western margin of the continent was northward transport of the Pacific plate with respect to the North American plate. This might have attenuated a crust that had been rendered ductile by Challis volcanism and plutonism.

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UPPER MESOZOIC METHOW-PASAYTEN SEQUENCE, NORTHEASTERN CASCADE RANGE, WASHINGTON AND BRITISH COLUMBIA [1]

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ABSTRACT

A Lower Jurassic through Upper Cretaceous sequence of sandstone, shale, and conglomerate in the northeastern Cascade Range reflects tectonic chronologies in flanking crystalline terranes of the Columbian orogen to the east and the Pacific orogen to the west. Progressive unroofing of an eastern Triassic-Jurassic magmatic and regional metamorphic complex in the Okanogan terrane is manifested in the sedimentary record by Jurassic-Lower Cretaceous volcaniclastic rocks (Sinemurian-Neocomian) and plagioclase arkose and shale (Aptian-Albian) deposited on submarine fans in a westward-deepening basin. In early Late Cretaceous time (Cenomanian-Turonian?), the inception of deformation, magmatism, and metamorphism in the north Cascades orogenic core to the west caused emergence and erosion of a Permian-Jurassic chert-greenstone supracrustal complex, reflected in the Methow-Pasayten sequence by deposition of chert-pebble conglomerate interbedded with volcaniclastic rocks free of granitic detritus. Deformation of the sedimentary sequence ensued immediately and is associated in time with Shuksan thrusting and probably Skagit metamorphism of middle Late Cretaceous age in the North Cascades core.

It is our interpretation that the Methow-Pasayten sequence occupied a fore-arc setting in Jurassic-Early Cretaceous time, analogous to the Great Valley sequence. As magmatism, tectonism, and metamorphism shifted westward from the Okanogan region to the North Cascades Coast Plutonic Complex in early Late Cretaceous time, possibly in association with the arrival of Wrangellia, the basin was filled and deformed in a rear-arc position.

INTRODUCTION

Moderately deformed sedimentary rocks of the Methow-Pasayten area in north-central Washington and southern British Columbia provide useful constraints on the Mesozoic evolution of the Cordillera in the Pacific Northwest. This region has long resisted tectonic interpretation because of very complex geologic relations, which have been studied most intensively in crystalline terranes of the North Cascades (Misch, 1966; McTaggart, 1970) and the Okanogan region (Hawkins, 1968; Menzer, 1970; Hibbard, 1971; Fox and others, 1977), which lie respectively to the east and west of the Methow area (Fig. 1). Mapped in part by Barksdale (1975), Coates (1970, 1974), Staatz and others (1971), Jeletzky (1972), Tennyson (1974), and Kleinspehn (1982), this sliver of dominantly marine sedimentary rocks was downfaulted in latest Mesozoic or Tertiary time along two major regional faults and thus escaped the erosion that evidently destroyed the rest of the apparently more extensive original sequence (Cole, 1973). It is often referred to informally as the "Methow graben," although it is not a graben in the topographic sense, nor are the bounding faults necessarily normal faults. The graben is the southernmost extension of the Tyaughton trough (Jeletzky and Tipper, 1968; Eisbacher, 1974), and it lies along the boundary between the Columbian and Pacific orogens established in Canadian literature (Wheeler and others, 1972), of which the Okanogan complex and North Cascades are respective elements. It contains one of the more complete and instructive Jurassic-Cretaceous sedimentary sequences in the Pacific Northwest. This paper summarizes the geology of the graben and suggests its implications for regional tectonic history.

[1] Reprinted with modifications from: Howell, D. G.; McDougall, K. A., editors, 1978, Mesozoic paleogeography of the western United States: Society of Economic Paleontologists and Mineralogists Pacific Section Symposium 1, p. 499-508.

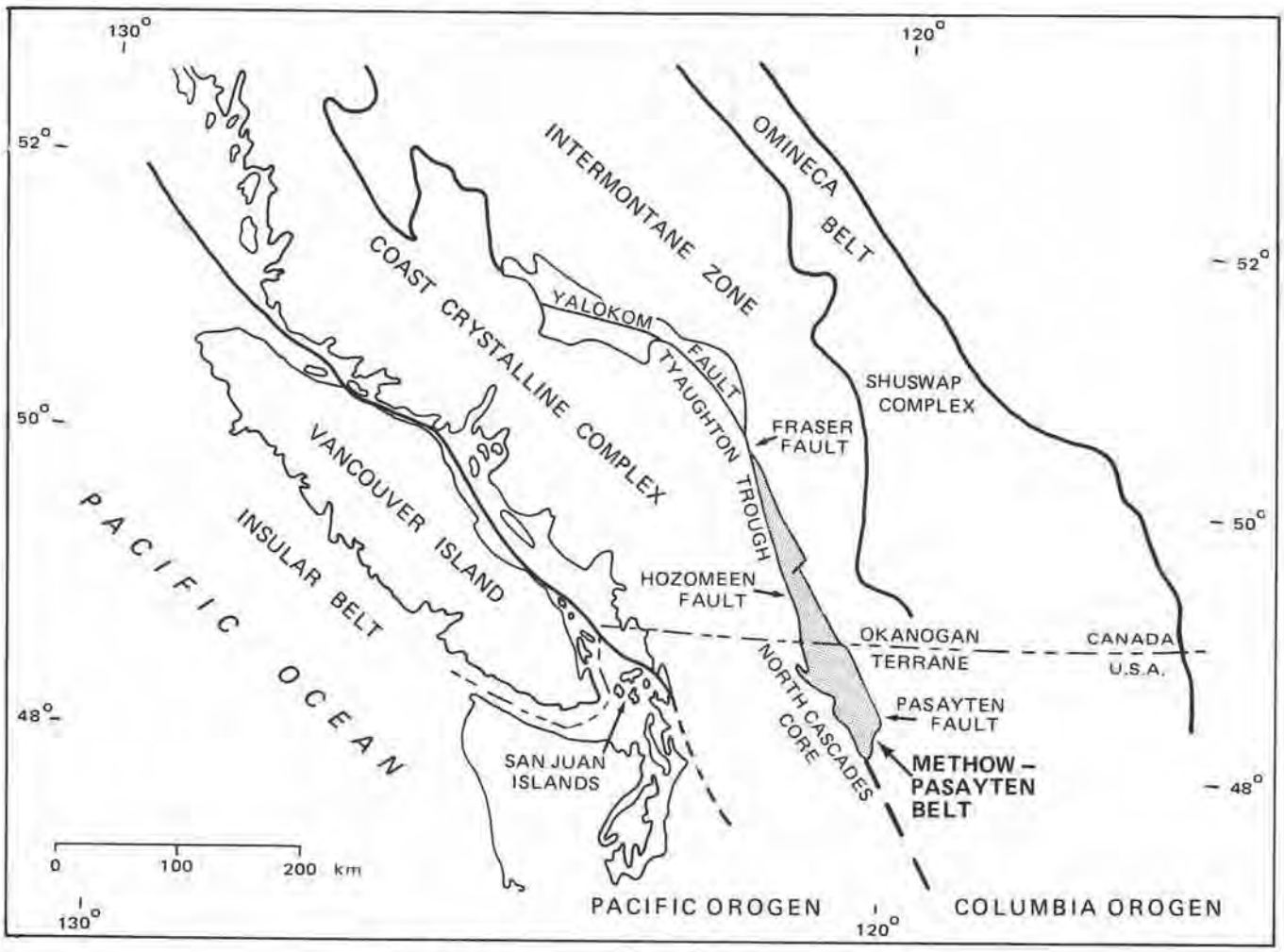

Figure 1.–Tectonic setting of the Methow-Pasayten belt.

STRATIGRAPHY AND STRUCTURE OF THE METHOW-PASAYTEN SEQUENCE

Jurassic and Cretaceous rocks of the Methow-Pasayten sequence occupy a belt 275 km long and up to 40 km wide, bounded by two steeply dipping strands of the Fraser-Yalokom fault system. These faults separate the sedimentary rocks from the Okanogan-Shuswap terrane (Columbian orogen) on the east, and from the North Cascades crystalline core (Pacific orogen) on the west. The western boundary fault is actually a complex fault system (Ross Lake-Jack Mountain fault system of Misch, 1966; Hozameen fault of Monger, 1970; Twisp fault of Barksdale, 1975) along which a block of submarine basalt, bedded chert and shale, the Hozameen Goup of Permian to Middle Jurassic age (Tennyson and others, 1982; Haugerud, 1984), has been emplaced between the Mesozoic sedimentary rocks and the infrastructural rocks of the Cascade core. Serpentinite locally occupies the fault zone (Monger, 1970; Staatz and others, 1971). The sedimentary column within the graben (Fig. 2) reaches a thickness of 15 to 20 km, representing Early Jurassic to early Late Cretaceous time. Marked thickness variations and facies changes, along with a variety of sedimentary structures, allow inferences about the paleogeographic setting of the rocks preserved in the graben. These features provided the data for a basin analysis (Cole, 1973), the results of which are referred to below. The Jurassic and Lower Cretaceous part of the section is marine, whereas Upper Cretaceous rocks include both marine and nonmarine units. It is in some cases difficult to correlate marine and nonmarine units with certainty; paleontologic ages,referred to below are better established for marine units.

The oldest sedimentary rocks within the graben are marine Jurassic and Lower Cretaceous mudstones, volcanic sandstones, and volcanic breccias of the Newby, Ladner, and Dewdney Creek Groups (Barksdale, 1975; Coates, 1974). These rocks were derived from a volcanic terrane to the east and deposited in a submarine-fan environment in a basin that deepened to the west.

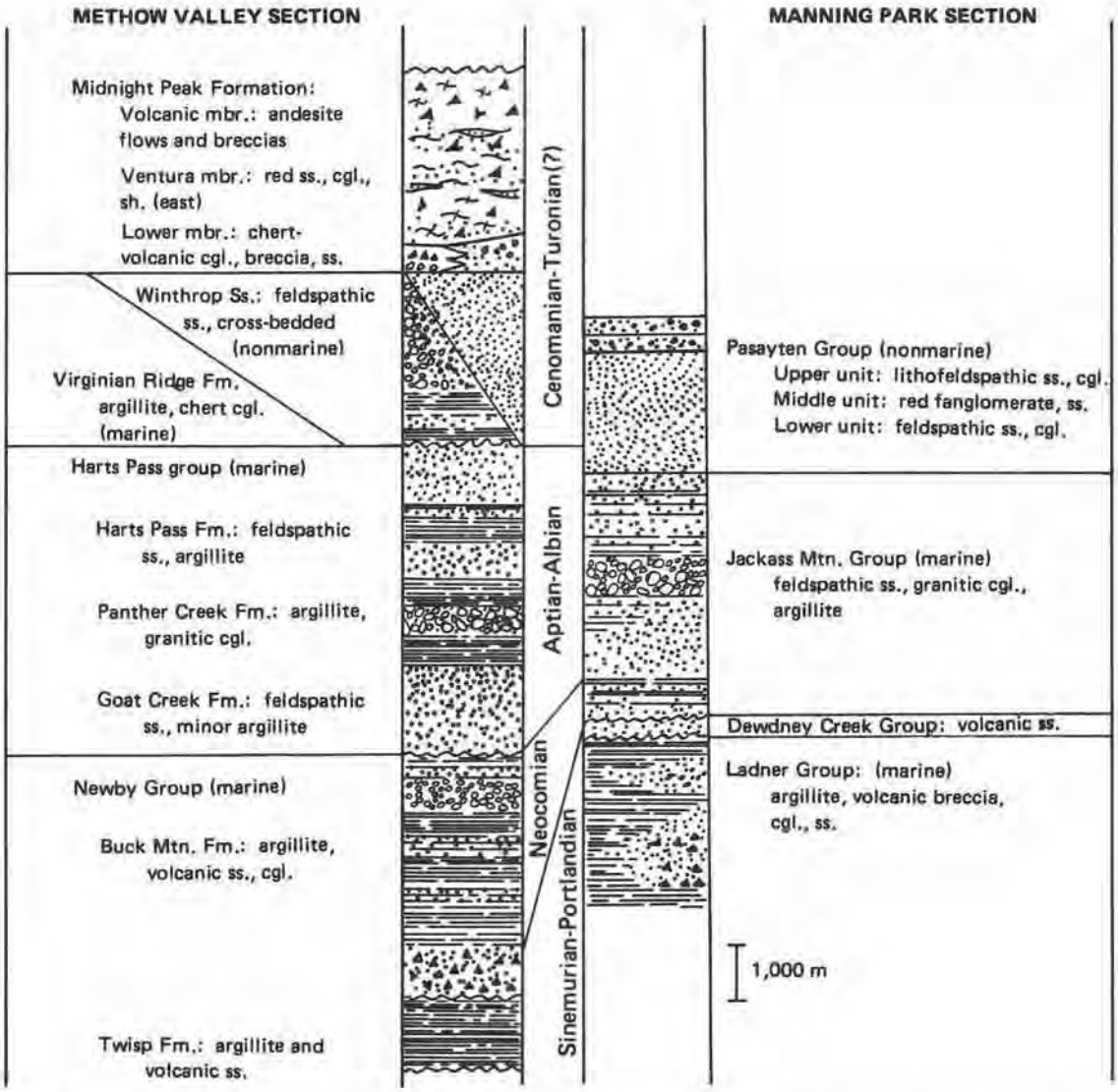

Figure 2.–Stratigraphic columns for Jurassic-Cretaceous rocks in the Methow Valley, Washington (after Barksdale, 1975), and Manning Park, British Columbia (after Coates, 1974). Thicknesses and lithologies generalized.

Sedimentary structures which allow reconstruction of transport directions (Table 1) include small ripple-drift cross-laminations and elongate clasts. Thin-bedded, graded siltstones and fine-grained, parallel-laminated sandstones of general "flysch" aspect are useful in inferring a depositional environment; a submarine-fan complex seems consistent with all of these features.

Figure 3 is a diagrammatic representation of Jurassic and Early Cretaceous (Neocomian) paleogeography which incorporates these conclusions. Fans shown in the diagram were not all active at the same time; the locus of deposition, as deduced from exposed rocks, apparently shifted back and forth, in a manner typical of modern fans. The lack of shoreline or shallow marine deposits along the east side of the graben indicates that the Jurassic-Early Cretaceous shoreline lay an unknown distance to the east. We cannot say how far west this depositional regime extended, except to point out that volcanic sandstones of the Nooksack Group on the west side of the Cascades are correlative with and petrologically similar to those of the Newby, Ladner, and Dewdney Creek Groups (Tennyson, 1972). Misch (1966) described the partly correlative Cultus and Bald Mountain Formations on the west side of the Cascades, comprising a variety of marine clastic rocks. To our knowledge, no paleocurrent work has been done on these rocks which might shed light on their relationship to the Methow basin.

Petrology of Jurassic and Lower Cretaceous (Neocomian) Methow sandstones (Table 1) suggests an

Table 1.–Stratigraphy and sedimentology of the Methow-Pasayten sequence in Washington and British Columbia

Washington

Age	Unit	Thickness (ft, m)	Sedimentation rate (ft, m/m.y.)	Petrology*	Paleocurrent azimuth**
Cenomanian-Turonian(?)	Midnight Peak Fm.	10,400 3,170			
	Ventura Mbr.	2,040 622		$Q_{24}F_{38}L_{38}$ V/L = .59	
	Winthrop Sandstone	13,500 4,115	2,887 880	$Q_{26}F_{54}L_{20}$ V/L = .56	225° (*n* = 201)
	Virginian Ridge Fm.	11,600 3,536		$Q_{29}F_{25}L_{46}$ V/L = .27 C/L = .57	55° (*n* = 64)
Albian	Harts Pass Fm.	7,900 2,408			
	Panther Creek Fm.	5,200 1,585	1,882 555	$Q_{32}F_{62}L_{6}$ V/L = .32	296° (*n* = 122)
Aptian	Goat Creek Fm.	5,120 1,561			
Barremian	Buck Mtn. Fm. (Newby Group)	14,500 4,420	2,416 736		
Portlandian	Twisp Fm. (Newby Group)	4,000 1,220	166 51	$Q_{8}F_{30}L_{62}$ V/L = .98	232° (*n* = 73)

British Columbia

Age	Unit	Thickness (ft, m)	Sedimentation rate (ft, m/m.y.)	Petrology*	Paleocurrent azimuth**
Albian-Turonian(?)	Upper unit	1,000 305	769 234	$Q_{24}F_{25}L_{51}$ V/L = .94	
	Middle unit	1,000 305		$Q_{31}F_{52}L_{17}$ V/L = .71	303° (*n* = 6)
	Lower unit	8,000 2,438		$Q_{29}F_{68}L_{3}$ V/L = .05	254° (*n* = 12)
Hauterivian-Albian	Jackass Mtn. Group	14,000 4,267	848 258	$Q_{15}F_{52}L_{33}$ V/L = .91	
Sinemurian-Portlandian	Dewdney Creek Group	1,000 305	80 24	$Q_{3}F_{16}L_{81}$ V/L = .97	
	Ladner Group	6,000 1,829	250 76	$Q_{3}F_{35}L_{62}$ V/L = .98	

* Q = quartz grains; F = feldspar grains; L = lithic grains, including chert; V/L = volcanic/lithic ratio; C/L = chert/lithic ratio.
** *n* = number of readings.

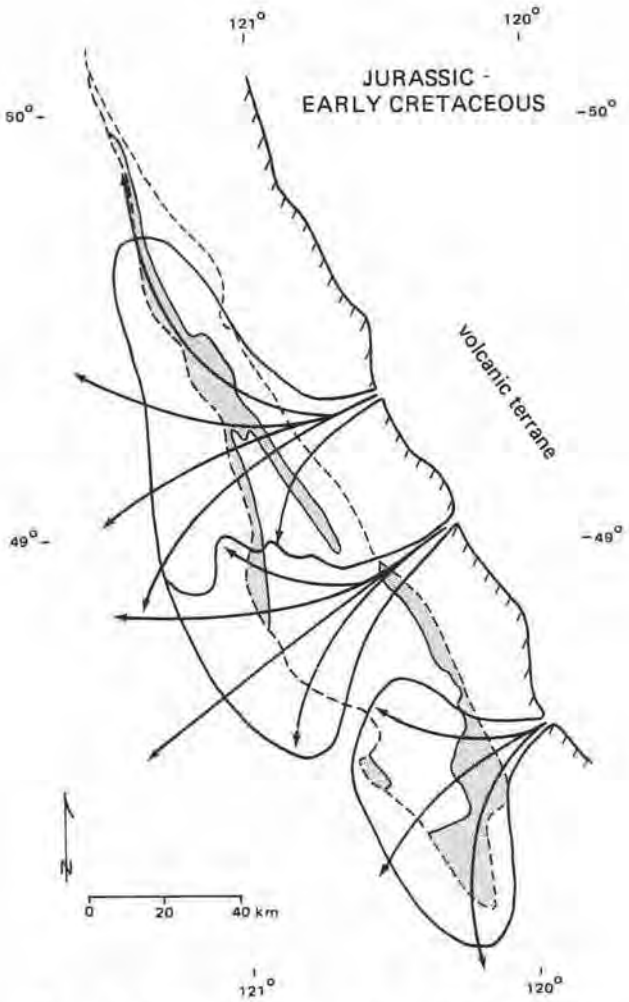

Figure 3.–Schematic representation of Jurassic-Early Cretaceous (Neocomian) paleogeography in the Methow-Pasayten belt (dashed line), showing submarine fans and inferred volcanic upland to the east. Outcrop areas are stippled.

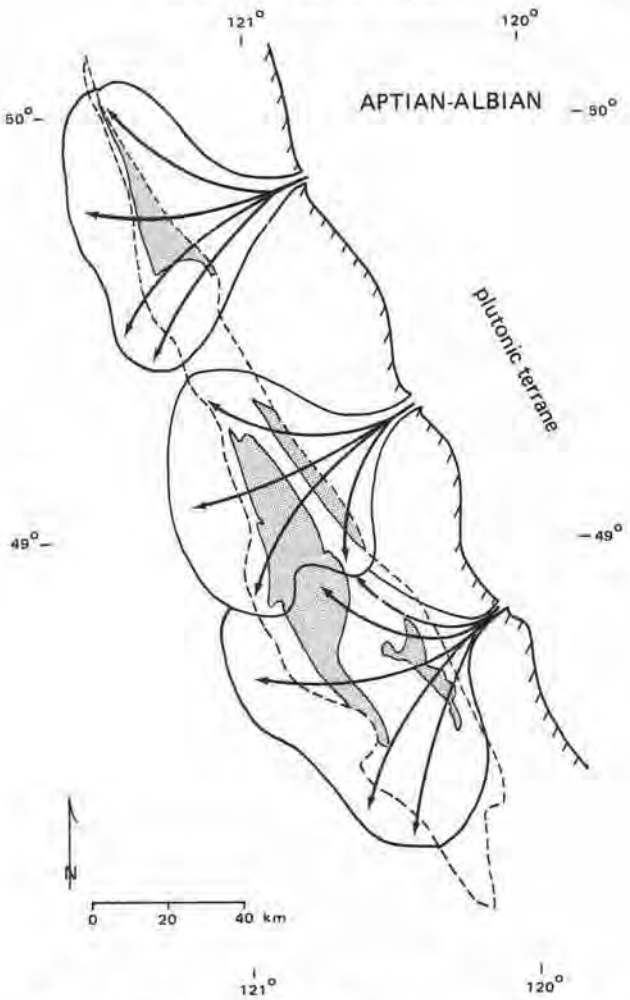

Figure 4.–Schematic representation of later Early Cretaceous (Aptian-Albian) paleogeography in the Methow-Pasayten belt (dashed line). Submarine fans received arkosic detritus from an unroofed plutonic source to the east. Outcrop areas are stippled.

exclusively volcanic source. They are free of granitic detritus except for lowermost Cretaceous rocks. Small amounts of granitic material in Neocomian units announce a shift to the arkosic composition typical of Aptian and younger east-derived rocks. Sedimentation rates, although impossible to calculate with precision, seem to show a marked increase at about the same time (Table 1).

Succeeding the volcanic turbidites with at least local unconformity (Barksdale, 1975) is a sequence of Aptian-Albian plagioclase arkose, argillite, and granite-bearing conglomerate of the Goat Creek, Panther Creek, and Harts Pass Formations (here referred to informally as the Harts Pass group) of Barksdale (1975) and the Jackass Mountain Group of Coates (1974). Paleocurrent analyses of these rocks, based on groove casts, elongate clast orientations, ripple-drift cross-laminations, and flute marks, indicate westward transport, a conclusion supported by westward decrease in average sandstone bed thickness, sandstone/shale ratios, and maximum clast size across the graben. Graded sandstone and siltstone beds, parallel laminations, contorted laminations, scoured bases, and flute and groove marks suggest deposition by turbidity currents on west-sloping submarine fans; proximal environments are suggested by fluxoturbidites on the east, and somewhat more distal environments are suggested on the west side of the graben (Fig. 4). Conglomerates are most easily interpreted as channel fillings. Sandstones of these units show marked consistency in their composition; typical modes are approximately Q_{30}:F_{60}:L_{10}. K-feldspar is very rare. The proportion of volcanic lithic fragments and chert is minor.

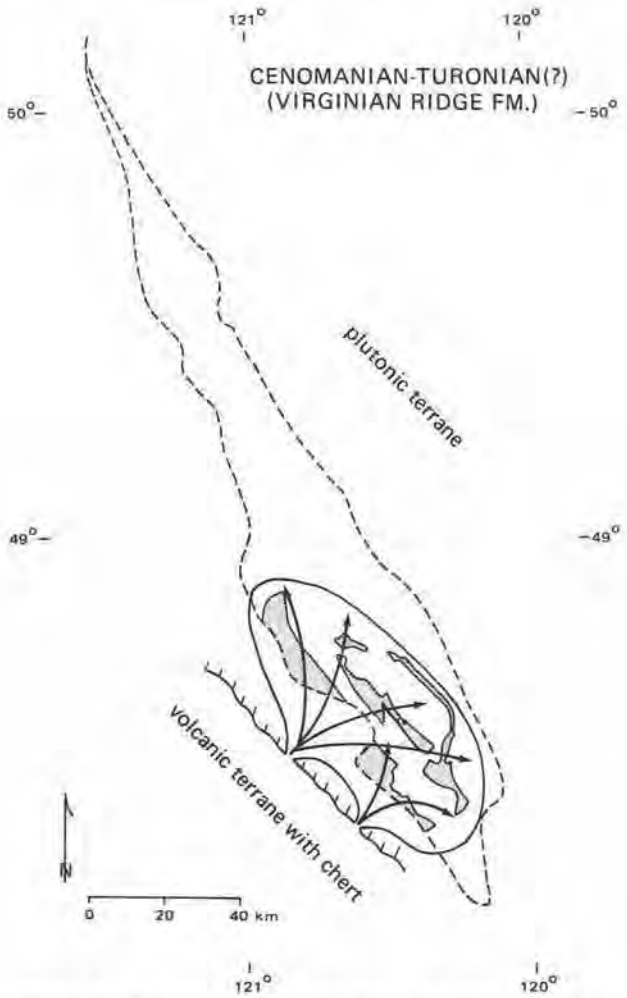

Figure 5.–Schematic representation of paleogeography during a part of Cenomanian-Turonian(?) time in the Methow-Pasayten belt (dashed line), showing deposition of marine chert-pebble conglomerate and mudstone (Virginian Ridge Formation) derived from the west. Outcrop areas are stippled. See also Figures 6 and 7.

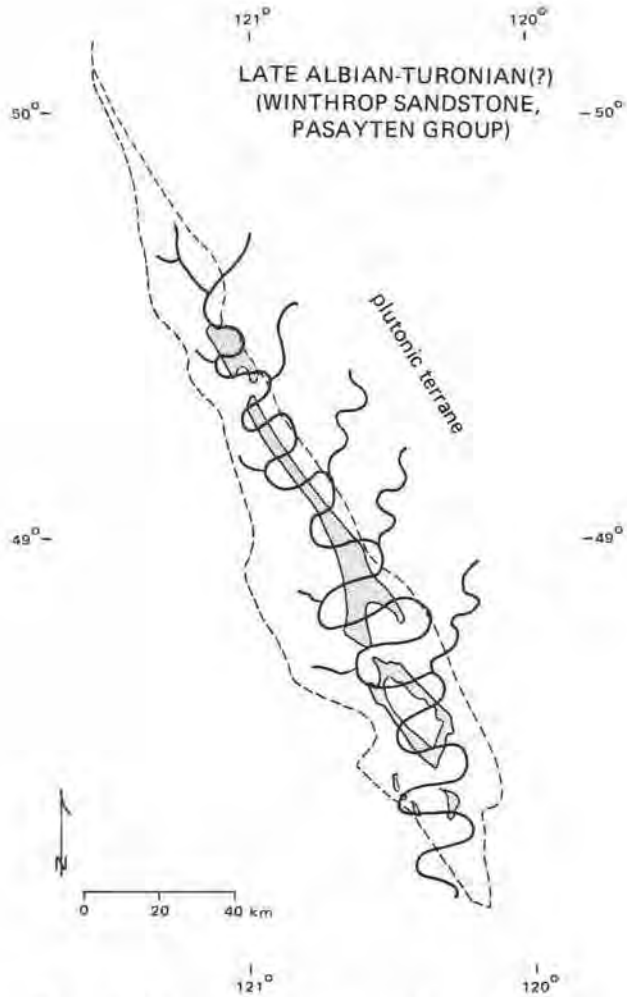

Figure 6.–Schematic representation of paleogeography during a part of late Albian to Turonian(?) time in the Methow-Pasayten belt (dashed line), showing deposition of plant-bearing arkosic sandstone (Winthrop Sandstone) on the south, and related nonmarine sandstones and conglomerates (Pasayten Group) on the north. Inferred drainage to the south and southeast. Outcrop areas are stippled. See also Figures 5 and 7.

Lower Upper Cretaceous strata show complex local variations and facies relationships (Fig. 5). They are at least locally unconformable on older rocks (Barksdale, 1975), and they show important departures from the earlier depositional framework. Most notable is a west-derived, mostly marine chert-pebble conglomerate-and-mudstone unit, the Virginian Ridge Formation. It coarsens and thickens markedly to the west, ranging from a thickness of about 300 m on the east side of the graben with a maximum grain size in the gravel range, to more than 3,500 m on the west side, where clasts reach 6 cm in diameter. Sandstone/shale ratios decrease toward the east, and paleocurrent indicators show eastward transport. Typical sedimentary structures include graded bedding, convolute laminations, and ripple-drift cross-laminations in the siltstones; sandstones show grading and rare cross-bedding. Conglomerates are lenticular and graded in the central and east parts of the graben but continuous and massive along the west side. A likely depositional environment is a fan-delta or delta.

Composition of Virginian Ridge sandstones and conglomerates similarly indicates a change in provenance. The typical sandstone mode of Q_{30}:F_{25}:L_{45} includes about 40 percent chert grains, counted either as lithic or quartzose grains. Volcanic lithic grains are generally altered basaltic fragments. Quartz grains tend to be unstrained, suggesting a volcanic origin. Some plagioclase is present. Conglomerates are completely free of granitic cobbles, instead comprising

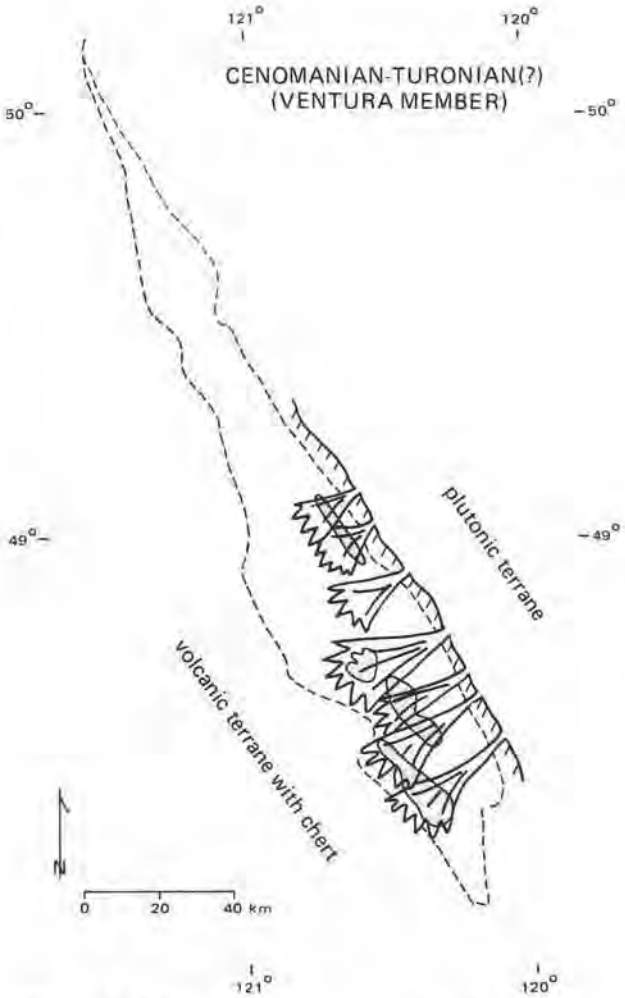

Figure 7.–Schematic representation of paleogeography during a part of Cenomanian-Turonian(?) time in the Methow-Pasayten belt (dashed line), showing deposition of redbeds of highly variable composition in an alluvial-fan complex (Ventura Member). Volcanic and cherty sediment (Midnight Peak Formation) are mixed on the west side of the belt. Outcrop areas are stippled. See also Figures 5 and 6.

chert almost exclusively, with a subordinate component of altered mafic volcanic clasts. Sandstones and conglomerates are locally tuffaceous (Tennyson and Whetten, 1974).

The Virginian Ridge Formation intertongues on the east with a plant-bearing arkosic unit, the Winthrop Sandstone (Barksdale, 1975) or the Pasayten Group (Coates, 1974). This unit thickens markedly to the east (maximum thickness more than 4,000 m) and is absent to the west. It is characterized by fining-upward sequences, trough cross-bedding, small cross-laminations, and climbing ripples, reflecting a fluvial or deltaic environment of deposition. Transport was to the south-southeast (Fig. 6). Composition of this sandstone is virtually identical to that of Aptian-Albian arkoses.

Locally, in the southeast part of the graben, the Winthrop Sandstone and the Virginian Ridge Formation grade into redbeds of highly variable composition (Ventura Member; Fig. 7) and andesitic flows and pyroclastic rocks of the Midnight Peak Formation (Barksdale, 1975), with which the redbeds are usually associated. Volcaniclastic rocks include lahars and water-laid graded tuffs which demonstrate mixing of volcanic and cherty sediment on the west side of the belt (Tennyson and Whetten, 1974). Facies variations between chert-rich sediment, plagioclase arkose, redbeds, and volcanic rocks are complex. Ages are uncertain because of difficulty in correlating plants and marine invertebrates and because of poor preservation, but the rocks appear to represent an interval between late Albian and Coniacian, with an emerging consensus on a Cenomanian age (Barksdale, 1975).

The sedimentary rocks are deformed into large, evidently parallel folds and cut by steeply dipping longitudinal faults and cross faults. Fold wavelengths range up to 10 km. Penetrative deformation is present only locally, along faults or in fold cores. Fold axes are roughly parallel to boundary faults. Longitudinal faults tend to substitute for anticlinal axes.

Barksdale's mapping (1975) has demonstrated angular unconformities within the Jurassic-Cretaceous section, which testify to pre-Late Jurassic(?), late Neocomian, and Albian deformation, but major deformation accounting for the present geometry is well bracketed between late stages of deposition in Cenomanian-Turonian(?) time (97-88 m.y. on the time scale of Palmer, 1983) and cross-cutting plutons about 85 m.y. old (Santonian). It is doubtful that the depth of burial during deformation exceeded the maximum thickness of the sedimentary column, about 15 km, since continental sediments only slightly older than the deformation are widely preserved, and since there is little evidence of plastic behavior or penetrative deformation. Furthermore, incipient metamorphic minerals, present only locally, are limited to zeolites, prehnite, and pumpellyite.

Sense of movement on boundary faults is uncertain. Stratigraphic separation amounts to almost the thickness of the sedimentary column along the Hozameen fault, and perhaps somewhat less on the Pasayten fault (Fig. 1). Some strike-slip movement is likely, but unproven. The Hozameen-Jack Mountain fault system was interpreted by Misch (1966) as a combined eastward overthrust and dextral strike-slip fault, but field relations suggest that the near-vertical Hozameen fault in fact truncates and thus postdates the subhorizontal Jack Mountain thrust (Staatz and

others, 1971; Tennyson, 1974). Age of thrusting is not known, since ages of the rocks involved are speculative. Age of activity on the Hozameen fault itself, however, is bracketed between Cenomanian, the age of the youngest rocks involved, and 88 to 84 m.y. (Coniacian-Santonian on the time scale of Palmer, 1983), the K-Ar ages of two plutons that cut the fault (Coates, 1974; Peter Misch, personal commun., 1972). The Pasayten fault on the east edge of the belt cuts Aptian-Albian strata and Paleocene rocks; it is overlain by the Tertiary Island Mountain volcanics (Lawrence, 1978). Late Cretaceous activity cannot be ruled out, and it is perhaps suggested by the fanglomerate character of Upper Cretaceous redbeds along the east side of the belt (Cole, 1973). Lawrence (1971, 1978) has studied petrofabric evidence for various possible offset histories.

A summary of the late Mesozoic depositional and deformational history of the area, then, includes three main phases: (1) Sinemurian-Neocomian deposition of east-derived turbidites rich in volcanic detritus in a marine basin deepening westward; (2) Aptian-Albian deposition of east-derived plagioclase arkose, granitic conglomerate, and mudstone in a similar marine environment; and (3) Cenomanian shoaling and partial emergence of the basin, deposition of east-derived fluvial-deltaic plagioclase arkose and west-derived marine chert-pebble conglomerate and argillite, local volcanism, folding, faulting, and intrusion.

TECTONIC SIGNIFICANCE

The Methow-Pasayten belt lies along the boundary between the Columbian orogen and the Pacific orogen as defined by Canadian workers (Wheeler and others, 1972), in this case the Okanogan terrane and the North Cascades core, respectively. An important question is the relationship between these two infrastructural terranes. They are composed of similar rock types and yield radiometric ages that show substantial synchroneity of latest Mesozoic and Tertiary plutonism. The evolution of each infrastructure will be examined here briefly and correlated with the supracrustal record preserved by sedimentary rocks in the Methow graben.

The Okanogan terrane has yielded radiometric and geologic evidence for regional metamorphism of a late Paleozoic-earliest Mesozoic "eugeosynclinal" complex and intrusion of ultramafic to granitic plutons between Late Triassic and Middle Jurassic time (Hibbard, 1971; Rinehart and Fox, 1972; Miller and Engels, 1975). During most of Jurassic time a volcanic arc evidently mantled the crystalline rocks so produced, since quartzofeldspathic detritus does not appear in east-derived sedimentary rocks of the Methow-Pasayten sequence older than Lower Cretaceous. Uplift and cessation of volcanism allowed erosion to unearth the Jurassic-Neocomian infrastructure during middle Cretaceous time; Hibbard (1971) argued, on the basis of metamorphic mineral assemblages, that about 16 km of unloading must have occurred between Early Jurassic and late Early Cretaceous time in the Okanogan area. It is not clear from available radiometric-age data, summarized in Figure 8, whether the Okanogan terrane was host to two plutonic pulses—one in Late Triassic to Middle Jurassic time (~200-150 m.y. ago), and the other in middle Cretaceous time (~110-85 m.y. ago)—or whether there was actually a continuum, and plutons intruded during the apparent gap simply were not uplifted sufficiently for cooling until early Late Cretaceous time. Plutonism logically would have accompanied the Middle Jurassic-Early Cretaceous volcanism evident in the supracrustal record, which argues for the latter speculation and confines any interlude to late Neocomian-Aptian time (130-110 m.y. ago), a span for which no volcanic and few plutonic rocks have been reported (Engels and others, 1976; Fox and others, 1977). It was during this interval that unroofing of the plutonic rocks occurred and arkosic sedimentation in the Methow-Pasayten belt began. Resumption of volcanism in late Albian or Cenomanian time and the rash of plutonic-rock dates between 110 and 85 m.y. are almost certainly related. Hibbard (1971, p. 3033) reported nonpenetrative deformation in association with relatively shallow intrusive activity in the approximate interval 110 to 90 m.y., or Albian-Cenomanian time. Rinehart and Fox (1972) described strike-slip faulting and thrusting which post-date the Jurassic (Fox and others, 1977) Similkameen batholith. This episode of deformation and shallow plutonism, then, is evidently the same event which affected the Methow-Pasayten belt.

The evolution of the North Cascades core (Misch, 1966) differs significantly. Lower Mesozoic rocks, where present, are marine sedimentary and volcanic rocks, and there is little evidence for a major regional metamorphic-plutonic event of this age, the only exceptions being blueschist-facies metamorphism with one K-Ar date of 219 m.y., and one quartz-diorite pluton (Marblemount) with an intrusive age of 220 m.y. (Misch, 1966). The Methow-Pasayten sequence demonstrates that there was no significant emergent Cascade landmass in pre-Late Cretaceous time. In early Late Cretaceous time, however, supracrustal rocks must have emerged above base level in the Cascades core—the late Paleozoic to early Mesozoic marine sedimentary and volcanic rocks still preserved on the flanks of the Cascade orogen. Destruction of these rocks produced

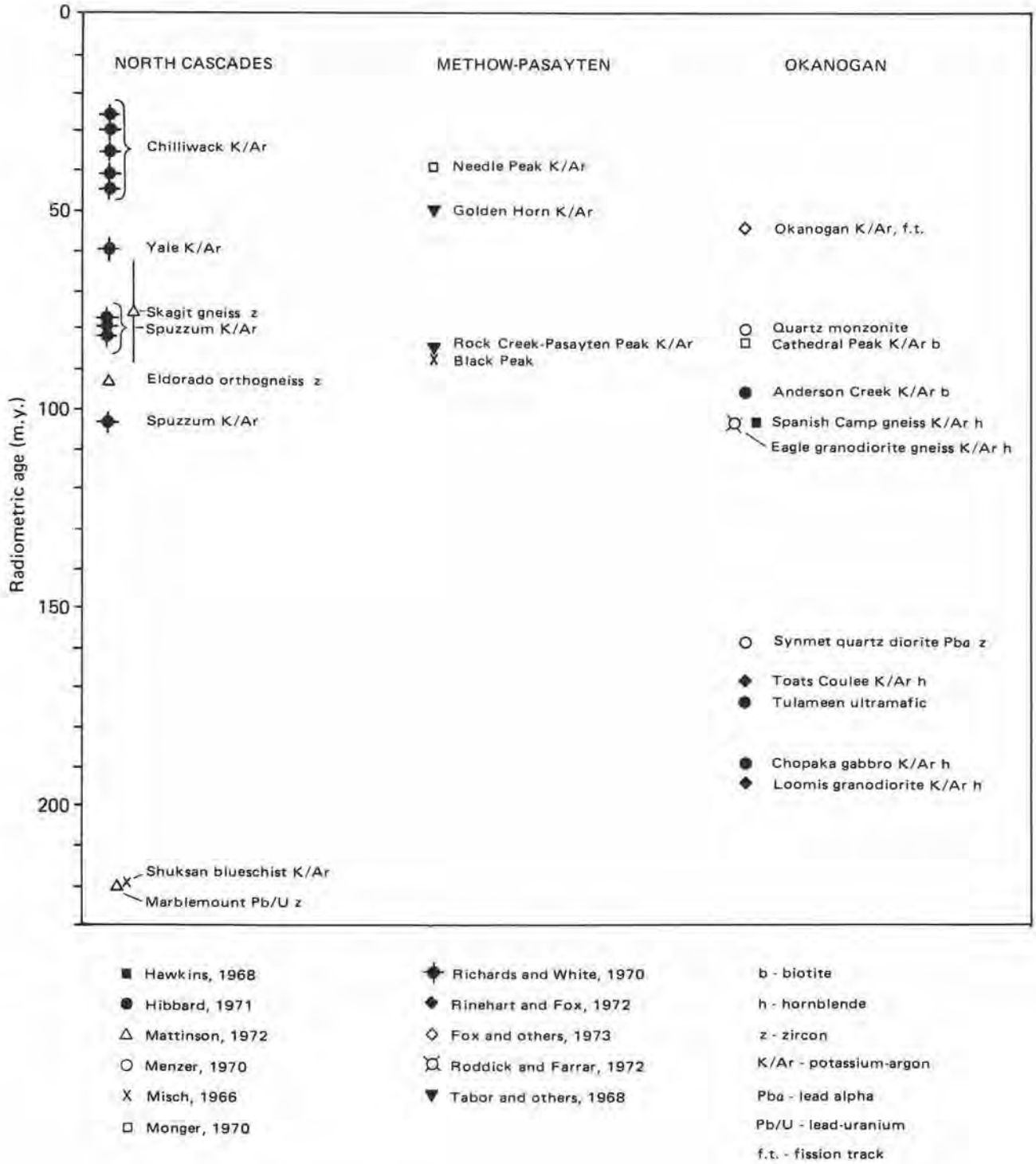

Figure 8.–Radiometric ages from the North Cascades core, Methow-Pasayten belt, and Okanogan terrane. For additional data, see Engels and others (1976) and Fox and others (1977).

the chert-rich detritus in the Methow section. Uplift most likely was a function of the regional compressional deformation which Misch (1966) has shown to have gripped the North Cascades in middle Cretaceous time. Major regional metamorphism evidently began about that time; zircons from the Skagit Gneiss suggest an age of 60 to 90 m.y. for metamorphism (Mattinson, 1972). Intrusion accompanied the metamorphism; Cascade plutons yield radiometric ages from about 100 m.y. through Late Cretaceous and Tertiary. Significantly, granitic detritus is not present in Cretaceous sediment of Cascade provenance in the Methow section, which reinforces the inference that the Cascade infrastructure was still in early stages of development. The middle Cretaceous deformation is apparently the same event that affected the Methow rocks and the Okanogan terrane. Existing information, then, implies that the Okanogan infrastructure dates from Triassic-Jurassic time, while that in the Cascades dates only from the Late Cretaceous. The later event seems to have affected both terranes, whereas the earlier event appears not to be recorded in the Cascades.

The Methow-Pasayten belt is part of a string of Mesozoic successor basins along the length of the Canadian Cordillera. It belongs to the Jurassic-Cretaceous Bowser tectonostratigraphic assemblage (Eisbacher, 1974), and it lies within the Tyaughton trough (Jeletzky and Tipper, 1968), interpreted as a sediment prism shed from the rising Omineca crystalline belt in the Columbian orogen (Eisbacher, 1974), an interpretation which fits the pre-Cenomanian part of the Methow sequence well. The Methow section has a striking resemblance in most respects to the Great Valley sequence in California, even down to such details as the Albian age of the shift from volcanic to arkosic sands (Dickinson and Rich, 1972; Cole and others, 1973). Both sequences were deposited on submarine fans seaward of a Jurassic-Cretaceous magmatic arc. They have divergent evolutions beginning in early Late Cretaceous time, however, as the Great Valley remained essentially open to the west throughout the Late Cretaceous, while the Methow basin filled and was walled off by the rising Cascade orogen, from which it received sediment. Late Cretaceous plutonic-rock ages in the Cascades are similar to those in the Sierra Nevada (Lanphere and Reed, 1973), but plutonism did not shift westward in the Sierra Nevada, as it apparently did in Washington.

A complete explanation for the abrupt appearance of a western source is beyond the scope of this paper (see Hamilton, 1978), but it is a major question in the regional tectonic story. The western boundary-fault system of the graben was active about the same time as deposition of the Virginian Ridge chert-rich facies, but activity on this fault might also be interpreted as a subordinate component of a regional event. Clearly, the sense and amount of slip on the western boundary-fault system must be independently established before the problem can be fully evaluated. The Hozameen fault has been described as a Jurassic suture (Anderson, 1976). Our data, however, argue for a middle Cretaceous rather than Jurassic age for major tectonism. Permian to Middle Jurassic ages from Hozameen and Bridge River terranes (Tennyson and others, 1982; Haugerud, 1984; Potter, 1986) require only that accretion of the Hozameen-Bridge River complex took place in Late Jurassic or Early Cretaceous time. Subsequent middle Cretaceous accretion of Wrangellia farther west probably drove uplift of the Hozameen-Bridege River complex and activity on the Hozameen fault.

Although reconstruction of the Late Cretaceous history of the North Cascades core is outside the focus of our investigation, it is worth pointing out that the Methow sequence unequivocally demonstrates that an oceanic environment of evident regional extent lay west of the Okanogan terrane until middle Cretaceous time, when two important things happened. First, volcanism evidently ceased in the eastern source area of the Methow sediments. Second, a positive area of chert and mafic volcanic rocks, and lacking granitic rocks, appeared on the present site of the North Cascades core. At about the same time, North Cascades deformation and plutonism, Skagit metamorphism (Mattinson, 1972), and a major phase of intrusion of the Canadian Coast Plutonic Complex began (Roddick and Hutchison, 1974). These events may indicate the arrival of the complex of allochthonous terranes which compose the Insular Belt of British Columbia (Wrangellian superterrane). Strike-slip juxtaposition (Monger and Ross, 1971; Monger and others, 1972; and others) is another possibility, but we are hesitant to draw conclusions on the basis of existing data. The story preserved in the Methow sediments hints, in any case, that Late Cretaceous arc-type tectonism in the Cascades Coast Plutonic Complex constituted an episode fundamentally separate from earlier Mesozoic Okanogan arc activity farther east.

ACKNOWLEDGMENTS

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Editor's note: Although the spelling Hozameen is used in some literature, including this reprinted paper, the U.S. Geological Survey uses Hozomeen, which is the spelling of the range in Canada and the mountain in Washington.

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Rb-Sr AND K-Ar DATING OF MID-MESOZOIC BLUESCHIST AND LATE PALEOZOIC ALBITE-EPIDOTE-AMPHIBOLITE AND BLUESCHIST METAMORPHISM IN THE NORTH CASCADES, WASHINGTON AND BRITISH COLUMBIA, AND Sr-ISOTOPE FINGERPRINTING OF EUGEOSYNCLINAL ROCK ASSEMBLAGES

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ABSTRACT

K-Ar and Rb-Sr mineral dates and outcrop-scale (tens to hundreds of meters) whole-rock Rb-Sr isochron dates for rocks of the Shuksan Metamorphic Suite and nearby garnet-amphibolite and barroisite-muscovite schist range from 164 million to less than 100 million years, indicating a Late Jurassic minimum age for blueschist metamorphism and post-Jurassic Ar loss or Sr re-equilibration for various causes. K-Ar and Rb-Sr dates for Vedder Complex and Garrison schist are mostly in the range from 238 million to 285 million years, with a few values scattered to 167 million years, indicating a Late Carboniferous to Early Permian minimum age for albite-epidote-amphibolite, locally associated with blueschist, metamorphism. In addition to distinction in metamorphic age, the rock groups are distinguished by their whole-rock Rb-Sr analyses. Except at low Rb/Sr ratios (where all rocks, regardless of age or association, have overlapping present-day Sr isotopic ratios of 0.703 to 0.707), the Vedder rocks have higher $^{87}Sr/^{86}Sr$ ratios. Thus the Shuksan rocks cannot be reset Vedder Complex rocks.

Whole-rock Sr of Shuksan Suite rocks is identical in distribution on a $^{87}Sr/^{86}Sr$ vs $^{87}Rb/^{86}Sr$ diagram with that of Jurassic and Lower Cretaceous Franciscan rocks of California and Oregon. Sr in Upper Jurassic and Lower Cretaceous Nooksack Group rocks is somewhat less radiogenic, and the analyses of these rocks are less scattered on an isochron diagram, as are analyses of the Great Valley rocks of California. The Leech River Complex, attached to Wrangellia and thus several tectonic units west of the North Cascades, has a Franciscan-Shuksan whole-rock Sr fingerprint but Cenozoic metamorphic overprint. The few currently available analyses of Chilliwack Group rocks show a Sr isotopic fingerprint resembling that of the Franciscan and Shuksan suites, in contradiction with fossil evidence of a distinctly greater age.

INTRODUCTION

This paper summarizes our efforts to date multiple metamorphic episodes and eugeosynclinal rock suites in the North Cascades, west of the Straight Creek fault in Washington and southern British Columbia. That north-south-trending fault (Fig. 1), first recognized by Vance (1957) to the south of the Skagit River valley and shown by Misch (1966) to continue north across the Skagit into British Columbia, separates the mostly Mesozoic Skagit Crystalline Core on the east from a folded stack of thrust plates on the west (Misch, 1966). The geology of the entire region has been reviewed by Misch (1966, 1977) based on many years of study by himself and his students. More recent contributions to metamorphic petrology and detailed re-mapping of critical areas have come from Brown (1974, 1977) and his students (for example, Brown and others, 1981; Haugerud and others, 1981).

Lowest in the stack of Northwest Cascades thrust plates, below the Church Mountain thrust (Fig. 1), are low-grade Upper Paleozoic to Lower Cretaceous rocks studied and named by Misch (1966). Most important for our discussion are the Middle Jurassic Wells Creek Volcanics and the overlying Upper Jurassic to Lower Cretaceous, volcanic-arc-derived Nooksack Group exposed in the Mount Baker window. Related rocks are present in the Fraser Valley, the San Juan Islands, and southwest of Darrington.

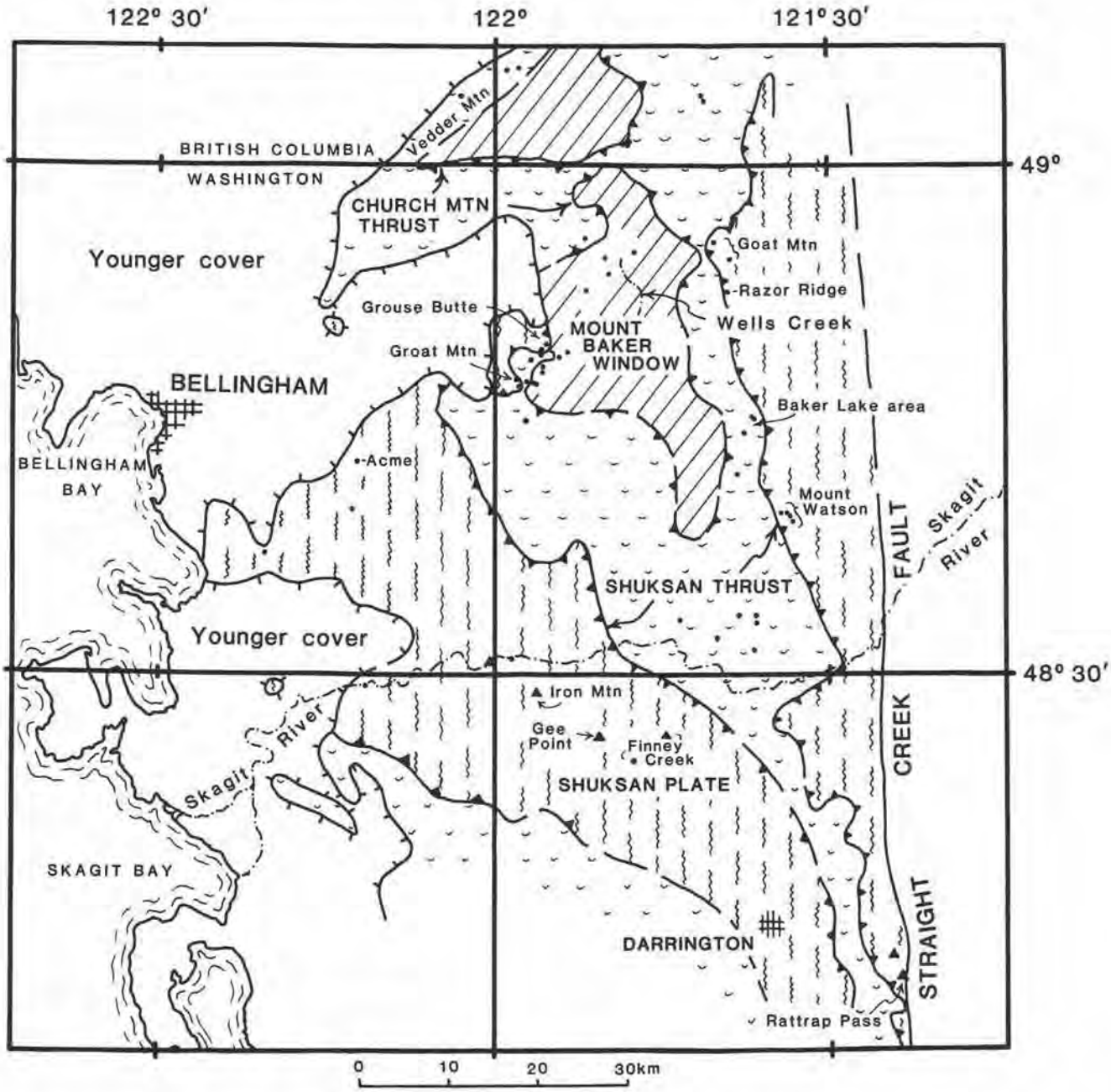

Figure 1.–Map of the Northwest Cascades showing distribution of major rock units and sample localities.

Above the Church Mountain thrust and below the Shuksan thrust, the Church Mountain plate consists in large part of eugeosynclinal volcanic and sedimentary strata of the Chilliwack Group (named by Daly, 1912) on the north side of the International Border and assigned by him to the Carboniferous. This group was traced southward to the Skagit River valley by Misch (1952, 1966), where it was shown by Danner (1957, 1966) to range from Devonian to Permian; it was restudied in the type area by Monger (1966). Laterally extensive outcrops of Chilliwack have stratigraphic continuity, as have parts of more subordinate Mesozoic eugeosynclinal strata (including Nooksack Group equivalents) in the plate. All of these units have been metamorphosed in a high-pressure variety of prehnite-pumpellyite facies during the mid-Cretaceous orogeny, as have the rocks below the Church Mountain thrust (Misch, 1966, 1971). A second major constituent of

the Church Mountain plate is the imbricate zone (Misch, 1962, 1966, 1980) which tends to occupy the upper portion of the plate but whose share in the plate ranges form nearly zero to 100 percent. This zone consists of small to large thrust slices of a variety of rocks, telescoped from originally separate belts and brought together by the mid-Cretaceous (plus some inferred earlier) thrusting. The rocks involved in the imbricate zone comprise the Paleozoic and Mesozoic strata referred to above, plus previously metamorphosed units including the Precambrian and "Caledonian", high-grade metaplutonic, primitive-continental-crustal Yellow Aster Complex, and the Late Paleozoic, meta-igneous and metasedimentary Vedder Complex of albite-epidote-amphibolite facies, locally associated with somewhat later blueschist metamorphism. Altered peridotites are present along thrust planes.

The highest structural unit in the area is the Shuksan plate, composed of Darrington Phyllite and Shuksan Greenschist of the Shuksan Metamorphic Suite of Misch (1966). This unit displays blueschist facies metamorphism that predates the stacking of the thrust plates. Eocene (Johnson, 1984) sediments overlie the thrust stack unconformably and were folded and faulted together with the thrust plates during the Eocene (Misch, 1966; Tabor and others, 1984). Tertiary igneous rocks cut through and overlie the thrust plates.

North Cascade geochronometry began with K-Ar dating done by J. L. Kulp at Lamont Geological Observatory for a joint project with Misch (1964, 1966; Engels and others, 1976) on Darrington Phyllite (samples PM 16, 17 and 18, giving dates of 115 ± 3, 110 ± 4 and 107 ± 3 m.y., respectively), on blueschist (PM 24, giving 265 ± 8 m.y.) from Groat Mountain (a rock at that time called Shuksan because it duplicated the lithology of typical Shuksan crossite schist, but now assigned to the Vedder Complex because of its greater age), and several metamorphic rocks of the Skagit Crystalline Core (~45 m.y.) and granitic plutons of the North Cascades. At the time that work was done, it was not clear how well it touched on the themes that have emerged with further work on North Cascade rocks—all the Late Paleozoic, Mesozoic, and Eocene metamorphic episodes were recognized. The Cretaceous K-Ar dates were interpreted by Misch as a result of resetting during the mid-Cretaceous thrusting event. The Paleozoic age for a tectonic slice of blueschist at the sole of the Shuksan plate north of Groat Mountain was duplicated by a crossite K-Ar date (PM 9, giving 246 ± 41 m.y.) done by Geochron Labs. Inc. for a blueschist from upper Finney Creek in the interior of the coherent Shuksan plate. Given those results, Misch correlated the Permian Groat Mountain blueschist slice with the main Shuksan. Later work was unable to duplicate the Finney Creek date (note its original large analytical uncertainty), the same and nearby rocks yielding only Mesozoic K-Ar dates when run at much better precision in different laboratories (Table 3 and Brown and others, 1982). The Paleozoic age of the Groat Mountain blueschist was confirmed by Armstrong and others (1983).

The major U-Pb zircon dating study of Mattinson (1972) was largely concerned with the Cascade Crystalline Core and its southeastern extension near the Columbia River, but it also established Precambrian and Early Paleozoic ages in the Yellow Aster Complex of the Northwest Cascades, confirming the great age inferred on the basis of geologic evidence by Misch (1966).

Our joint studies began in the late 1970s, but before they were complete, geologists for the proposed Skagit Valley Nuclear Site instigated a crash K-Ar dating program (Sayre and others, 1979) that provided numerous analyses of Shuksan Suite whole rocks, mostly Darrington Phyllite, which gave Cretaceous dates clustered around 125 m.y. Whetten and others (1980) and Vance and others (1980) compiled geochronometry from a variety of sources in support of a common Late Jurassic age for most ophiolitic rocks in northwestern Washington, including some closely associated with the Shuksan thrust.

Geochronometry done at the University of British Columbia on rocks collected by Western Washington University students has been presented by Brown and others (1982)—on the Jurassic-Early Cretaceous Shuksan metamorphism, and by Armstrong and others (1983)—on the Late Paleozoic metamorphism of the Vedder Complex. Additional results presented here strengthen the case made in those papers for *two* blueschist metamorphic events; in addition, we discuss numerous new whole-rock Sr isotopic data for Shuksan and related rocks.

The analytical methods used for Rb-Sr and K-Ar at the University of British Columbia are conventional ones, described in Armstrong and others (this volume). It suffices to say here that decay constants used for all results discussed in this paper are those recommended by the I. U. G. S. Geochronometry Subcommission (Steiger and Jäger, 1977) and Sr isotopic analyses are adjusted so that the E and A and NBS Sr standards give $^{87}Sr/^{86}Sr$ ratios of 0.70800 and 0.71022, respectively. Raw Rb-Sr data are presented in Table 1, calculated Rb-Sr dates in Table 2, K-Ar data and dates in Table 3.

Table 1.–Rb-Sr data for rocks and minerals of the Northwest Cascades, San Juan Islands, and Leech River area*

Sample number*	Rock type	Latitude, Longitude	Material analyzed	Sr (ppm)	Rb (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
		Darrington Phyllite					
Alger	Medium dark-grey phyllite	48° 36.7' 122° 20.0'	R	78.1	57.8	2.14	0.7092
PM 1	Dark-grey phyllitic slate	48° 29.3' 122° 11.7'	R	65.7	40.5	1.786	0.7079
PM 23	Crumpled dark-grey phyllite with silty laminae	48° 39.7' 122° 12.6'	R	51.6	74.1	4.16	0.7135
Acme P	Dark-grey crinkled phyllite	48° 42.5' 122° 12.1'	R	97.8	119	3.54	0.7127
Acme S	Medium-grey phyllitic siltstone	48° 42.5' 122° 12.1'	R	413	23.8	0.167	0.7067
RH78 D19	Dark-grey crinkled phyllite	48° 38.9' 121° 33.2'	R	94.8	88.3	2.70	0.7120
RH78 E1	Medium-grey phyllite	48° 39.4' 121° 33.6'	R	85.9	132	4.44	0.7154
Baker 7	Massive, dark-grey phyllitic argillite	48° 48.85' 121° 55.90'	R	85.2	78.6	2.67	0.7102
Baker 9	Medium-grey phyllitic slate	48° 49.15' 121° 55.45'	R	98.3	47.4	1.396	0.7080
PR85-237 A	Dark-grey phyllitic slate	48° 47.25' 121° 57.65'	R	93.0	53.2	1.656	0.7082
PR85-237 B	Medium-grey greywacke	48° 47.25' 121° 57.65'	R	131	36.9	0.818	0.7063
PR85-254	Dark-grey slate with sandy laminations	48° 47.05' 121° 57.4'	R	165	52.8	0.925	0.7070
TLR1	Dark-grey massive phyllitic argillite	48° 56.85' 121° 38.4'	R	210	73.5	1.015	0.7080
GMR-B	Dark-grey crinkled phyllite	48° 55.3' 121° 40.4'	R	50.6	54.9	3.14	0.7116
GMR-D	Medium-grey thinly laminated phyllite	48° 55.3' 121° 40.4'	R	59.2	94	4.60	0.7139
GMR-E	Medium-grey siliceous phyllite with disrupted laminae	48° 55.3' 121° 40.4'	R	34.0	83.5	7.10	0.7191
GMT-B1	Medium-grey, crinkled, finely laminated siliceous phyllite	48° 54.4' 121° 38.8'	R	73.1	100	3.98	0.7129

Table 1.–Rb-Sr data (continued)

Sample number	Rock type	Latitude, Longitude	Material analyzed	Sr (ppm)	Rb (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
		Darrington Phyllite (cont'd)					
GMT-D	Medium-grey crinkled, finely laminated siliceous phyllite	48° 54.4' 121° 38.8'	R	65.9	124	5.45	0.7142
RR-B	Medium-grey lustrous phyllite	48° 52.5' 121° 39.0'	R	16.5	13.4	2.35	0.7098
RR-C	Medium-grey siliceous phyllite	48° 52.5' 121° 39.0'	R	89.0	33.3	1.084	0.7081
RR-D	Dark-grey scaly phyllite	48° 52.5' 121° 39.0'	R	107	130	3.53	0.7120
RR-E	Dark-grey crinkled phyllite	48° 52.5' 121° 39.0'	R	59.6	117	5.69	0.7152
		Shuksan blue- and greenschists					
PM 2	Greenish-grey phyllitic blueschist	48° 30.7' 122° 0.0'	R	104	7.5	0.208	0.7041
PM 2a	Greenish-olive-grey muscovite schist with coarse amphibole porphyroblasts	48° 28' 121° 56'	R	134	52.8	1.137	0.7062
			E	444	13.1	0.085	0.7044
			A	32.3	4.3	0.380	0.7048
			M	49.5	126	7.41	0.7205
RH78 E69	Micaceous blueschist	48° 39.4' 121° 33.7'	R	140	62.7	1.30	0.7057
RH78 E78	Blue amphibole-rich schist	48° 39.7' 121° 34.3'	R	84.1	0.6	0.021	0.7038
JV-Rattrap	Crossite schist	48° 12.0' 121° 23.4'	R	94.7	17.0	0.520	0.7049
JV-Shuksan 3	Crossite schist	48° 26.3' 121° 44.8'	R	118	4.9	0.120	0.7043
JV-A877	Greenschist	48° 13.5' 121° 24'	R	106	0.1	0.003	0.7034
		Nooksack Group					
PM 28	Slate	48° 54.7' 121° 47.5'	R	112	33.6	0.865	0.7062
B 1	Pencil slate	48° 47.9' 121° 55.5'	R	77.6	45.2	1.686	0.7079

Table 1.–Rb-Sr data (continued)

Sample number	Rock type	Latitude, Longitude	Material analyzed	Sr (ppm)	Rb (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
		Nooksack Group (cont'd)					
B 3s	Shale with silty laminae	48° 48.25' 121° 55.65'	R	194	57.8	0.861	0.7060
B 3g	Greywacke	48° 48.25' 121° 55.65'	R	214	29.1	0.393	0.7055
B 9	Argillite	48° 48.9' 121° 53.6'	R	285	48.9	0.495	0.7052
B 13	Shale	48° 47.15' 121° 56.35'	R	65.9	42.4	1.860	0.7074
B 14	Pencil slate	48° 47.1' 121° 56.25'	R	66.5	41.5	1.807	0.7076
JS74-37	Sandstone	48° 52.5' 121° 51.9'	R	162	51.2	0.914	0.7054
JS74-12	Sandstone with mica flakes	48° 48.75' 121° 54.2'	R	159	43.2	0.787	0.7064
JS74-69c	Volcaniclastic sandstone	48° 53.5' 121° 49.7'	R	384	26.1	0.197	0.7049
JS56	Coarse volcaniclastic grit	48° 54.35' 121° 49.9'	R	338	33.0	0.282	0.7054
		Leech River Complex					
LR ph	Pelite	48° 32.9' 124° 21.6'	R	170	67.3	1.15	0.7075
LR q	Sandstone	48° 32.9' 124° 21.6'	R	237	40.5	0.495	0.7071
LR sl	Pelite	48° 32.9' 124° 21.6'	R	246	72.4	0.852	0.7072
LFSSE-13	Sandstone	48° 30.6' 123° 49.45'	R	254	40.6	0.463	0.7065
LF5-1-6	Sandstone	48° 30.95' 123° 52.7'	R	424	32.9	0.225	0.7064
LF4-26-12	Sandstone	48° 31.75' 123° 53.05'	R	543	48.2	0.257	0.7058
LFMME-18-J11	Pelite	48° 33.3' 123° 48.0'	R	325	43.1	0.384	0.7062
LF5-1-1	Pelite	48° 30.5' 123° 52.3'	R	191	86.6	1.314	0.7085
LF5-5-13	Pelite	48° 30.5' 123° 58.7'	R	251	74.5	0.861	0.7074

Table 1.–Rb-Sr data (continued)

Sample number	Rock type	Latitude, Longitude	Material analyzed	Sr (ppm)	Rb (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
		Leech River Complex (cont'd)					
LF4-10-1	Pelite	48° 34.1' 123° 49.2'	R	175	100	1.66	0.7080
LF4-29-3	Greenstone	48° 34.9' 123° 56.1'	R	190	28.6	0.435	0.7045
LFMME-156-J11	Greenstone	48° 33.45' 123° 48.8'	R	160	1.2	0.021	0.7052
LF5-2-8	Greenstone	48° 30.95' 123° 55.85'	R	226	28.6	0.365	0.7037
		Vedder Complex					
PM 101	Amphibole-muscovite-albite schist	48° 45.1' 121° 36.9'	R	264	33.8	0.371	0.7069
	Subcalcic hornblende with glaucophane and minor actinolite		H	55.6	16.7	0.871	0.7083
	Phengitic muscovite		M	331	195	1.71	0.7123
PM 102 a	Albite-amphibole-muscovite-quartz schist	48° 45.1' 121° 36.9'	R	68.7	35.3	1.49	0.7097
PM 103	Albite-amphibole-muscovite schist	48° 45.1' 121° 36.9'	R	113	35.8	0.913	0.7086
PM 104	Albite-bearing amphibole-white mica schist	48° 45.1' 121° 36.9'	R	73.8	57.2	2.24	0.7125
PL90-8	Albite-muscovite schist	48° 45.0' 121° 36.7'	R	211	51.2	0.704	0.7084
			M	52.1	230	12.82	0.7554
PL2-106	Blueschist block	48° 44.1' 121° 37.6'	R	372	7.2	0.056	0.7055
		Garrison Schist					
MB8054J-1	Albite-epidote-amphibole schist	48° 27.65' 123° 1.92'	R	114	1.5	0.037	0.7049
JVO-507	Albite epidote amphibolite	48° 39.1' 122° 51.9'	R	361	3.4	0.027	0.7052
MB7981J-1C	Albite-epidote-amphibole schist	48° 33.97' 122° 58.72'	R	168	1.4	0.025	0.7043

Table 1.–Rb-Sr data (continued)

Sample number	Rock type	Latitude, Longitude	Material analyzed	Sr (ppm)	Rb (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
		Chilliwack Group					
B4	Greywacke	48° 48.2' 121° 56.5'	R	120	15.6	0.377	0.7066
BC72-37A	Greywacke	48° 33.05' 121° 43.45'	R	84.7	46.4	1.59	0.7100
BC72-43	Greywacke	48° 31.2' 121° 36.7'	R	91.8	32.0	1.01	0.7077
BC72-78-2	Greywacke	48° 31.9' 121° 40.0'	R	133	17.7	0.384	0.7064
BC72-157B	Greywacke	48° 33.4' 121° 38.65'	R	186	16.3	0.253	0.7065
BC72-52C	Argillaceous greywacke	48° 32.7' 121° 36.8'	R	63.1	37.8	1.73	0.7091
TLR-2	Slate with silty laminae	48° 56.8' 121° 38.55'	R	136	63.7	1.351	0.7073
BC72-32	Meta-andesite	48° 55.8' 121° 51.8'	R	183	8.6	0.136	0.7048
BC72-82	Meta-andesite	48° 33.1' 121° 36.9'	R	53.5	42.5	2.30	0.7108
JM 5357	Keratophyre	49° 01.1' 121° 48.0'	R	40.8	5.9	0.418	0.7063
JM 5360	Keratophyre	49° 01.4' 121° 48.6'	R	29.4	0.5	0.049	0.7052
JM 5361	Greenstone	49° 01.2' 121° 47.9'	R	70.5	0.7	0.029	0.7050
JM 5366	Greenstone	49° 04.1' 121° 41.0'	R	319	1.0	0.0091	0.7054
JM 5367	Greenstone	49° 04.0' 121° 40.75'	R	254	2.1	0.024	0.7049
B6	Greenstone	48° 48.75' 121° 56.9'	R	204	1.9	0.028	0.7048

*Samples collected by Peter Misch or in his company have PM prefix. Samples provided by other workers are indicated by prefixes as follows: RH, Ralph Haugerud; PR, Paul Rady; JV, Joe Vance; JS, Jon Sondergaard; LF, Lee Fairchild; PL, Peter Leiggi; MB, Mark Brandon; BC, Bruce Christenson; JM, Jim Monger. All others collected by Richard Armstrong.

Table 2.–Rb-Sr dates and Sr isotope initial ratios for mineral isochrons and local whole-rock suites

Suite	Date (m.y. ± σ)	$^{87}Sr/^{86}Sr$ (initial ratio ± σ)	MSWD
Darrington Phyllite			
Acme (PM23, Acme P, Acme 5)	122 ± 4	0.7064 ± 0.0001	1.4
Mount Watson (RH78 D19, RH78 E1)	137	0.7067	
W. of Mount Baker-Grouse Butte area (Baker 7, Baker 9)	122	0.7056	
Groat Mountain (PR85-237A, PR85-237B, PR85-254)	146 ± 42	0.7048 ± 0.0007	10.4
Goat Mountain -W near Twin Lakes (TLR-1, GMR-B, GMR-D, GMR-E)	124 ± 5	0.7061 ± 0.0002	3.4
Goat Mountain -S near Ruth Creek (GMT-B1, GMT-D)	62	0.7094	
Razor Ridge (RR-B, RR-C, RR-D, RR-E)	110 ± 5	0.7063 ± 0.0002	2.1
Shuksan blue- and greenschist			
Iron Mountain (PM 2a Minerals and Whole Rock)	153 ± 8	0.7040 ± 0.0002	6.0
(PM 2a Muscovite - Whole Rock)	160	0.7036	
(PM 2a amphibole - Whole Rock)	130	0.7041	
Mount Watson (RH78 E69, RH78 E78)	105	0.7038	
Leech River Complex			
Port Renfrew (LR ph, RL q, R L sl)	42 ± 15	0.7068 ± 0.0002	0.9
Vedder Complex			
Baker Lake (PM 101 Minerals and Whole Rock)	280 ± 39	0.7052 ± 0.0005	16.5
(PM 101 Muscovite - Whole Rock)	283	0.7054	
(PM 101 Amphibole - Whole Rock)	197	0.7059	
Baker Lake (PL90-8 Muscovite - Whole Rock)	273	0.7057	

Table 3.–K-Ar dates

Sample number	Rock type	Latitude, longitude	Material analyzed	% K	Radiogenic ^{40}Ar (cc x 10^6)	% Radiogenic	Date (m.y. ± σ)
			Shuksan blue- and greenschist				
PM 2a	Muscovite schist with amphibole porphyroblasts	48° 28' 121° 56'	MS	8.15	49.54	91.4	150 ± 5
			AM	0.251	1.465	72.1	144 ± 5
RH 78 E69	Micaceous blueschist	48° 39.4' 121° 33.7'	WR	2.12	10.362	94.3	122 ± 4
RF75096*	Amphibolite[1]	48° 10' 121° 32'	AM	0.573	0.3272	38.7	141 ± 6
RF75098* -RF75116*	Blueschist[2]	48° 12.0' 121° 23.4'	WR	0.447	1.7463	79.9	97.2 ± 2
RF75120*	Crossite schist[3]	48° 26.3' 121° 44.8'	AM	0.0921	0.3854	27.4	104.5 ± 7
			Vedder Complex				
PM 101	Albite-muscovite-amphibole schist	48° 45.1 121° 36.9'	MS	6.15	73.22	96.4	283 ± 9
			AM	0.583	6.297	84.9	259 ± 9
PL90-8	Albite-muscovite schist	48° 45.0' 121° 36.7'	MS	7.79	89.50	95.8	274 ± 9
RH78 D61	Quartz-mica semischist	48° 41.7' 121° 38.2'	MS	4.31	34.66	97.1	196 ± 7
			Garrison Schist				
JVO-507	Albite-epidote amphibolite	48° 39.1' 122° 51.9'	AM	0.499	6.001	89.7	286 ± 10
MB7981J-1C	Albite-epidote-amphibole schist	48° 33.97' 122° 58.72'	AM	0.132	0.8967	62.6	167 ± 6
S161*	Amphibolite	48° 31.3' 123° 8.0'	AM	0.523	5.261	91.8	242 ± 7
			Chilliwack Group				
BC72-202B	Metamorphosed granitic clast	48° 31.8' 122° 40.6'	WR	1.146	7.222	85.8	155 ± 7

* Analysis by D. L. Turner, Univerisity of Alaska
1 Amphibolite of Helena Ridge (Vance, 1957; Vance and others, 1980)
2 Blueschist 1 km north of Rattrap Pass (Vance, 1957)
3 Crossite schist, Finney Creek locality for PM 9 date of 246 ± 41 m.y. by Geochron Labs.

DARRINGTON PHYLLITE

The mid-Cretaceous reconnaissance K-Ar dates reported by Misch (see above) were interpreted as reset during the mid-Cretaceous thrusting event, because this orogeny *postdates* Shuksan metamorphism. Shuksan Suite rocks suffered cataclasis, mylonitization and destructive retrogression at this time, and previously largely unmetamorphosed Paleozoic and Mesozoic rocks, now beneath the Shuksan plate, were recrystallized in prehnite-pumpellyite facies. All Darrington Phyllite samples examined by Misch have undergone various degrees of renewed penetrative deformation during the mid-Cretaceous event, and none of that later fabric is annealed. Recrystallization in the Shuksan plate was restricted to growth of prehnite and pumpellyite on a very local scale where the rocks were severely broken up next to thrust contacts.

We have analyzed a large suite of whole-rock samples, including multiple samples from individual localities scattered across the length and breadth of Darrington exposures. Figure 2 shows that the results plot in a band about 0.003 wide in $^{87}Sr/^{86}Sr$, with initial ratio about 0.7055, ranging from near zero to 7 in $^{87}Rb/^{86}Sr$ ratio, and having a slope corresponding to a Rb-Sr date of about 130 m.y.

Much better isochrons are observed for locality suites—two or more samples from within ten to hundreds of meters of each other. These local isochrons give dates ranging from ~62 to 146 ± 42 m.y.(Table 2), with most between 110 and 137 m.y., and averaging 117 m.y. These are essentially identical to mineral dates by K-Ar or Rb-Sr for the same or associated rocks. The one distinctly younger whole-rock date, 62 m.y., comes from an area near whole-rock K-Ar dates of 52 to 73 m.y. for greywacke and greenstone beneath the Shuksan Plate along the Mount Baker Highway (Sayre and others, 1979), suggesting resetting caused by nearby Cenozoic igneous rocks. Misch has mapped quartz diorite along Wells Creek that may be the local heat source.

The general concordance of whole-rock and mineral systems implies that isotopic re-equilibration of minerals on a sub-centimeter scale and of whole-rock systems on a scale of tens of meters continued together into Early Cretaceous time in fine-grained rocks nearly everywhere in the Shuksan thrust plate and was rapidly terminated, perhaps by mid-Cretaceous emplacement of the Shuksan plate over cooler rocks, before the plate reached its present location by further brittle movements. Some Cenozoic resetting has occurred locally.

Whether the depositional age of metasedimentary rocks can be determined by whole-rock Rb-Sr dating is controversial. Our evidence for large-scale metamorphic resetting would suggest a negative response and certainly demands caution. Some previous work on rocks in other parts of the world has demonstrated overwhelming resetting so that the depositional age is erased (Peterman, 1966; Gebauer and Grünenfelder, 1974), but other studies have reached more optimistic conclusions, especially where low-grade or unmetamorphosed sediment is concerned (Bofinger and others, 1968; Moorbath, 1969; Clauer, 1976, 1979; Cordani and others, 1978).

When cogenetic samples of clastic sediment become open systems, with Rb, Sr, and Sr-isotopic exchange among themselves, an original isochron for the time of deposition (or for the time of diagenetic reconstitution of sediment minerals) is altered. Samples having high Rb/Sr ratios tend to lose radiogenic Sr easily, and those having low Rb/Sr ratios (which are usually richer in Sr) tend to acquire radiogenic Sr. Sample analyses will become scattered about the original isochron, and the apparent isochron will rotate toward a slope corresponding to the time of metamorphism. The rotation will be about a point of average composition of the suite; the initial ratio will increase significantly for any suite which has a Rb/Sr ratio as high as average crustal values or above. Extreme resetting will result in an *approach* to a purely metamorphic isochron, but the evidence of an older age will be preserved in the elevated initial ratio. We have simulated this behavior on computer model systems that we feel reasonably represent real rocks and processes.

It is important that the rocks analyzed be representative of the overall rock system and not biased towards a few domains having high Rb/Sr ratios in a larger body of rock with lower Rb/Sr ratios. In that special case, the isochron rotation will be about a point near the isochron intercept, and the intercept will not provide a clue to the amount of resetting—only the metamorphism will be dated. We think that some previous attempts to date sediments, where only the metamorphic resetting was observed, involved such a sampling bias in favor of high Rb/Sr samples. In sampling the Darrington Phyllite we collected typical specimens of large, apparently homogeneous units, and samples selected for analysis were spread over the range of Rb/Sr ratios at each locality but not strongly shifted towards a few samples having higher Rb/Sr ratios and against samples having low Rb/Sr ratios and higher Sr concentrations. We thus feel that our data give a representation of the Sr isotopic character of the whole unit, and not just high Rb/Sr domains within it.

Tempering the view that sedimentation ages can be determined by Rb-Sr dating, but offering encourag-

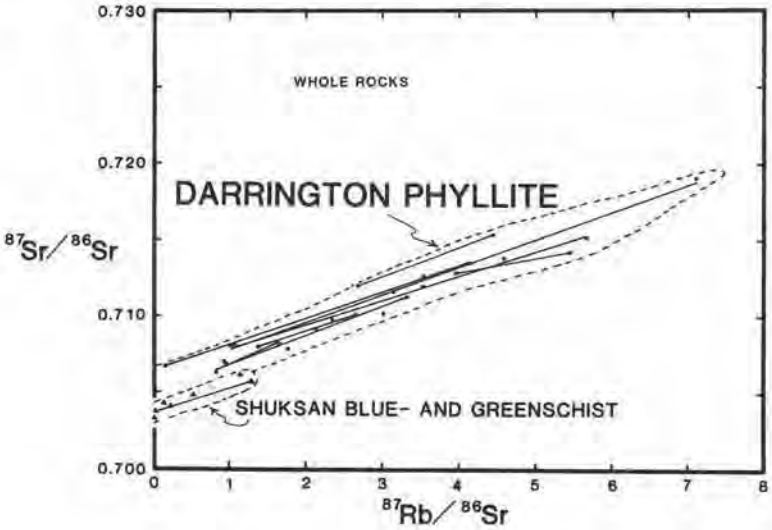

Figure 2.–Rb-Sr isochron plot of Shuksan Suite whole rocks. Dashed outlines enclose fields for the Darrington Phyllite and Shuksan blue- and greenschist. Solid lines are isochrons (Table 2) for local sample suites that show much less scatter than the whole suite.

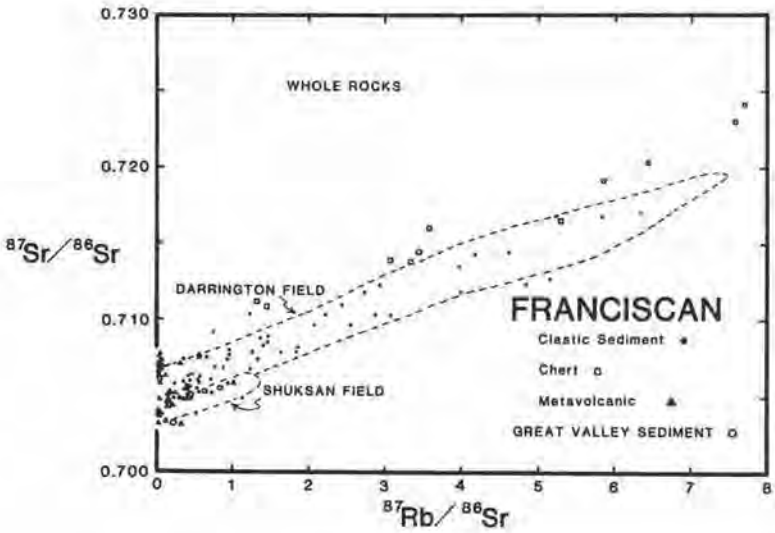

Figure 3.–Isochron plot for greywacke-argillite, chert, and metavolcanic samples of the Franciscan and Great Valley assemblages in California and Oregon. (Compiled from Coleman, 1972; Peterman and others, 1967; Davis, 1969; Sinha and Davis, 1971; Ghent and others, 1973; Suppe and Foland, 1978; Chyi and others, 1982). All isotope ratios have been adjusted to be consistent with a value of 0.7080 for the E and A Sr standard. Clastic sediment is shown as dots, chert as squares, and metavolcanic rock as triangles. Stratigraphic age for the samples analyzed as indicated by fossils ranges from Upper Jurassic through about Middle Cretaceous (165 to 91 m.y.) and metamorphic age based on K-Ar dating ranges from 160 to 80 m.y. (Coleman and Lanphere, 1971; Suppe and Armstrong, 1972). A few analyses of higher grade blueschist blocks (not shown) scatter towards higher $^{87}Sr/^{86}Sr$, and some have even higher Rb/Sr ratios than the clastic sediments shown here. One chert analysis plots to the right of the figure as well. For comparison, the envelopes for Darrington Phyllite and Shuksan green- and blueschist are shown. The coincidence with Franciscan analyses is dramatic. The Darrington envelope has a slope corresponding to a Rb-Sr date of about 130 m.y. and 0.705 initial ratio. The Franciscan cherts scatter about an isochron of about 140 m.y. and 0.708 initial ratio. For the Franciscan, these dates are about the same as the times of **both** deposition and metamorphism.

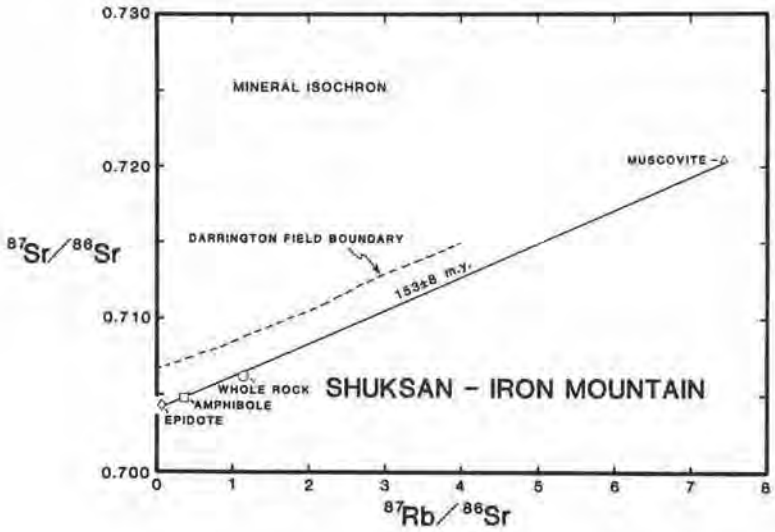

Figure 4.–Rb-Sr isochron plot for minerals and whole-rock of sample PM 2a from Iron Mountain. A portion of the Darrington field boundary is shown by a dashed line.

ing evidence of the resistance of sediment Sr isotopic systems to resetting under some conditions, are reports of provenance isochrons preserved in fresh and metamorphosed sediments (Dasch, 1969; Hofmann and others, 1974; Spanglet and others, 1978). Thus, Sr in metasedimentary rocks may reflect some composite of provenance, age of deposition, and diagenesis, and it may be affected by much later metamorphism. It may be difficult or impossible to assign geologic meaning to an observed correlation of $^{87}Sr/^{86}Sr$ and $^{87}Rb/^{86}Sr$ ratio. However, our own experience (for example, Armstrong and others, 1986, and this volume) is that the crustal residence age cannot be completely erased. Drastic Sr isotope equilibration of old sediment will be indicated by increased initial ratios, and model ages for well-sampled suites [such as the bulk earth model of Cameron and others (1981) which give a date of 156 m.y. for the Darrington Phyllite suite] will give a reasonable, albeit imprecise, estimate of depositional age. On that basis we interpret the whole-rock Sr data to support a Jurassic age of deposition for the Darrington Phyllite, with a Late Jurassic to Early Cretaceous metamorphic overprint.

Quite apart from deriving an age, we can use whole-rock Sr data for well-sampled sedimentary suites as a basis for comparison between suites. Close similarity would be consistent with similar age and history, although several different histories might be contrived to give similar end results (trade-offs being apparent initial ratio vs. age of suite and depositional/diagenetic age vs. degree of resetting in later events). Toward that goal of comparative Sr isotopic signatures, we show in Figure 3 Sr isotopic results for a large number of rocks from the Franciscan Complex of California and Oregon. On the same figure, we show the field for Darrington clastic sedimentary rocks. The isotopic similarity of Franciscan (most of those data coming from the eastern, and thus older parts of the Franciscan with Late Jurassic and Early Cretaceous stratigraphic ages) and Darrington clastic sedimentary rocks is quite dramatic. Beyond the Sr isotope data in Figures 2 and 3, there is also similarity to some extent in metamorphic type and grade and in metamorphic mineral dates (to be discussed later). The identity of Sr isotopic character supports the proposition of overlapping age and similar metamorphic chronology. Lithologically, however, the Darrington, which consists of well-foliated, typically graphitic and quartz-veined, semipelitic and pelitic phyllites with lesser but significant amounts of schistose, plagioclase-arkosic metagreywacke and lacks cherts, is quite different from typical Franciscan. The Darrington is far more similar to the South Fork Schist of southwestern Oregon (M. C. Blake, Jr., oral commun. to Misch).

On Figure 3 the points plotted for Franciscan cherts scatter about an isochron whose slope is similar to that of the Franciscan clastic sedimentary rocks (approximately 150 m.y., a composite of depositional age and metamorphic resetting), but that has a higher initial $^{87}Sr/^{86}Sr$ ratio (0.707 to 0.708 compared to $\sim$0.705 for the clastic rocks). The chert initial ratio is in the range for Jurassic-Early Cretaceous seawater indicating depositional and (or) diagenetic equilibration. The Franciscan clastic initial ratio is lower than the ratio for Mesozoic seawater because the sediments were derived from a volcanic arc source (initially 0.703 to 0.704); a slight ^{87}Sr increase due to finite crustal residence time, an unknown amount of seawater interaction, and perhaps some degree of metamorphic re-equilibration. The Darrington clastic rock initial ratio would have the same rationalization; a predominantly quartz dioritic source is indicated by its sedimentary petrography (Misch, 1966).

SHUKSAN BLUESCHIST AND GREENSCHIST

As already mentioned, in the 1970s it became clear that Shuksan metamorphism was mid-Mesozoic, not Permian. However, the *exact* age of this event remained elusive inasmuch as additional Early and mid-Cretaceous K-Ar dates done for scattered Shuksan Suite samples, mostly of Darrington Phyllite, left it open to debate how much resetting had occurred. In 1976 R. B. Forbes, planning a joint project with Misch, suggested dating by K-Ar of muscovite-blue amphibole pairs, and in 1978 our late colleague G. M. Miller discovered a suitably coarse-grained muscovite-rich blueschist in a coherent stratigraphic section in the interior of the Shuksan plate that was well exposed in a new road cut on the southwest side of Iron Mountain. In 1980 Armstrong obtained Late Jurassic concordant K-Ar dates on muscovite and amphibole from a sample (PM 2a) that Miller and Misch had collected at this locality.

We present here our new Ar and Sr dates for the coarse-grained blueschist, PM 2a, from the Iron Mountain area. Our new K-Ar dates of 150 ± 5 m.y. for muscovite and 144 ± 5 m.y. for blue amphibole (Table 3) are essentially concordant with Rb-Sr mineral-whole-rock dates of 160 and 130 m.y. (Table 2) for the same mineral fractions or, alternatively, a three-mineral plus whole-rock isochron date of 153 ± 8 m.y. (Fig. 4). These dates add to published K-Ar dates of 148 and 164 m.y. from coarse garnet-amphibolite and barroisite-muscovite schist, respectively, at Gee Point, nearby (Brown and others, 1982). Our Iron Mountain dates show that the Jurassic Gee Point dates are reproducible over a large area—even though the

Gee Point samples recrystallized at higher temperatures and before regional blueschist facies conditions. We can be certain that the metamorphism producing the coarse blueschists occurred during or before Kimmeridgian time and the Gee Point metamorphism was earlier, but perhaps not much older. As observed elsewhere (Coleman and Lanphere, 1971; Suppe and Armstrong, 1972), coarse-grained white mica is more retentive of radiogenic daughter products than amphibole in a blueschist metamorphic environment and is thus the mineral most suitable for dating such rocks.

K-Ar (122 ± 4 m.y.) and whole-rock Rb-Sr (105 m.y.) dates for micaceous blueschists of the Mount Watson area (Haugerud, 1980) are in the range common to the Darrington Phyllite dates and most probably have been partially reset due to proximity of the Tertary Chilliwack composite batholith or to deformation under prehnite-pumpellyite conditions during mid-Cretaceous thrusting. We also list in Table 3 several other previously unpublished K-Ar dates of 141 to 97.2 m.y. for Shuksan and associated rocks, courtesy of R. B. Forbes and D. L. Turner of the University of Alaska. These samples were collected by J. A. Vance of the University of Washington. Sample RF75120 from upper Finney Creek was run as a check on the Late Paleozoic amphibole date of sample PM 9 and gave only 104.5 ± 7 m.y., less than half the previous, much less precise, date given by Geochron Laboratories Inc. in the 1960s. Samples RF75096 and RF75098 are from southeast of Darrington (Vance, 1957). The first is an amphibolite from the Murphy Creek side of Helena Ridge and gives date of 141 ± 6 m.y. that was discussed by Vance and others (1980) in their review of Mesozoic ophiolitic rocks in western Washington. The latter is a blueschist from 1 km north of Rattrap Pass, which gives a typical Shuksan-Darrington date of 97.2 ± 2 m.y.

Whole-rock and Sr isotopic data for Shuksan blue- and greenschists, as seen on Figure 2, are spread over a smaller range in Rb/Sr ratios and have less radiogenic Sr than Darrington Phyllite. Initial $^{87}Sr/^{86}Sr$ ratios in the range 0.703 to 0.704 are indicated; they are quite normal for moderately altered ocean-floor basalts or fresh volcanic arc basalts (Basaltic Volcanism Study Project, 1981). Comparison of Shuksan Sr isotope and Rb/Sr data with the Franciscan compilation (Fig. 3) shows a similar range of Rb/Sr ratios and a similar lower limit of initial $^{87}Sr/^{86}Sr$ ratios (~0.703), but the Franciscan metavolcanic rocks show greater scatter towards a Mesozoic seawater value (~0.708). We could probably observe the same scatter if we analyzed a larger and more variably altered sample suite of Shuksan rocks.

Sr ISOTOPIC COMPARISON WITH OTHER MESOZOIC ROCK SUITES

On Figure 5 (parts a and b) we show Sr isotope data for two other Mesozoic suites for comparison with the Darrington Phyllite. These data have been accumulated in a more haphazard and incomplete fashion than those for the Darrington suite and thus may not equally well represent the rock units. Nevertheless, they present our progress to date in characterizing other mostly sedimentary rock suites south of 49° N latitude. The Nooksack Group of the Mount Baker window (Misch, 1966; see Fig. 1) is a mildly metamorphosed (prehnite-pumpellyite facies, nonschistose to phyllitic), fossiliferous (upper Oxfordian or lower Kimmeridgian through upper Valanginian, according to J. A. Jeletzky's identifications of Misch's collections), volcaniclastic sequence at least 1.5 km thick. The source terrane was predominantly andesitic, and the Nooksack rests on Middle Jurassic andesite and dacite breccias and flows. An arc setting is indicated. The Nooksack samples that were analyzed were supplied by J. N. Sondergaard (1979) who remapped part of the type area. The results (Fig. 5a) scatter about an isochron parallel with the Darrington trend but show less scatter about the isochron, a slightly lower initial ratio, and lower range in Rb/Sr ratios. In all these characteristics they match the Great Valley sequence of California, as shown by a few data on Figure 4. The bulk-earth model date (Cameron and others, 1981) for Nooksack rocks is 124 m.y. Fossils and sedimentary petrography have already assigned these two suites a partially similar age (Late Jurassic and Cretaceous) and comparable tectonic setting (adjacent to a volcanic arc) (Misch, 1966; Sondergaard, 1979; Ingersoll, 1979), so the Sr data indicate nothing new or remarkable.

The Leech River Complex of southern Vancouver Island (Muller, 1977) has been recently restudied by Fairchild (Fairchild and Cowan, 1982), and Rusmore (1982). It consists of clastic sediment interbedded with volcanic rocks and chert that have together undergone sub-greenschist- to amphibolite-grade metamorphism. A Jurassic to Cretaceous depositional age and Eocene metamorphism are indicated by lithologic matching, scanty fossil evidence, and 39 to 41-m.y. K-Ar dates on metamorphic minerals (Muller, 1977). The Leech River sedimentary rocks have been accreted to the southwestern end of a fragment of Wrangellia, which lies outboard of the Northwest Cascades and San Juan imbricate zones. The Sr isotopic analyses of a varied suite of Leech River rocks are plotted in Figure 5b. The scattered results occupy the low Rb/Sr end of the field in which Darrington and Franciscan data plot,

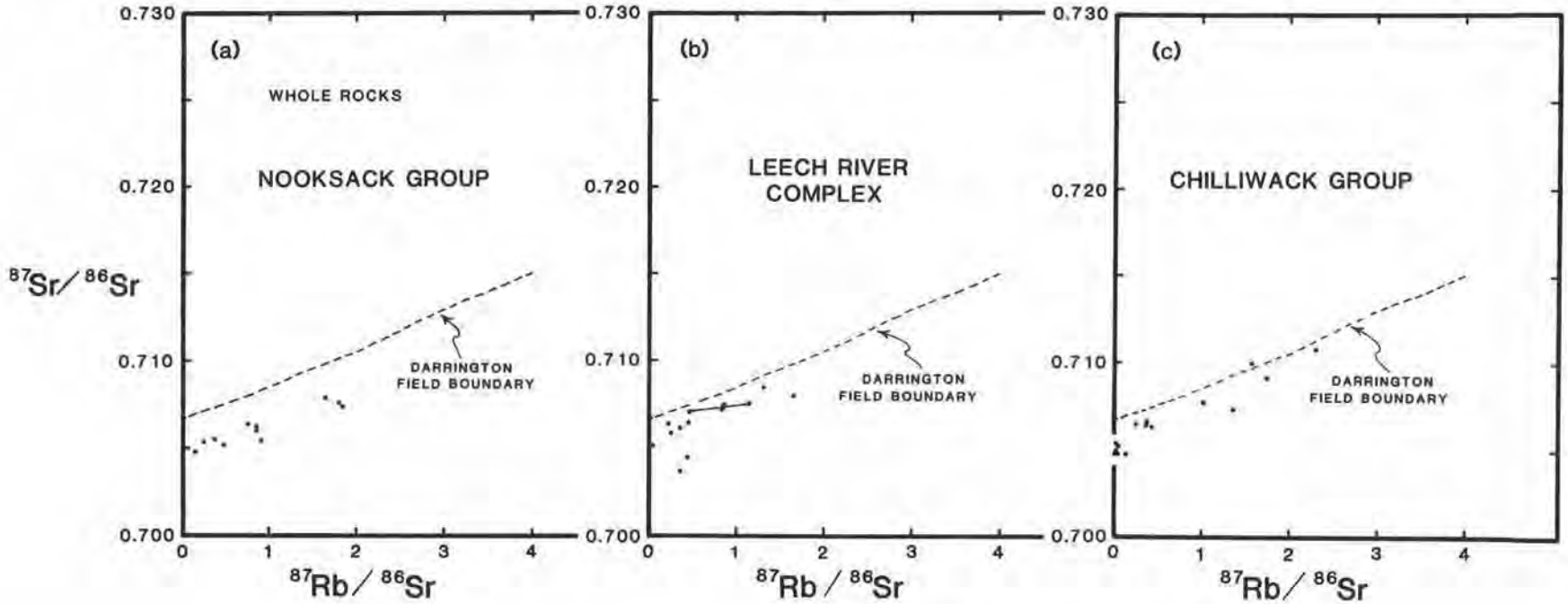

Figure 5.–Rb-Sr isochron plots for whole-rock analyses of (a) Nooksack Group, (b) Leech River Complex, and (c) Chilliwack Group rocks. A portion of the Darrington field boundary is shown by a dashed line in each case. The solid line in the Leech River plot is a local whole-rock isochron for samples from near Port Renfrew.

consistent with the geologically inferred Jurassic-Cretaceous age of deposition. Three samples from one series of road cuts near Port Renfrew (on the southwest coast of Vancouver Island) lie on an isochron with a 42 m.y. date. This agrees exactly with mineral dates for the metamorphism in Eocene time and is thus a further example of local-scale resetting of whole-rock Sr systems during metamorphism.

VEDDER COMPLEX

New mineral dates for small tectonic slices of muscovite-amphibole schist and muscovite schist from the imbricate zone below the Shuksan thrust north of Baker Lake, a unit not previously dated but recognized by Misch (1977, 1980) as containing Vedder-like rocks, are Late Paleozoic, in perfect agreement with dates on other Vedder Complex fragments (Armstrong and others, 1983). The new and previously reported Rb-Sr mineral isochrons are shown together on Figure 6. Comparison of Figure 6 with Figure 4 underlines the contrast of Vedder and Shuksan rocks. Armstrong and others (1983) reported Vedder K-Ar and Rb-Sr dates scattered from 219 m.y. to near-concordant at 277 to 285 m.y. Our new K-Ar dates near Baker Lake are 274 ± 9 and 283 ± 9 m.y. for white mica (samples PL90-8 and PM 101) and 259 ± 9 m.y., for barroisitic amphibole (sample PM 101). The respective Rb-Sr dates for mineral-whole-rock pairs are 283, 273, and 197 m.y., and the two-mineral plus whole-rock isochron date is 280 ± 39 m.y. This demonstrates that Vedder rocks can be recognized a priori on the basis of geologic character and that they have a very reproducible geochronometry.

A somewhat weathered quartz-mica semischist (sample RH78 D61) tectonic block sampled by Haugerud (1980) northwest of Mount Watson, south of Baker Lake, presumably from the imbricate zone, gave a K-Ar date for muscovite of 196 ± 7 m.y. It is probably another fragment of the Vedder Complex but is too poorly exposed and limited in extent and lithologic variability to make that a certainty. The weathered state of the rock precluded a whole-rock Sr analysis.

The Vedder Complex whole rocks plot generally *above* the Darrington field (Fig. 7). Only samples of low Rb/Sr ratio lie in the Darrington Field. This is further convincing evidence that Vedder rocks are older than Shuksan and are distinguishable on whole-rock isotopic as well as geologic grounds. If the Shuksan rocks were merely reset Vedder rocks, then their initial ratios would be much higher than observed. The difference in whole-rock Sr *and* mineral dates precludes any confusion between the two rock groups.

GARRISON SCHIST

The Garrison schist of the San Juan Islands (Vance, 1977) and the Vedder Complex have a partially similar metamorphic character—albite epidote amphibolite being typical of both units (Misch, 1977). The argument for similarity was enhanced by a 242 ± 7 m.y. (recalculated to new decay constants) K-Ar date reported by Vance (1977) for Garrison schist on Orcas Island (Table 3). K-Ar dates for two other Garrison schist samples provided by J. A. Vance and M. T. Brandon are 286 ± 10 m.y. for hornblende and 167 ± 6 m.y. for soda-actinolite (Table 3). These dates both support and confuse the comparison with Vedder rocks (Brandon and others, 1983).

Further comparison comes from whole-rock Sr analyses of three Garrison samples shown on Figure 7. The Garrison schist of the San Juan Islands has a very low Rb concentration and low Rb/Sr ratio compared to Vedder rocks and on the basis of the $^{87}Sr/^{86}Sr$ ratio cannot be distinguished from Shuksan-Darrington samples. If we argue that the one low K-Ar date is reset due to a complicated later Mesozoic history, then we can conclude that Garrison and Vedder rocks are similar in apparent age of metamorphism. Whole-rock Sr does not support the correlation but is, by itself, not grounds for rejection of the possibility.

CHILLIWACK GROUP

The Sr analyses we now have for Chilliwack Group rocks do not provide a geologically consistent story. As can be seen on Figure 5c, the analyses fall largely in the Darrington field. The fossil evidence for a Devonian to Permian age (Danner, 1957, 1966; Misch, 1966) cannot be dismissed. Metamorphic grade is very low (prehnite-pumpellyite) and identical with that of the Nooksack Group and Wells Creek Volcanics (Misch, 1966, 1971, 1977).

The results are partly expectable. For any Phanerozoic age, low Rb/Sr-ratio metavolcanic samples are indistinguishable — giving 0.704 to 0.705 present-day *and* initial $^{87}Sr/^{86}Sr$ ratios, as might be expected for rocks of a slightly altered ocean floor or somewhat less altered volcanic arc. The puzzle is the ^{87}Sr deficiency in a few samples of clastic metasediment. There are several possible explanations, all ad hoc. The first relates to the caution already expressed that these samples were not as deliberately collected to represent large volumes of typical rocks and thus may not be representative. Second, the effects of metamorphism may be more drastic than expected, or hoped, so that all isotopic evidence of a Paleozoic age has been erased. Third, we may have in our Chilliwack collec-

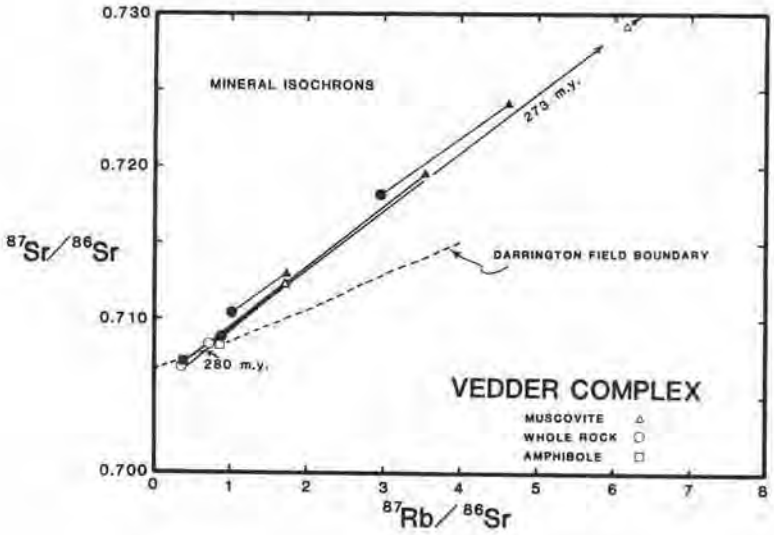

Figure 6.–Rb-Sr isochron plot for whole rocks and separated minerals of the Vedder Complex. Published data (Armstrong and others, 1983) are shown with solid symbols, new data (Table 1) with open symbols. A portion of the Darrington field boundary is shown by a dashed line.

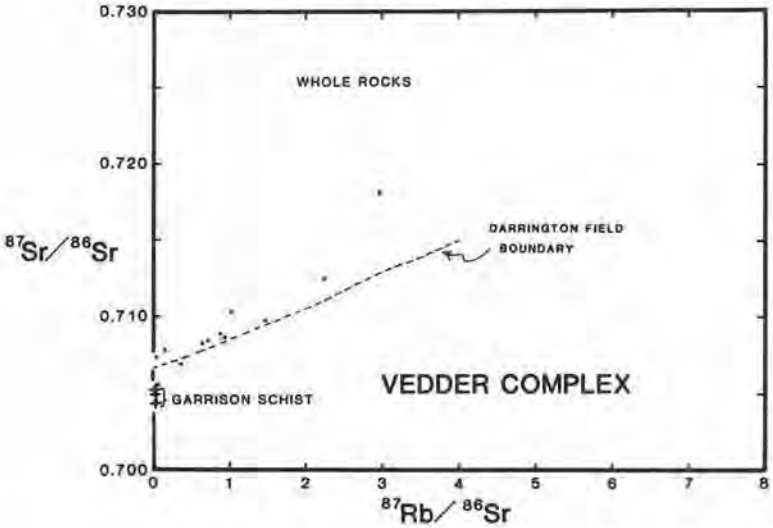

Figure 7.–Rb-Sr isochron plot for whole rocks of the Vedder Complex and Garrison schist. A portion of the Darrington field boundary is shown by a dashed line. The generally more radiogenic character of Vedder Complex rocks is evident.

tion samples of Mesozoic sedimentary slices within the Church Mountain plate. We leave the answer to the future. Our analyses have only found a problem to be investigated more systematically.

CONCLUSION

New geochronometry for Shuksan-Darrington and Vedder Complex rocks has confirmed the distinction between the units as recognized on geologic grounds (Misch, 1977, 1980) and on the basis of geochronometry (Brown and others, 1982; Armstrong and others, 1983). Published and new age determinations for metamorphism of Shuksan, Darrington, Vedder, and Garrison rocks are summarized in histograms on Figure 8. The older dates, mostly for relatively coarse-grained mineral phases, probably best indicate the time of temperature decrease after the first metamorphic culmination—Late Jurassic for Shuksan-Darrington rocks, and Late Carboniferous to Early Permian for Vedder-Garrison rocks. The isotopic dates for coarse-grained Darrington rocks give whole-rock K-Ar and many whole-rock Rb-Sr isochron dates of 100 to 126 m.y., some even less than 100 m.y. This may be the result of slow cooling following the period of maximum temperature, and there is also the likelihood of resetting associated with thrusting and prehnite-pumpellyite-facies metamorphism in mid-Cretaceous time, as well as local Cenozoic resetting. Various times and degrees of metamorphism have been demonstrated in California where older coarse-grained blueschists are mixed by sedimentation and tectonics into younger finer grained blueschists (Suppe and Armstrong, 1972). Isotopic dates for coarse-grained

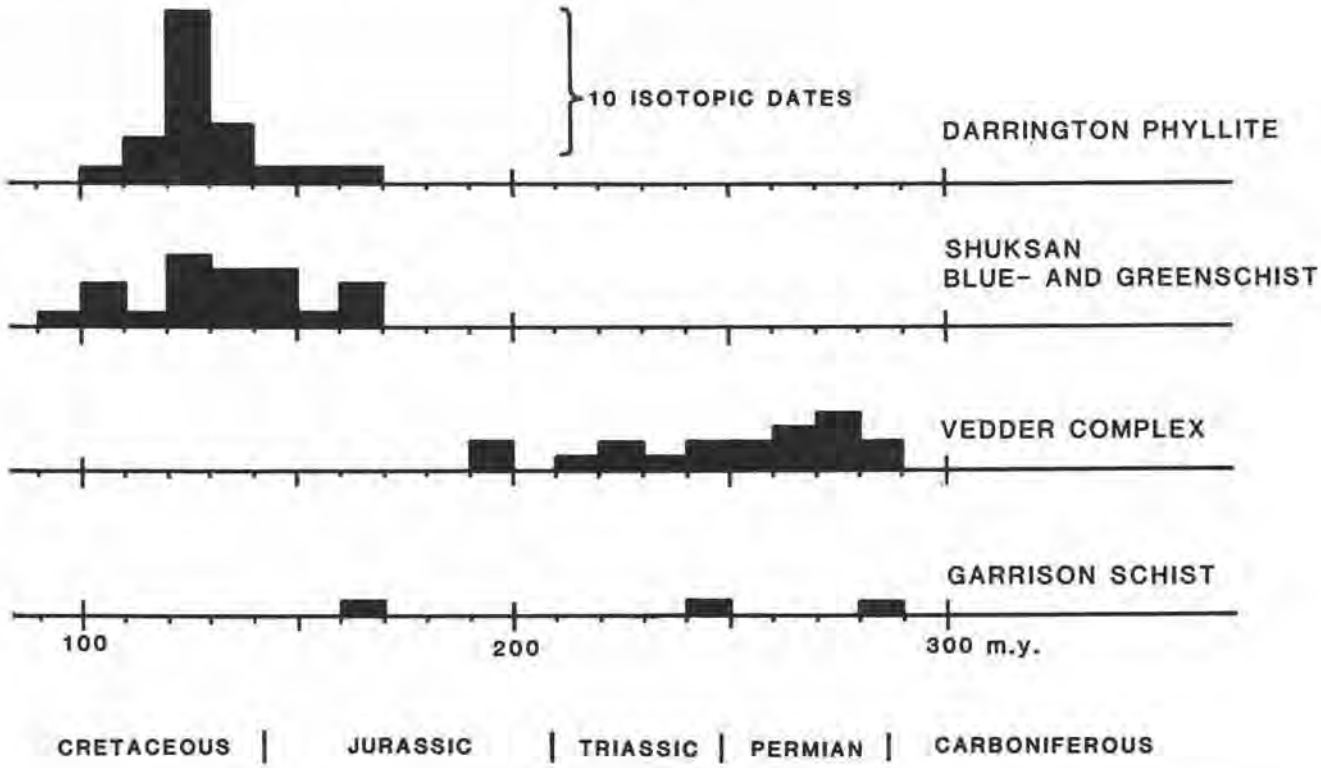

Figure 8.–Histograms of K-Ar and Rb-Sr dates for Darrington Phyllite, Shuksan blue- and greenschist, Vedder Complex, and Garrison schist. Late Jurassic and Early Permian-Late Carboniferous minimum ages, respectively, for two metamorphic events are clearly distinguished.

Vedder Complex rocks range from 238 to 285 m.y. Some rocks, especially finer grained ones, give dates spread downwards towards 200 m.y., for much the same reasons as the younger Shuksan dates.

Metamorphic age, Sr isotopic signature, and, in a general way, metamorphic type together broadly match Shuksan-Darrington with the main body of Franciscan rocks in California and Oregon (eastern and central belts). However, the Franciscan assemblage is much more varied lithologically, and it includes younger material in its western parts—strata as young as Tertiary have been accreted—whereas in Washington younger accretion has stepped west of the Cascade Mountains and San Juan Islands into the Olympic Mountains (Tabor and Cady, 1978). The isotopic similarity does not establish tectonic or stratigraphic continuity between western California-Oregon and northwestern Washington. In that context it is appropriate to note that the Leech River Complex, as discussed, and the Pacific Rim Complex of Vancouver Island (Muller, 1977) have some features in common with the Franciscan and Darrington rocks (although the Leech River has been pervasively overprinted by early Cenozoic metamorphism). Both complexes lie outboard of Wrangellia and are separated from the Northwest Cascades by several thrust plates, including Wrangellia itself.

Nooksack Group rocks, structurally beneath the Church Mountain and Shuksan plates, are similar in age, Sr isotopic signature, depositional environment (forearc or backarc basin) to early Great Valley rocks of California.

How the present Cascade geometry was achieved remains a subject on which there is no general agreement. In northwestern Washington and southwestern British Columbia there are many more Mesozoic stratigraphic packages than we have discussed here that would have to be taken into account in any overall synthesis of paleogeography and tectonics.

The first K-Ar dates reported for the Shuksan Suite by Misch (1966) led to comparison with Triassic blueschists in California, Oregon, and British Columbia (Hotz and others, 1977), but the one reliable date was actually from a tectonic slice of Vedder Complex blueschist at the sole of the Shuksan plate. Thus the particular comparison was misleading, but there may yet be a tie between the North Cascades and central Oregon. Avé Lallemant and others (1980) report 255

to 262 m.y. K-Ar dates for amphibole in metamorphic and plutonic (Canyon Mountain) rocks and Walker (1981,1983) has reported 225 to 279 m.y. U-Pb dates for zircon from the same ophiolite and arc-plutonic rocks included in Triassic and Jurassic melange in Oregon. The similarity of the central Oregon and Vedder-Garrison rocks in isotopic age and setting within a Mesozoic tectonic melange invites further comparisons.

For Chilliwack rocks, the oldest group studied, we have provided a few Sr isotopic analyses that only add confusion and hint at unresolved problems for future work.

ACKNOWLEDGMENTS

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POSSIBLE TECTONOSTRATIGRAPHIC TERRANES IN THE NORTH CASCADES CRYSTALLINE CORE, WASHINGTON

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ABSTRACT

The southeast part of the North Cascades comprises five possible tectonostratigraphic terranes: (1) the Nason terrane, (2) the Ingalls Tectonic Complex, (3) the Swakane terrane, (4) the Mad River terrane, and (5) the Chelan Mountains terrane.

The Nason terrane is composed predominantly of the Chiwaukum Schist, which is mostly pelitic mica schist in its upper part but grades into mica schist containing amphibolite and rare marble in its lower part. The Chiwaukum has been regionally metamorphosed in two episodes: an early Barrovian event and a later Buchan event accompanied by intrusion of the Late Cretaceous syntectonic to late-tectonic Mount Stuart batholith and other plutons. Within this terrane, an antiformal core and other uplifted regions of the Chiwaukum have been transformed into banded gneiss by metasomatic and intrusive processes. On the west, separated from the Nason terrane by the Evergreen fault, is the Tonga Formation, a correlative of the Easton Schist, but not as clearly correlated with the Chiwaukum as earlier proposed. An early, regional, Alpine metamorphism of the Tonga created blueschist and phyllite; the Chiwaukum and Tonga may have been widely separated at that time. The later, Buchan episode of metamorphism in the Chiwaukum Schist was shared with the Tonga Formation.

The Ingalls Tectonic Complex is a dismembered ophiolite that was thrust over the Nason terrane prior to intrusion of the Late Cretaceous Mount Stuart batholith and contemporaneous regional metamorphism. Radiolarians in chert and U-Pb ages of zircons from gabbro in the complex indicate a Late Jurassic age for these components. A possible serpentine-slide origin for a metaperidotite layer of Ingalls ophiolite in the Chiwaukum Schist suggests that during the deposition of the protolith shale the Ingalls was adjacent. On the basis of this argument, the age of the Chiwaukum protolith is probably Late Jurassic to Late Cretaceous.

The Swakane terrane is composed of the Swakane Biotite Gneiss, a uniform metaclastite(?) or metadacite(?) containing rounded zircons more than 1,650 million years old. A preliminary Rb-Sr whole-rock isochron yields an age of about 690 million years, tentatively interpreted as a probable minimum stratigraphic age.

Tectonically overlying the Swakane terrane is the Mad River terrane, a unit of metachert, metabasite, orthogneiss, marble, and ultramafite. Complex U-Pb systematics of zircons from the orthogneiss suggest a mixture of Paleozoic or older zircons and Late Cretaceous or Tertiary metamorphic zircons. Locally a zone of tectonic imbrication separates the Mad River terrane from the Nason terrane, but much of the contact is intruded by a "stitching" pluton, the Late Cretaceous Ten Peak pluton. Recrystallized mylonite separates the Mad River terrane from the Chelan Mountains terrane.

The Chelan Mountains terrane is composed of two northwest-trending belts. A southwest belt of metaplutonic rocks includes the Dumbell-Marblemount plutons of Triassic age and the Entiat pluton of Late Cretaceous age. The northeast belt is composed of metasedimentary and metavolcanic rocks grading from greenschist facies on the northwest (Cascade River Schist of Misch), through amphibolite facies (the younger gneissic rocks of the Holden area), to ultrametamorphic migmatite and anatectic(?) tonalite plutons (including the Entiat pluton) of the Chelan Complex on the southeast. Lithologic similarity between the metasedimentary and meta-igneous rocks of the Chelan Mountains and Mad River terranes suggests a correlation which would make but one terrane of the two. The two terranes as here described, however, are separated everywhere by faults.

Most zircon U-Pb ages from plutons of the Chelan Complex reflect Late Cretaceous magmatic crystallization, although some $^{207}Pb/^{206}Pb$ ages, mostly from migmatite, indicate probable Triassic metaplutonic and Permian(?) or Triassic (or younger) metasedimentary and metavolcanic protoliths. Continued metamorphism partially recrystallized the anatectic plutons of the Chelan Complex.

All the described terranes were recrystallized by Late Cretaceous and earliest Tertiary regional metamorphism (~90-60 million years ago) and intruded by synkinematic to late-kinematic tonalite to granodiorite plutons, some of which stitch terranes together. Possible post-Late Cretaceous rotation of paleomagnetic poles with low inclination in the Mount Stuart batholith suggests that the amalgamated North Cascades terranes moved northward after the Late Cretaceous to arrive at their present site against the North American plate.

INTRODUCTION

Before about 1970, the North Cascades range was considered to be for the most part an in-place eugeosynclinal accumulation of metasedimentary and metavolcanic rocks. (See, for example, McKee, 1972, p. 82-104.) Using fossil evidence from mostly unmetamorphosed rocks west of the Straight Creek fault (Fig. 1), Danner (1970) first suggested the presence of exotic components in the general area of the North Cascades. Proceeding from paleomagnetic evidence, Beck and Noson (1971) first postulated that parts of the crystalline core had been transported a large distance northward; and Davis (1977) and Davis and others (1978) proposed that the crystalline core had been transposed northward by large-scale faulting during the Mesozoic from a position south of, and continuous with, the Okanogan crystalline belt. Following these leads, Coney and others (1980) included the North Cascades in their composite suspect terrane, Cascadia.

Our work suggests that there are some fundamental differences in age and geologic history among tracts of metamorphic rocks in the North Cascades crystalline core. Although we cannot identify whence the tracts came, or how or when they arrived at their present position, these suspect tracts may indeed be tectonostratigraphic terranes; they now constitute a patchwork that forms at least part of the crystalline core of the North Cascades.

RECOGNITION OF TERRANES IN HIGHLY METAMORPHOSED ROCKS

We use the term "tectonostratigraphic terrane" in much the same way as Beck and others (1980, p. 454): "A tectonic (or stratigraphic) terrane is defined as a fault-bounded geologic entity characterized by a distinctive stratigraphic sequence and a structural history differing markedly from those of adjoining neighbors." Features that meet the criteria in this definition may be diffcult to identify in metamorphic rocks. The terranes that we describe here clearly have distinctive lithologies, and clues to their protolith ages indicate that they may have formed in different geologic eras. They share, however, a Late Cretaceous and earliest Tertiary metamorphism that obscures their premetamorphic structural history. Regional metamorphism and stitching plutons mask the faults bounding the terranes.

At best, the terranes outlined here should be considered tentative, at worst, speculative. In the course of our discussion, we interpret published descriptions of rocks that were not mapped with terrane concepts in mind. We hope that our speculations will lead others to consider the geology of the North Cascades in a nontraditional manner.

NATURE OF THE TERRANES

At least five candidate tectonostratigraphic terranes make up the southern part of the North Cascades: (1) the Nason terrane, composed mostly of Late Jurassic(?) metapelite; (2) the Ingalls Tectonic Complex, a dismembered ophiolite of Late Jurassic protolithic age; (3) the Swakane terrane, composed of the Swakane Biotite Gneiss, of uncertain origin but of apparent Precambrian protolith age; (4) the Mad River terrane, composed of mostly marine metasedimentary and metavolcanic rocks, of unknown but possibly Paleozoic or older protolithic age; and (5) the Chelan Mountains terrane, composed of metamorphosed late Paleozoic(?) or Triassic (or younger) sedimentary and volcanic rocks and metamorphosed Triassic plutons. Figure 1 shows the locations of the terranes and their probable extensions to the north as we now recognize them; Table 1 summarizes their major characteristics and emphasizes differences.

Nason Terrane

The Chiwaukum Schist and banded gneiss derived from the Chiwaukum by metamorphic and igneous processes are the major units of the Nason terrane (Fig. 2). We have named the terrane for Nason Creek (near U.S. Highway 2), where rocks typifying the terrane are exposed (Tabor and others, in press). The Nason terrane is bounded by the Leavenworth fault on the east and by other faults, such as the Evergreen

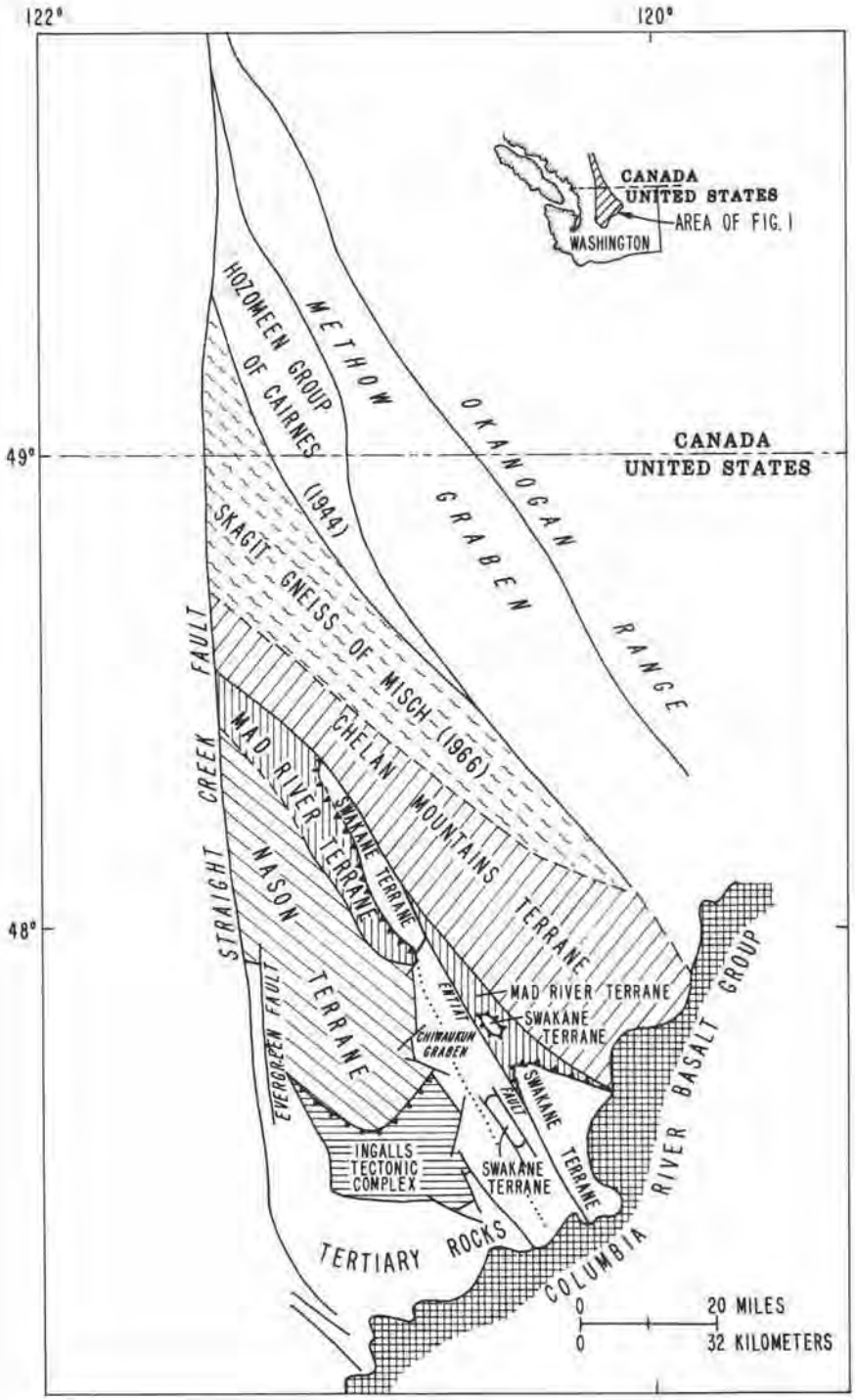

Figure 1.–Locations of possible major tectonostratigraphic terranes in the North Cascades, Washington. Patterned units are discussed in text. Contacts between terranes are faults and are dashed where inferred, dotted where concealed. Barbed contacts are thrust faults.

Table 1.–Summary of major characteristics of tectonistratigraphic terranes in the North Cascades

Characteristics	Nason terrane	Ingalls Tectonic Complex	Swakane terrane	Mad River terrane	Chelan Mountains terrane
Protolith rock type	Al-rich shale overlying shale and mafic volcanic rocks and minor limestone; rare ultramafite; fairly homogeneous	Ultramafite, diabase, gabbro, basalt, graywacke, shale, and chert; heterogeneous	Sandstone or shale; possibly dacitic pyroclastic and rare mafic volcanic rocks and marble; no ultramafite; highly homogeneous	Chert, mafic volcanic rocks, limestone, and minor shale; gabbroic intrusions; common ultramafite; heterogeneous	Cherty graywacke, arkose, shale, mafic volcanic rocks, and marble; minor ultramafite; gabbroic to tonalitic igneous plutons; heterogeneous
Structural and metamorphic style	Multiple deformations; isoclinal and broad regional folding of foliation, and compositional layering on WNW axes; metamorphic segregation and igneous injection where uplifted along antiformal fold and on the northwest	Melange with penetrative foliation on the east; static thermal recystallization on the west	Very rare folds; mostly planar foliation; homoclinal or broadly arched along NW axis; gneissic segregation and migmatite	Complexly folded on all scales; northwest alignment of structure; mostly without development of gneissic segregation	Some isoclinal folding and larger folds along NNW axes; migmatization on southeast end
Small bodies of metamorphosed silicic intrusive rocks	Not abundant	Uncommon	None	Abundant	Common in metasedimentary rocks
Contact relations	In thrust contact with the Ingalls Tectonic Complex; probably imbricated with the Mad River terrane in a wide tectonic zone	Thrust over the Nason terrane; contact with other terranes covered by Tertiary rocks	In thrust contact with the Mad River terrane	Probably imbricated with the Nason terrane; probably thrust over the Swakane terrane; probably faulted against the Chelan Mountains terrane	Probably faulted against the Mad River terrane
Age(s) of protolitic components	Probably Late Jurassic to Late Cretaceous	Late Jurassic	Precambrian	Pre-Late Cretaceous	Permian(?) or Triassic or younger metasedimentary and metavolcanic rocks; Triassic plutons

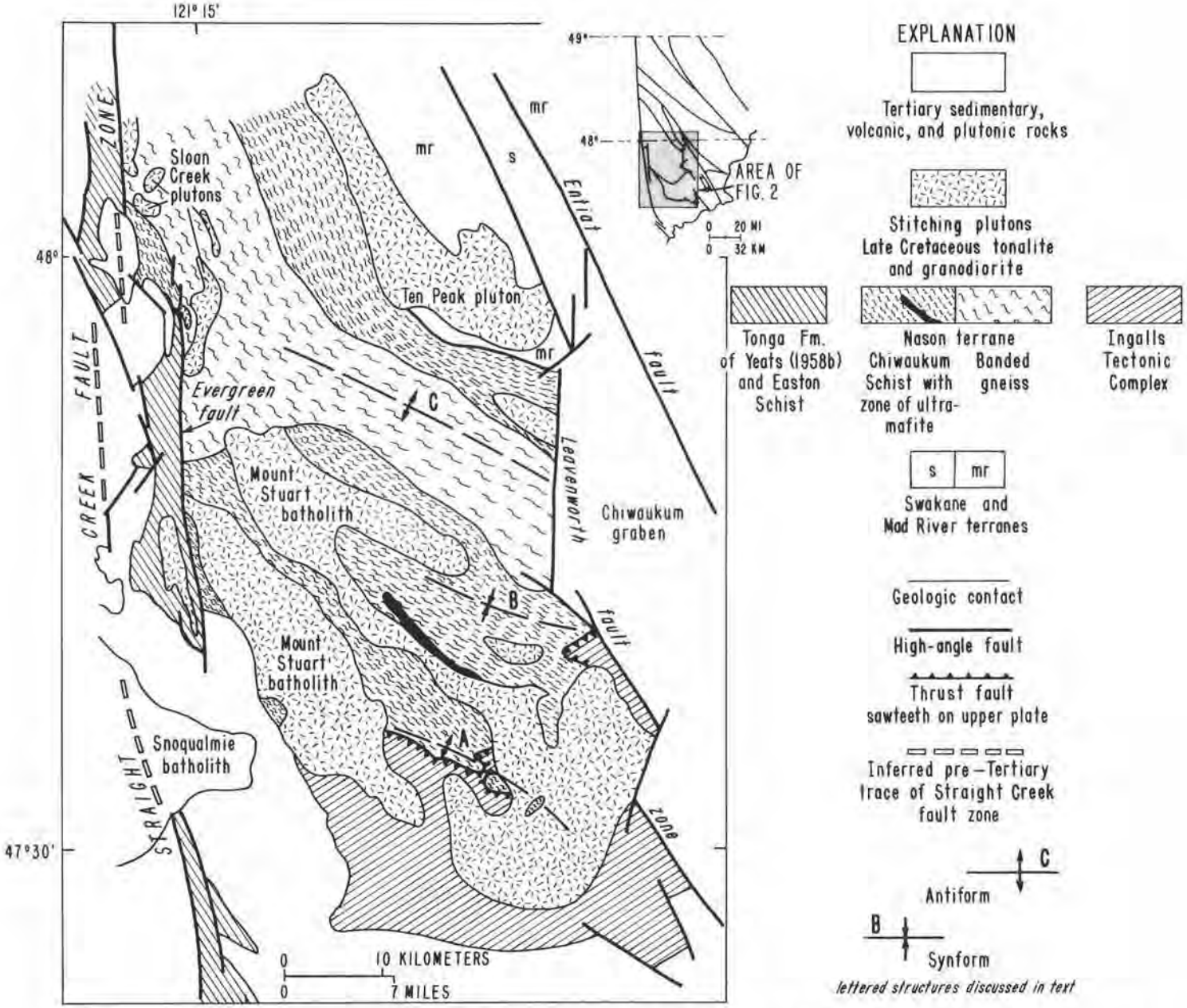

Figure 2.–Simplified geologic map of the Nason terrane and the Ingalls Tectonic Complex, based on reports by Crowder and others (1966), Tabor and others (1980, 1982a, 1982b), and Frizzell and others (1984).

fault in the Straight Creek fault zone, on the west. On the northeast, the Nason terrane appears to be faulted against the Mad River terrane, and on the south it is overthrust by the Ingalls Tectonic Complex.

The Chiwaukum Schist is predominantly graphitic garnet-biotite-quartz schist, commonly containing one or more of the aluminum silicate minerals staurolite, kyanite, and sillimanite. The lower part of the section contains abundant hornblende-biotite schist, hornblende gneiss, cummingtonite schist, amphibolite, rare calc-silicate schist, and marble. In spite of the high degree of metamorphism, the Chiwaukum locally retains relict sedimentary features such as graded bedding. The metasedimentary and metavolcanic rocks grade into banded gneiss consisting of interlayered biotite gneiss, gneissic tonalite, and relics of mica schist and amphibolite. The metapelite and banded gneiss of the Nason terrane have been described in detail. (See Page, 1939, p. 16-32; Bryant, 1955, p. 76-109; Oles, 1956, p. 41-111; Yeats, 1958a, p. 19-28; Ford, 1959, p. 27-111; Rosenberg, 1961, p. 21-34; Van Diver, 1964, p. 15-38; Crowder and others, 1966; Plummer, 1969, p. 54-124; 1980; Heath, 1971, p. 12-26; Getsinger, 1978, p. 13-67; Kaneda, 1980; McDougall, 1980, p. 11-17).

Ultramafic rocks, mostly serpentinized metaperidotite and talc-anthophyllite or tremolite rocks, are dispersed throughout the Chiwaukum Schist and banded gneiss but are mostly concentrated near the thrust contact with the Ingalls Tectonic Complex, near the probable faulted contact with the Mad River terrane, and near other major faults. Some of the ultramafite bodies may be pieces of the Ingalls Tectonic Complex, as discussed below.

The Chiwaukum Schist has undergone a complex metamorphic history. According to Plummer (1980), in an early high-pressure, high-temperature, Barrovian regional metamorphism, rocks locally attained the sillimanite zone of the amphibolite facies. A second, lower pressure, Buchan episode produced more sillimanite, as well as cordierite and andalusite. The second phase culminated about 90 m.y. ago with the intrusion of the Late Cretaceous Mount Stuart batholith, Sloan Creek plutons, and Ten Peak pluton (Fig. 2; see also Kaneda, 1980; Tabor and others, 1980, 1982a).

The overall structure in the Nason terrane reflects a complex antiform on the south (Fig. 2, antiform A), pushed up(?) and intruded by the Mount Stuart batholith; a synform in the Chiwaukum Mountains area (Fig. 2, synform B); and an antiform in the Little Wenatchee River area (Fig. 2, antiform C). The major antiform in the Little Wenatchee River area brings up banded gneiss and other light-colored gneissic rocks characterized by metasomatic and intrusive material that have transformed the Chiwaukum Schist into light-colored tonalite gneiss. The presence of kyanite in many of the uplifted rocks, in contrast to staurolite on the limbs of the antiform, indicates that the upward arching was late, in the second episode of metamorphism. Exposure of rocks formed at greater depth in the more northern antiform (C), in contrast to those in the southern antiform (A), is not surprising, inasmuch as the entire Cascade Range in this area appears to be progressively uplifted toward the north. Local recrystallized mylonite zones in gneiss of the Little Wenatchee antiform suggest pronounced shear parallel to foliation. The antiform might be viewed as an incipient gneiss dome. (See, for example, Fox and Rinehart, 1971.) The major structural elements in the Nason terrane have a much more westerly trend than do structures in the terranes to the east.

We do not know the age of the protolith of the Chiwaukum Schist. According to arguments given below relating the Chiwaukum protolithic shale to the Ingalls Tectonic Complex, the Chiwaukum protolith may be Late Jurassic to Late Cretaceous in age. However, Lowes (1972, p. 167) and Misch (1977) have suggested that the Chiwaukum Schist is correlative with the Settler Schist in British Columbia and offset about 200 km along the Straight Creek fault. The lithology and metamorphic history of the Settler Schist and the Late Cretaceous Spuzzum plutons make this correlation quite convincing. The Settler Schist yields a Rb-Sr isochron age of 214 m.y. (Triassic), which has been interpreted as a sedimentary age (Bartholomew, 1976, p. 100), but could also reflect the provenance age of the original sediment. The probable offset of the Nason terrane has important implications, as discussed below.

Important to understanding the Nason terrane is the relation between the Chiwaukum Schist and the Tonga Formation of Yeats (1958b), which crops out just west of the terrane, separated from it by the high-angle Evergreen fault (Fig. 2). In earlier reports, Tabor and others (1982a,d) suggested that the Tonga Formation was a lower grade equivalent of the Chiwaukum Schist, a correlation that is only partly correct.

The character of the phyllite and rare outcrops of blue-amphibole-bearing greenschist in the Tonga Formation led Yeats (1958a, p. 41; Yeats and others, 1977, p. 267; Misch, 1966, p. 111) to correlate the Tonga with the Easton Schist south of the Snoqualmie batholith (Fig. 2) and the Shuksan Metamorphic Suite of Misch (1966, p. 103) to the north. The Shuksan Metamorphic Suite (Easton Schist) has most recently

been considered Late Jurassic in protolithic age, probably metamorphosed during the Early Cretaceous (Brown and others, 1982, p. 1095).

Anearly episode of high-pressure, low-temperature Alpine-type metamorphism in the Tonga Formation created phyllite and blue-amphibole-bearing greenschist. A second episode of higher temperature, Buchan-type metamorphism (Misch, 1971; Yeats and others, 1977, p. 267) upgraded some of the metapelite and meta-arenite to staurolite-garnet-mica schist and hornblende-mica schist, respectively, while the greenschist became amphibolite; these higher grade rocks closely resemble much of the Chiwaukum Schist. Furthermore, a north-to-south increase in grade in the Tonga (greenschist facies through staurolite zone of the amphibolite facies; Yeats, 1958a) parallels a south-to-north increase in grade (garnet zone through sillimanite zone) in the Chiwaukum Schist (Tabor and others, 1982d, in press). These two units clearly shared the latest episode of metamorphism, although under different conditions and in different crustal positions.

The early high-pressure, low-temperature metamorphism in the Tonga contrasts with the early episode of Barrovian metamorphism in the Chiwaukum Schist. Although evidence of an early Alpine metamorphism in the Chiwaukum is absent, Kaneda (1980, p. 61) documented an early folding in the Chiwaukum that could have been in a lower grade facies, and low-rank greenschist-facies rocks in the Chiwaukum are preserved adjacent to the Mad River terrane on the east. (See Tabor and others, 1980.) Nevertheless, without more definitive evidence that the early metamorphism of these two units took place under the same conditions, we assume that the two units must have been separated during the early metamorphism, and thus that their original stratigraphic continuity is less certain. They probably were separate terranes but were joined by middle or Late Cretaceous time.

Ingalls Tectonic Complex

The Ingalls Tectonic Complex crops out in a wide belt wrapping around the south end of the Mount Stuart batholith (Fig. 2). Although we here consider the complex to correspond to the Mount Ingalls terrane of Silberling and others (1984), we continue to use its established name. The most abundant rocks in the complex are serpentinite, serpentinized peridotite, and metaperidotite, but the complex includes as well flysch-type sandstone and argillite, radiolarian chert, pillow basalt, diabase, and gabbro. All the rocks are stratigraphically associated and tectonically intermixed; the assemblage is an ophiolite complex (Hopson and Mattinson, 1973; Miller, 1977, 1980b; Miller and Frost, 1977, p. 287). Southwick (1974) proposed that the unit consists of ophiolite and island-arc material juxtaposed during subduction. Miller (1980b) and Cowan and Miller (1980) speculated that the complex formed in a leaky-transform-fault zone.

The metamorphic grade of the Ingalls Tectonic Complex ranges from prehnite-pumpellyite facies and greenschist facies in rocks at the southwest corner of the complex to pyroxene-hornfels facies in rocks adjacent to the Mount Stuart batholith. The mostly static thermal metamorphism caused by the intrusion has clearly overprinted an older schistose fabric. However, rocks along the west side of the complex are less schistose and retain more original structures than do rocks to the east near the bounding thrust and along the east margin of the Mount Stuart batholith.

The age of the gabbro in the Ingalls Tectonic Complex is Late Jurassic according to a U-Pb date on zircon (Southwick, 1974, p. 391). Chert in the complex contains radiolarians restricted to the Late Jurassic (E. A. Pessagno, Jr., Univ. of Texas, personal commun., 1977; Tabor and others, 1982b). Considering the contemporaneity of two such disparate members of the complex, we assume a Late Jurassic age for most of the other components, although the complex could contain tectonic slices of rocks of other ages. The age of tectonic mixing of the complex is Late Jurassic or Early Cretaceous.

Miller (1980a, 1980b, p. 390-404) suggested that the Ingalls Tectonic Complex and the Chiwaukum Schist are in thrust contact and were juxtaposed after the early regional metamorphism of the Chiwaukum, but before the intrusion of the Mount Stuart batholith and its associated episode of regional metamorphism. He thought that the Ingalls had been thrust (obducted) onto a continental margin of Chiwaukum Schist.

Whetten and others (1980) included the Ingalls Tectonic Complex in their Haystack thrust plate, a regional thrust plate exposed in the San Juan Islands and in other, isolated places in northwestern Washington.

Well within the Chiwaukum Schist of the Nason terrane (Fig. 2; Tabor and others, 1980), a narrow zone of ultramafic bodies contains a remarkably thin and continuous layer (150-200 m x 5 km) of metaperidotite that parallels compositional layering in the Chiwaukum and is associated with fine-grained sugary amphibolite similar to amphibolite in the Ingalls. Although imbrication of the Ingalls and Chiwaukum might explain the distribution of most ultramafite in the Chiwaukum, an alternative hypothesis is that the thin slab formed as a serpentine slide onto the protolithic sediments of the Chiwaukum. The main slide (and by analogy, smaller ones) may have descended

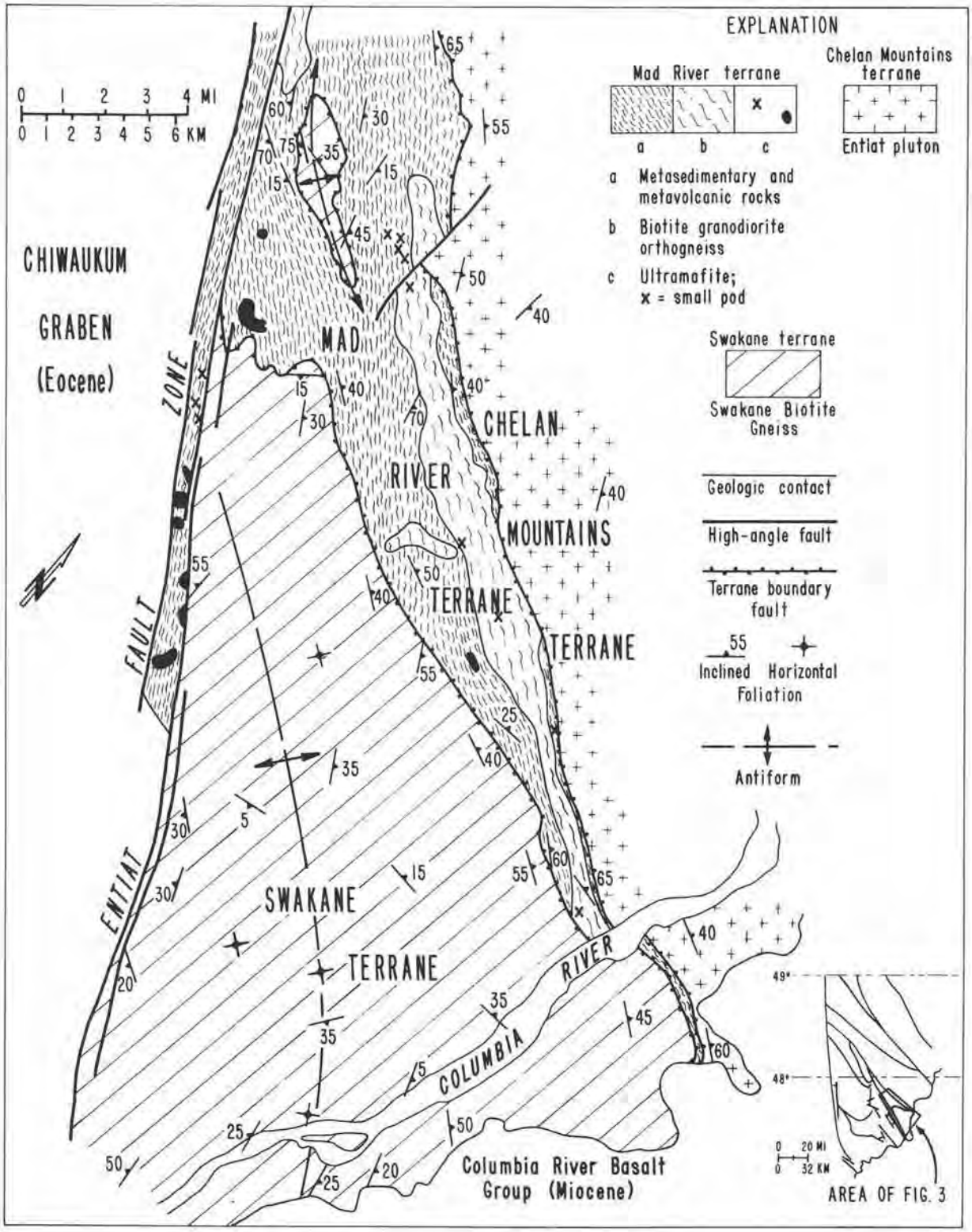

Figure 3.–Simplified geologic map showing the Mad River terrane overlying the Swakane terrane and adjoining the Chelan Mountains terrane. From Tabor and others (1980, in press).

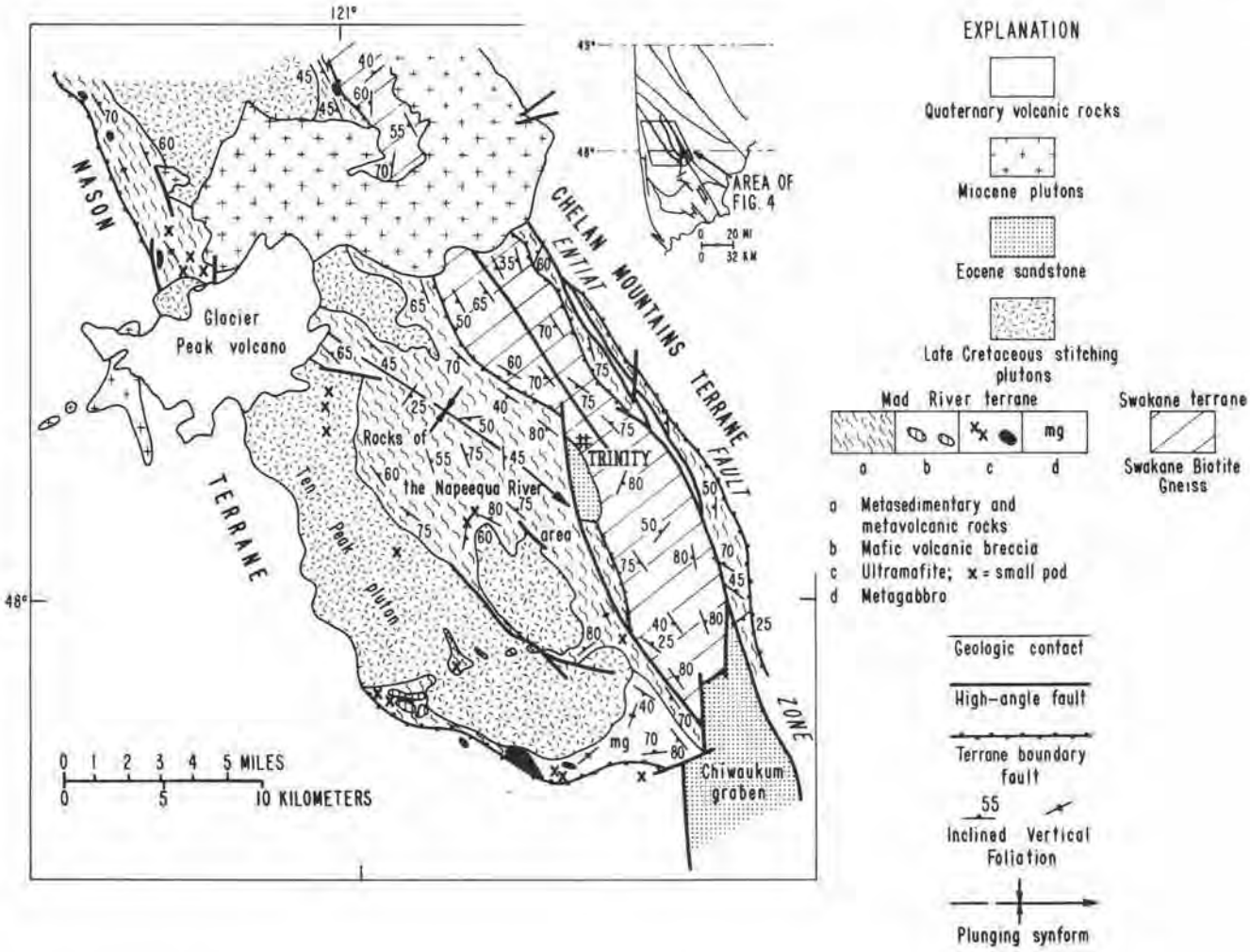

Figure 4.–Simplified geologic map of the central part of the Mad River terrane, showing contact with the Nason terrane. From Cater and Crowder (1967), Crowder and others (1966), and Tabor and others (1982a).

from an adjacent uplifted ophiolite mass of the Ingalls before metamorphism of the two units. The early metamorphism affecting both units may have been gradational, and the gradient may have been cut out by thrusting of the Ingalls over the Chiwaukum before the Late Cretaceous episode of metamorphism and intrusion. If this scenario is correct, the protolithic age of the Chiwaukum Schist is Late Jurassic to Late Cretaceous.

Swakane Terrane

In earlier reports, Tabor and others (1980, 1982c) included the Swakane Biotite Gneiss and the overlying and closely associated heterogeneous schist-and-gneiss unit in the tentatively named Roaring Creek terrane. Because the heterogeneous schist-and-gneiss unit overlies the Swakane in probable thrust contact, we now include that unit in the Mad River terrane (Figs. 1, 3, 4) and retain the Swakane Biotite Gneiss as the sole component of the Swakane terrane.

The Swakane Biotite Gneiss is a remarkably uniform granofelsic gneiss containing rare thin layers of hornblende schist, schistose amphibolite, and marble. Some amphibolite layers have been traced for many kilometers (Cater and Crowder, 1967).

At the south end of the Swakane terrane, the uniform foliation and subtle compositional layering in the Swakane Biotite Gneiss are folded into a broad antiform (Fig. 3). Farther to the northwest (Fig. 4), foliation in the Swakane is variable and steep; isoclinal folds are uncommon (Crowder, 1959, pl. 1, p. 833; Cater and Crowder, 1967). A horst of the Swakane terrane in the Tertiary Chiwaukum graben (Fig. 1) indicates that the contact with the terranes to the west lies buried in the graben and is not the conspicuous

Entiat fault cutting the Swakane and Mad River terranes.

Waters (1932, p. 616) and later workers (Chappell, 1936, p. 48-49; Page, 1939, p. 14-15; Crowder, 1959, p. 834) considered the gneiss to have been derived from clastic sedimentary rocks with rare interbedded basalt and/or mafic dikes and limestone. C. A. Hopson (quoted in Mattinson, 1972, p. 3773) suggested that the Swakane protolith was a silicic volcanic rock, a derivation championed by Cater (1982, p. 5-6). Much of their argument rests on the homogeneity of the Swakane. In other metamorphosed clastic sequences, the contrast between shale and sandstone beds is generally enhanced by metamorphism. The lack of pronounced compositional layering in the Swakane not only contrasts with other terranes in the North Cascades but also indicates a homogeneous protolith, such as a thick accumulation of dacite breccia or tuff (Cater, 1982, p. 6). The uncertain origin of the Swakane Biotite Gneiss makes interpretation of the age of its protolith difficult. Proceeding from a concordia plot of highly discordant U-Pb zircon ages, Mattinson (1972, p. 3773) favored the hypothesis that the zircon in the Swakane originally crystallized at least 1,600 m.y. ago and was metamorphosed about 415 m.y. ago and again 90 to 60 m.y. ago. U-Th-Pb analyses of zircon from garnet amphibolite in the Swakane (Tabor and others, in press) support Mattinson's assessment of its age. If the protoliths of the biotite gneiss or amphibolite were volcanic, the Middle Proterozoic or earlier age should represent its time of formation. However, apparent rounding of many zircons in both rocks suggests a detrital or possible xenolithic origin. In either case, the zircons provide a maximum stratigraphic age. Rb-Sr analyses of numerous whole-rock samples of the Swakane Biotite Gneiss yield a tentative isochron age of about 690 m.y., which R. J. Fleck and A. B. Ford (USGS, personal commun., 1985) interpret to be a probable stratigraphic age based on the Rb-Sr systematics and the arguments given above for a dacitic protolith. Even if the Rb-Sr age reflects resetting during an early metamorphism, it represents a minimum for the protolith and thus requires a Precambrian protolith age for the Swakane.

U-Pb ages of zircons from pegmatite gneiss in the Swakane Biotite Gneiss and from light-colored tonalite in migmatitic rocks of the Swakane indicate that the latest episode of regional metamorphism, which crystallized or recrystallized the gneiss, took place during the Late Cretaceous (Mattinson, 1972, p. 3779).

Mad River Terrane

Structurally overlying the Swakane terrane is the Mad River terrane (Figs. 3, 4). We include in this terrane the heterogeneous schist-and-gneiss unit of Tabor and others (1980, in press), the rocks of the Napeequa River area (Crowder and others, 1966; Cater and Crowder, 1967), as well as metasedimentary and metavolcanic rocks along the west side of the Entiat Mountains referred to as the "younger gneissic rocks of the Holden area" by Cater and Crowder (1967) and Cater and Wright (1967). The terrane is named for the Mad River, a major tributary of the Entiat River.

The Mad River terrane consists predominantly of hornblende schist, schistose amphibolite, micaceous quartz schist, and micaceous quartzite, as well as rare biotite gneiss, metaconglomerate, calc-silicate schist, and marble. The predominance of amphibolitic rocks and quartzitic schist suggests a supracrustal, oceanic protolith of mostly mafic lavas and chert. Within the metasedimentary rocks are masses of blastoporphyritic biotite gneiss, flaser gneiss, uniform biotite granodiorite gneiss, zoisite amphibolite, and amphibolite gneiss, all of probable intrusive igneous origin, as well as silicic metaporphyry dikes and small pods of ultramafic and talc-tremolite schist. A major body of metagabbro (Fig. 4) crops out south of the Ten Peak pluton on the edge of the Mad River terrane. This massive to slightly gneissic hornblende-zoisite metagabbro grades into highly foliated hornblende-zoisite gneiss and into schist. The common occurrence of layers of zoisite-amphibolite gneiss in rocks of the Mad River terrane to the north, in the area west of Trinity, indicates that this gabbro may have been a widespread constituent of the terrane.

The protolithic age of the rocks in the Mad River terrane is uncertain. U-Th-Pb zircon ages of two adjacent samples of biotite granodiorite orthogneiss (Fig. 3) are concordant to moderately discordant (Tabor and others, in press). The $^{207}Pb/^{206}Pb$ ages of zircons systematically increase from about 75 to 127 m.y. with increasing grain size, whereas the U-Th-Pb ages remain almost constant at about 75 m.y. The biotite orthogneiss appears to contain a mixture of newly crystallized Late Cretaceous and inherited, pre-Cretaceous, probably Paleozoic or older zircons.

U-Th-Pb analysis of zircons separated from a zoisite-amphibolite gneiss from the strongly foliated margin of the metagabbro south of the Ten Peak pluton (Fig. 4) does not clarify the protolithic ages. Although the $^{207}Pb/^{206}Pb$ ages of 88 to 91 m.y. appear to record Late Cretaceous metamorphism and associated intrusion of the nearby Ten Peak pluton, the U-Pb and Th-Pb ages of 45 to 50 m.y. demand considerable recrystallization during a Tertiary event (Tabor and others, in press). Apparently, some of the zircons originally were considerably older than Late Cretaceous, but their age has been fortuitously reduced to about 90

m.y. Although evidence of an Eocene thermal event is widespread in the Pacific Northwest and Canada, only recently has dynamothermal metamorphism of Eocene age been documented elsewhere in the North Cascades (Hoppe, 1982; Tabor and others, in press).

Evidence for Faulted Contacts of the Mad River Terrane

In contrast to the broadly folded uniform foliation in the Swakane terrane (Fig. 3), foliation in the rocks of the overlying Mad River terrane is commonly folded isoclinally on an outcrop scale. The contact between these two terranes is abrupt; it appears to be conformable with the uniform foliation in the Swakane, but not with the folds and foliation in the schist, indicating that the contact is a fault. Although the Swakane contains rare layers of quartzitic schist, schistose amphibolite, calc-silicate schist, and marble (rocks typical of the Mad River terrane), and the overlying Mad River terrane locally contains thin layers of biotite gneiss which resemble the Swakane, neither metamorphosed intrusive rocks nor ultramafites typical of the Mad River terrane occur in the underlying Swakane terrane. Sill-like bodies of metatonalite and tonalite flaser orthogneiss along the contact suggest an intruded fault zone (Cater and Crowder, 1967; Cater, 1982, p. 37-40; Tabor and others, in press).

The contact of the Mad River terrane with the Nason terrane (Fig. 4) is enigmatic. The southern margin of the metagabbro in the Mad River terrane is intensely deformed and is locally imbricated with fine-grained, low-grade schist of the Nason terrane (Tabor and others, 1980, in press). To the northwest, much of the contact between the two terranes is intruded by the Late Cretaceous Ten Peak pluton (Fig. 4), but a re-entrant on the southwest side of the pluton preserves micaceous quartzite, mica schist, ultramafite, and thick pods of mafic breccia bearing rounded marble clasts, which Tabor and others (in press) interpret as a metamorphosed volcanic breccia. The rocks in this zone around the south end of the Ten Peak pluton are unlike rocks elsewhere in the adjacent terranes, and intense local tectonism and evident tectonic mixing make them candidates for slices in a major fault zone between terranes. In contrast to our interpretation, Van Diver (1967) thought that this zone was a deep-seated fault associated with emplacement of the Ten Peak pluton. Cater (1982, p. 32), however, considered the zone a protoclastic border of the intrusive Ten Peak pluton. As we explain below, evidence that the Ten Peak was intruded at much greater depths than the Mount Stuart batholith in the center of the Nason terrane may require a large amount of uplift of the Ten Peak (and adjacent Mad River terrane) relative to the Nason terrane.

The contact between the pelitic schist of the Nason terrane (central schist belt of Crowder and others, 1966) and the characteristic micaceous quartzite and fine-grained amphibolite and ultramafic pods of the Mad River terrane (rocks of the Napeequa River area) continues beyond the Ten Peak pluton to the northwest, but the only suggestion that the contact is tectonic is a concentration of ultramafite pods and layers in the area of the contact.

Chelan Mountains Terrane

The Chelan Mountains terrane (Fig. 5) is composed of (1) metamorphosed marine sedimentary and volcanic rocks; (2) abundant plutonic rocks, including migmatite and associated tonalite of the Chelan Complex of Hopson and Mattinson (1971); and (3) plutons of the Dumbell-Marblemount belt. The terrane is named for the Chelan Mountains, which extend from the Columbia River northwestward along the west side of Lake Chelan to Holden.

Metasedimentary and metavolcanic rocks

The oldest components of the Chelan Mountains terrane are the metasedimentary and metavolcanic rocks, predominantly schistose amphibolite, biotite schist, quartzitic schist, and rare marble. These rocks have several names. Near 48° N. latitude along the northeast side of the terrane (Fig. 5), Tabor and others (1980) referred to them as the amphibolite and schist of Twenty-Five Mile Creek. Northward, the rocks are continuous with the "younger gneissic rocks of the Holden area" (Cater and Crowder 1967; Cater and Wright, 1967), which, in turn, can be traced almost continuously northwestward for about 30 km to the Cascade River area (northwest of Cascade Pass; Fig. 5); (Tabor, 1961, pl. 24; Grant, 1966, pl. II), where the metamorphism diminishes and the rocks were called the Cascade River Schist by Misch (1966, p. 112).

Cater and Crowder (1967) correlated the younger gneissic rocks of the Holden area proper with rocks along the west side of the Entiat Mountains and the rocks of the Napeequa River area in our Mad River terrane (Fig. 4). In the Cascade River area (Fig. 5, inset) lower grade rocks of both terranes, predominantly banded metachert, amphibolite, and talc schist are indistinguishable. (See, for example, Misch, 1979.) The units also share in common marble, metaconglomerate, a variety of metamorphosed plutons, and ultramafic rocks. This lithologic similarity between the marine metasedimentary and metavolcanic rocks of

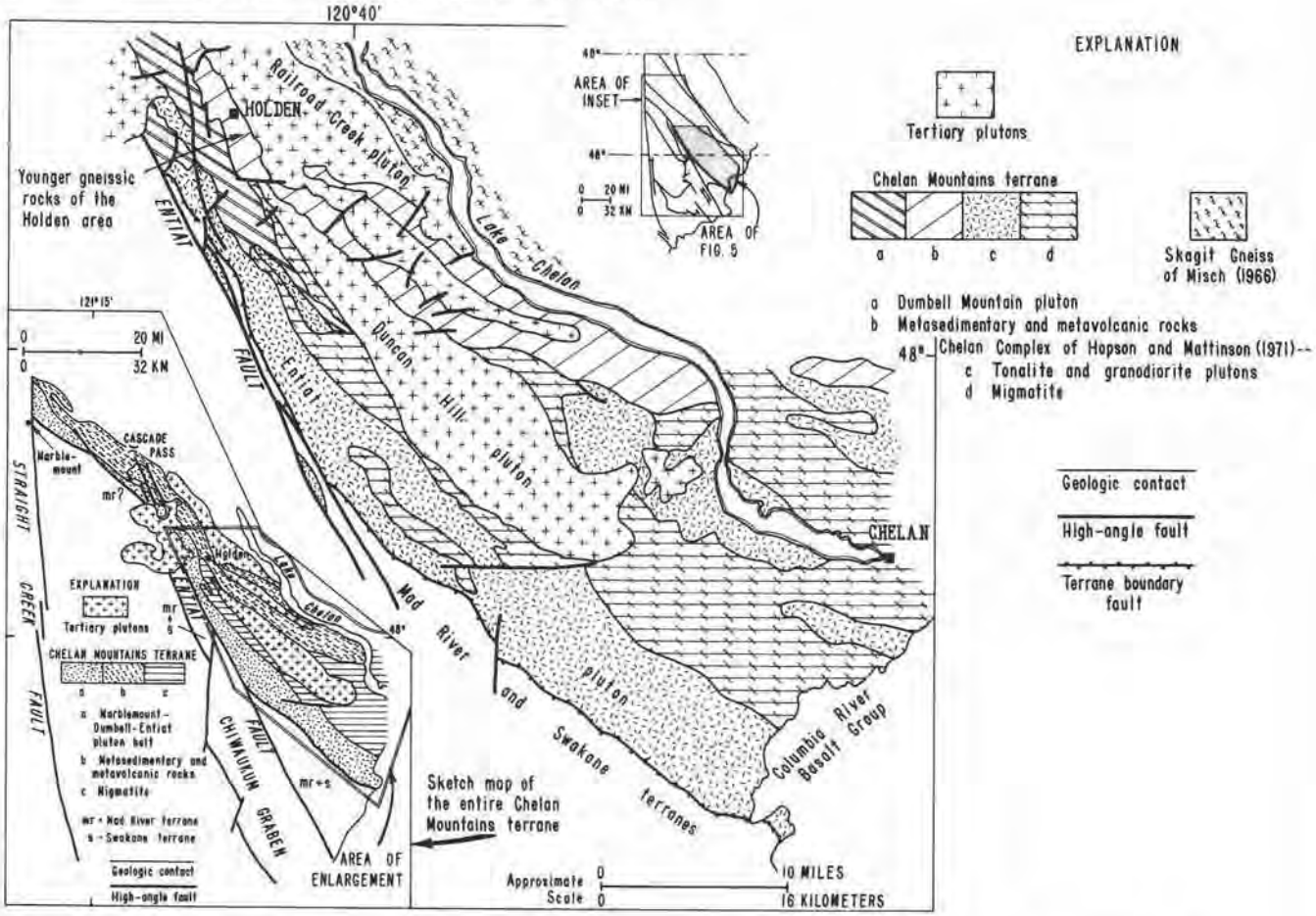

Figure 5.–Simplified geologic map of the southern part of the Chelan Mountains terrane. From Cater and Crowder (1967), Cater and Wright (1967), and Tabor and others (1980, in press).

the Chelan Mountains terrane and those of the Mad River terrane (Table 1) makes correlation attractive. However, the two units are nowhere continuous or in normal contact. The Cascade River Schist unconformably overlies the Marblemount Meta Quartz Diorite of Misch (1966), as described below, whereas the crystalline basement rocks appear to have been thrust over rocks of the Napeequa River area (based on recent U.S. Geological Survey reconnaissance studies). Although here we consider the Mad River and Chelan Mountains terranes to be separate, the supracrustal rocks of both terranes may once have been depositionally continuous and subsequently thrust over the Swakane Biotite Gneiss. In this alternative scenario, the Mad River and Chelan Mountains terrane are the same.

South of Holden, the supracrustal rocks in the Chelan Mountains terrane grade southward into migmatite as the proportion of light-colored tonalite sills and dikes and swirled tonalitic gneiss increases (Fig. 5). Schistose rocks of probable supracrustal derivation, especially schistose amphibolite, occur in small amounts throughout the migmatite terrane. Hopson and Mattinson (1971) referred many of the rocks of the Chelan Mountains area to the Chelan Complex, and we use their name to include the migmatite, as well as massive and gneissic tonalite.

Waters (1938) described in detail the feldspathization of supracrustal amphibolite and the production of migmatite in the Chelan Complex, a process he ascribed to contact metamorphism by massive igneous granitoid rock of the Chelan batholith. Hopson (1955) proposed that the massive igneous rocks, including tonalite along Lake Chelan and the elongate Entiat pluton, were derived by feldspathization and mobilization of amphibolite and schist during metamorphic processes. Later, Hopson and Mattinson (1971) emphasized that Late Cretaceous to earliest Tertiary ultrametamorphism produced numerous light-colored tonalite neosomes about 90 to 56 m.y. ago that were mobilized to a plastic mush by partial melting. In their view, most of the mobilized material was derived from the older igneous tonalite of the 220-m.y.-old (Triassic) Dumbell-Marblemount belt, and locally the ana-

tectic melt, or mush, intruded the supracrustal rocks. We think that the supracrustal rocks were also migmatized. Tabor and others (in press) confirm that the ultrametamorphism led to melting, but they also suggest that the massive tonalite plutons thus formed in the Late Cretaceous were further metamorphosed during the latest Cretaceous and earliest Tertiary.

Dumbell-Marblemount Belt

Extremely elongate tonalite plutons, including the Dumbell Mountain plutons in the Holden area (Fig. 5; Cater and Crowder, 1967; Cater and Wright, 1967; Cater, 1982, p. 9-23), crop out immediately southwest of the metasedimentary-metavolcanic belt. These plutons also have been traced northwestward (Tabor, 1961, pl. 24; Grant, 1966, pl. II), where they have been called the Marblemount Meta Quartz Diorite (Misch, 1966, p. 102). Zircons from the Marblemount and the Dumbell Mountain plutons yield relatively concordant 220-m.y. U-Pb ages (Mattinson, 1972, p. 3778). In the Holden area, Cater (1982, p. 9) considered these Triassic plutons to be relatively unmetamorphosed and intrusive into the adjacent metasedimentary and metavolcanic rocks. In the Cascade River area, where metamorphism is far less intense, Tabor (1961, p. 105-106) and Misch (1966, p. 112) described tonalite boulders in metaconglomerate as evidence that the protolith of the metasedimentary material was deposited on the Marblemount pluton. A recent study in the Cascade River area suggests that Marblemount plutonic rocks have been oceanic basement to nearly contemporaneous protolith basalt in the Cascade River Schist (Fugro Northwest, Inc., 1979, p. 13-15). There seems to be little doubt that the Dumbell-Marblemount belt is of igneous origin and that the metamorphism increases from greenschist facies in the Cascade River area southeastward to amphibolite facies in the Holden area. Although Cater's (1982, p. 9-10) conclusions that the deformational textures in the Dumbell are protoclastic and that the pluton is not significantly metamorphosed are untenable, the conflicting interpretations of the contact remain unresolved.

On the southeast, continuous with the Dumbell-Marblemount belt, are the elongate Seven-Fingered Jack and Entiat plutons, which, according to Cater (1982, p. 23), intrude the Dumbell plutons. These plutons (all shown as Entiat on Figs. 5 and 6) are intimately associated with the migmatite described above, and, as Hopson and Mattinson (1971) indicated, appear to be at least in part a final anatectic phase of the migmatite. More significantly, they may be mostly derived from the Triassic plutons of the Dumbell-Marblemount belt. As remobilized plutons, they might well intrude parts of their parent material. Cater (1982, p. 23-24) considered the younger Entiat pluton (and the Seven-Fingered Jack plutons) to be only incipiently metamorphosed. Features that we consider metamorphic Cater ascribed to protoclasis during intrusion at mesozonal depths. He did not accept the Late Cretaceous and earliest Tertiary regional metamorphic event in the North Cascades, although its validity is well established there and in contiguous British Columbia (Misch, 1971; Mattinson, 1972, p. 3778-3779; Fox and others, 1977; Monger and others, 1982; Tabor and others, 1980). Although proof that the Entiat is derived by partial anatexis of the Triassic plutons is elusive, the continuity of the tonalite plutonic belt adjacent to the supracrustal rocks and migmatite is consistent with a common origin for these plutons.

The overall picture of the Chelan Mountains terrane is of two belts: tonalite plutons on the southwest, and supracrustal rocks on the northeast. These belts grade from greenschist-facies rocks on the northwest, through well-recrystallized amphibolite-facies rocks in the Holden area, to migmatite and anatectic melt on the southeast.

Age Relations

U-Pb isotope ratios of zircon from biotite-quartz-oligoclase granofels in the metasedimentary and metavolcanic rocks (the younger gneissic rocks of the Holden area) give an older concordia intercept of 265 ± 15 m.y. if the younger intercept is assumed to be 90 to 60 m.y., the age bracket for the Late Cretaceous and earliest Tertiary metamorphism (Mattinson, 1972, p. 3773). The granofels near Holden was interpreted as a metamorphosed keratophyre by C. A. Hopson (as reported in Mattinson, 1972, p. 3273); thus the 265-m.y. (Permian) age is believed to represent the depositional age. The Permian age supports Cater's (1982, p. 9) conclusion that the supracrustal rocks were intruded by the Triassic Dumbell plutons, but, as we have discussed above for the Swakane Biotite Gneiss, the zircons could be detrital and the rock itself could be much younger. We can now conclude only that the protolithic age of the metasedimentary-metavolcanic component of the Chelan Mountains terrane is no older than Permian and could be Triassic or younger.

Predominantly discordant U-Pb ages of zircons and sphenes from the Chelan Complex reveal both the Late Cretaceous to earliest Tertiary plutonic and metamorphic events and the presence of older lead from the protoliths. As suggested by Hopson and Mattinson (1971) and Mattinson (1972, p. 3778), a 185-m.y.

age of zircon from banded migmatite just east of the Entiat pluton (Tabor and others, in press) and the mildly discordant ages (99-127 m.y.) of zircon from light-colored tonalite in migmatite (Mattinson, 1972, p. 3774) seem to reflect the influence of the pre-Cretaceous protoliths.

Zircon from the flaser-gneiss border of the Entiat pluton in the Chelan Mountains terrane adjacent to the supracrustal rocks of the Mad River terrane, with $^{207}Pb/^{206}Pb$ ages of 178 and 260 m.y., may have in part been picked up from the Mad River terrane by igneous assimilation or tectonic imbrication (Tabor and others, in press). Although we do not know whether such inherited zircon was present as detrital or primary grains in its source rock—even whether the Mad River rocks were in fact the source rock—the zircon has an inherited component with a Permian minimum age, but more likely a Precambrian age (Tabor and others, in press).

Nearly concordant ages from the center of the Entiat pluton and from mafic tonalite south of Lake Chelan clearly record the crystallization of zircon 85 to 75 m.y. ago. Because the plutons in these localities are massive, uniform tonalite displaying relict igneous textures, they probably were formed from true igneous melts during the Late Cretaceous. We assert that here, especially, the mineralogy and texture of the parent material of the anatectic plutons have been destroyed, and also that the superimposed effects of post-crystallization metamorphism are minimal.

Zircon with a $^{207}Pb/^{206}Pb$ age of 175 m.y. from tonalite along Lake Chelan (Mattinson, 1972, p. 3778) and zircon with $^{207}Pb/^{206}Pb$ ages of 178 and 260 m.y. from the flaser-gneiss border zone of the Entiat pluton give evidence for the retention of older lead even in parts of the anatectic melts.

Because of lithologic similarity, we think that much of the massive tonalite in the Chelan Complex was formed during the Late Cretaceous from melts derived by mobilization or anatexis of older rocks, as indicated by Hopson and Mattinson (1971). The gradation of the more massive plutons into migmatite and the spectrum of U-Th-Pb zircon ages representing mixtures of inherited and new components in this refractory mineral strongly support such an interpretation.

Metamorphic overprinting of the plutons in the Chelan Complex must have been contemporaneous with or must have immediately postdated their emplacement. Metamorphic sphene yields U-Pb ages of 86 to 70 m.y. (Mattinson, 1972, p. 3774), and K-Ar ages of hornblende and biotite are as much as 84 m.y. and 74 m.y., respectively (Tabor and others, in press).

Contacts of the Chelan Mountains Terrane with Adjoining Terranes

Much of the contact between the Chelan Mountains and Mad River terranes lies in the younger Entiat fault zone (Fig. 5), and we interpret the slivers of metatonalite mapped in the Mad River terrane to be fault slices.

In the area of the contact that we have studied in most detail, the Entiat pluton forms much of the contact between the Chelan Mountains and Mad River terranes. If, as we believe, the Entiat pluton was mobilized during the Late Cretaceous, it could be intrusive along this contact. Cater (1982, p. 23-28) indicated that the pluton is intrusive into wall rocks and older plutons, although we have found no unquestionable evidence of intrusion. The contact with the Mad River terrane at the southeast end of the pluton appears to be a deep-seated fault. The Entiat pluton has been highly deformed into a fine-grained flaser gneiss at the contact, and metachert nearby in the Mad River terrane is recrystallized mylonite.

To the northwest (Fig. 5), the western edge of the Chelan Mountains terrane is underlain by the Marblemount pluton, which is also faulted, locally at a low angle, against metasedimentary rocks of the Mad River terrane (Tabor, 1961, pl. 24).

Relation of the Skagit Gneiss of Misch (1966) to the Chelan Mountains and Swakane Terranes

The migmatitic Skagit Gneiss of Misch (1966), also referred to as the Custer Gneiss of McTaggart and Thompson (1967), lies northeast of the metasedimentary and metavolcanic rocks of the Chelan Mountains terrane (Fig. 1). A fundamental question is whether the Skagit is part of the terranes described here or constitutes a separate terrane. On the north, Misch (1966, p. 112-113; 1968) considered that much paragneiss of the Skagit was derived by metasomatic processes from the Cascade River Schist (which we include in the Chelan Mountains terrane). Babcock (1970) supported Misch's thesis with chemical comparisons, but in a later paper (Babcock and others, 1975) indicated a complex relationship inferred from isotopic studies. Much of the contact between the Chelan Mountains terrane and the Skagit Gneiss is between the Cascade River Schist and orthogneiss intimately associated with Skagit (Misch, 1979) or is intruded by terrane-stitching plutons ranging in age from late Cretaceous to Eocene (Tabor, 1961, pl. 24; Libby, 1964, p. 127; Cater and Wright, 1967; Misch, 1979). The derivation of the paragneiss of the Skagit

Gneiss from Cascade River Schist seems probable but is yet to be fully established.

At the southeast end of the Chelan Mountains terrane (Fig. 5), Cater and Wright (1967) and Cater (1982, p. 7) considered the Skagit to be correlative with the Swakane Biotite Gneiss and to be structurally overlain by the younger gneissic rocks of the Holden area (and, by implication, the Cascade River Schist) of the Chelan Mountains terrane. Important to their argument is the assumption that the younger gneissic rocks of the Holden area are equivalent to rocks of the Mad River terrane, which do indeed appear to structurally overlie the Swakane Biotite Gneiss. As we mentioned above, correlation of the metasedimentary and metavolcanic rocks in these two terranes is attractive and deserves further consideration, but is not directly germane to a Skagit-Swakane correlation.

Although the Skagit Gneiss superficially resembles the Swakane Biotite Gneiss, it differs in at least two respects: (1) The much more heterogeneous Skagit Gneiss commonly contains numerous layers of schist and amphibolite of probable supracrustal origin (Adams, 1961, p. 20-81; Libby, 1964, p. 71-91; Misch, 1966, p. 113; Staatz and others, 1972, p. 8-15). (2) In more uniform parts of the Skagit, the gneiss contains considerably more amphibole (mostly hornblende) than the Swakane. (See, for example, Adams, 1961, p. 31; Libby, 1964, p. 71, 74.) We do not know the age of the Skagit Gneiss, although Misch (1966, p. 113) stated that its protolithic age might be Paleozoic. Zircons from the Skagit Gneiss yield discordant U-Pb ages which, when plotted on a concordia diagram, have been interpreted as the ages of Precambrian detrital grains in the possible Paleozoic protolithic sediment (Mattinson, 1972, p. 3773). Although this zircon evidence somewhat parallels that of the Swakane, overall composition does not indicate to us that the Skagit Gneiss and Swakane Biotite Gneiss are correlative. For the present, the Skagit Gneiss of Misch should be considered separate from both the Swakane and Chelan Mountains terranes.

The contact between the supracrustal rocks of the Chelan Mountains terrane and the Skagit Gneiss is marked by elongate plutons, a relation suggesting that the plutons were emplaced along a major tectonic zone. In fact, viewed as a whole, the Chelan Mountains terrane is a swarm of elongate plutons; the whole terrane may be a long-lived zone of faulting and intrusion.

METAMORPHISM, PLUTONISM, AND EMPLACEMENT OF THE TERRANES

An exhaustive discussion of the complex emplacement history of the five North Cascades terranes described here is beyond the scope of this report; we can, however, outline some important aspects and restraints. Our knowledge of the terranes' amalgamation prior to their arrival at the North American plate margin is meager. The probable thrust contacts between the Mad River and the underlying Swakane terrane and between the Ingalls Tectonic Complex and the underlying Nason terrane may be evidence of prearrival assemblage of the paired terranes. The low-angle thrusts contrast with, and appear to be bounded by, probably younger, higher angle faults that bound the terranes elsewhere. Whetten and others (1980) considered the Ingalls Tectonic Complex to be part of a regional thrust plate, their Haystack thrust, exposed in the San Juan Islands and elsewhere west of the Straight Creek fault. In their view, the Haystack thrust was emplaced during the middle Cretaceous, presumably on top of already assembled and accreted terranes of the North Cascades. Vance and others (1980) also envisioned northward thrusting (obduction) of the Ingalls Tectonic Complex onto the Nason terrane during the middle Cretaceous. We have no evidence to dispute this timing, but if the large amount of post-Late Cretaceous offset along the Straight Creek fault of the Nason terrane proposed by Misch (1977) is correct, correlation of the Ingalls with rocks much to the south of the northward-displaced Nason Terrane seems unlikely.

Late Cretaceous and earliest Tertiary (90-60 m.y.) metamorphism in the Chelan Mountains terrane is well documented by the U-Pb isotope studies of Mattinson (1972). His (1972, p. 3374) analyses of zircons from metamorphic pegmatite in the Skagit Gneiss of Misch (1966) confirm the age of this regional event. Although U-Th-Pb ages have not been obtained from metamorphic country rock or from metamorphic differentiates in the other terranes, analyses of zircons from synmetamorphic plutons document Late Cretaceous metamorphism. The Sloan Creek plutons (Fig. 6) yield concordant 90-m.y. ages (Tabor and others, 1982d), and the Eldorado pluton yields 92-m.y. ages (Mattinson, 1972, p. 3774).

Conventional K-Ar ages for the plutons are concordant in the range 80 to 90 m.y., but younger ages are somewhat discordant, especially in the Chelan Mountains terrane (Tabor and others, 1980, in press). In the Nason terrane, however, K-Ar biotite and muscovite ages from well-recrystallized orthogneiss and metamorphic pegmatite and hornblende ages from amphibolite range from 81 to 85 m.y. In other terranes, K-Ar ages are more discordant and range from about 50 to 75 m.y. (Tabor and others, 1982d, in press).

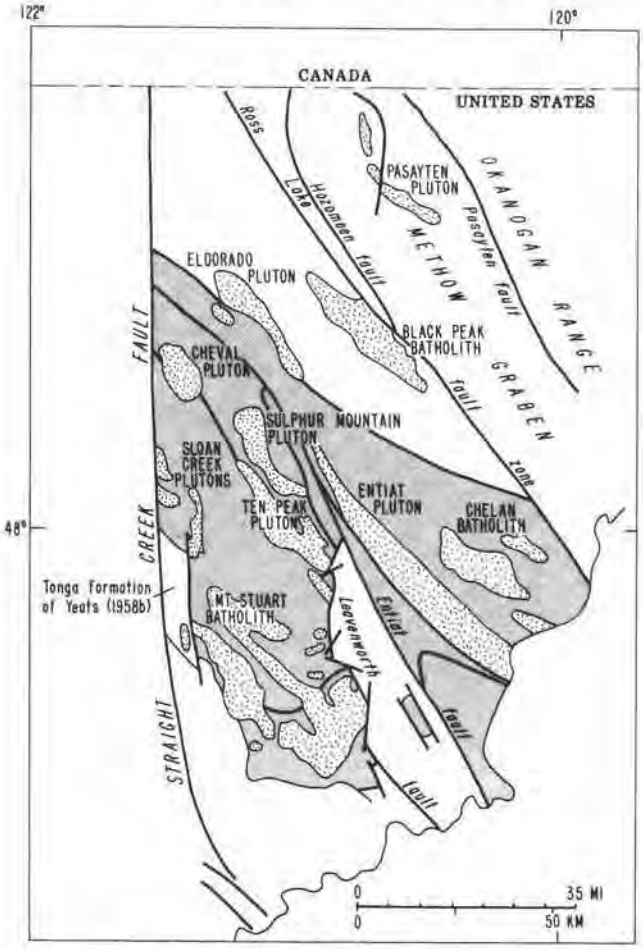

Figure 6.–Locations of major Cretaceous syntectonic plutons in the North Cascades. Shaded terranes and Tonga Formation of Yeats (1958b) are discussed in text. Compare with Figure 1.

The deep-seated, synmetamorphic plutons are characterized by both igneous and metamorphic features and a general lack of static thermal metamorphism in their host rocks. These plutons intrude all the terranes described here, and their shapes parallel the metamorphic grain (Fig. 6). The Mount Stuart batholith exhibits the least metamorphic character and has generated the most pronounced thermal effects in the country rocks (Erikson, 1977a, 1977b; Plummer, 1980; Tabor and others, 1980, 1982b). On its south end, the Mount Stuart batholith clearly reached relatively shallow depths, because the intruded rocks of the Ingalls Tectonic Complex are now hornfels. According to Kaneda (1980, p. 121-122), peak conditions of thermal metamorphism associated with the intrusion of the Mount Stuart batholith into the Chiwaukum Schist, about midway along the length of the pluton, involved temperatures of approximately 650° to 750°C and pressures of approximately 3 ± 1 kbar (corresponding a depth of approximately 10 km). The better developed metamorphic fabrics of many of the other plutons indicate that they were deeper seated; their general characteristics indicate that they were intruded in the mesozone of Buddington (1959), defined as extending from 7 to 16 km.

Zen and Hammarstrom (1984, p. 515) suggested that the Ten Peak and Sulphur Mountain plutons intruded at depths greater than 25 km, a circumstance inferred from their contained magmatic epidote. Relics of probable magmatic epidote in the Chelan and Entiat plutons (Tabor and others, in press) and epidote of possible igneous origin in the Chaval pluton (Boak, 1977, p. 52-53) indicate similar depths of formation. All the other plutons of Figure 6 apparently lack magmatic epidote (Heath, 1971, p.55-65; Tabor, 1961, p. 146-150; Adams, 1961, p. 93-114), although the deformation and metamorphism of the Eldorado pluton may have destroyed magmatic epidote. According to Buddington's (1959) depth criteria, the migmatitic aspect of the plutons in the Chelan Mountains terrane suggests that they were formed even deeper than the Ten Peak and Sulphur Mountain plutons. Furthermore, the continuing metamorphism which overprinted the plutons of the Chelan Mountains terrane, the discordant U-Th-Pb ages of these plutons, and their range of younger K-Ar ages indicate that these rocks remained at depth longer than those of the Nason terrane. The apparently different depths of intrusion of the plutons may be evidence that terranes were uplifted differentially along the terrane-bounding faults, but the history is complex, because some of the plutons, such as the Ten Peak and Eldorado, appear to straddle the faults (Fig. 6).

By latest Cretaceous time, some of the stitches seemed to have been in place in the patchwork quilt of the North Cascades. The territory embraced by metamorphism and plutonism (Fig. 6) indicates that the terranes were assembled by the Late Cretaceous.

On the west, the Late Cretaceous metamorphism and plutonism has affected part of the Easton Schist (that is, the Tonga Formation of Yeats, 1958b). Thus, at least some of the lower grade rocks that are mostly west of the Straight Creek fault were also assembled with rocks now in the crystalline core before Late Cretaceous time, but not earlier than 130 m.y. ago (Early Cretaceous), the maximum age being based on the probable age of blueschist metamorphism in the Easton Schist (Brown and others, 1982, p. 1095). The present distribution of the Easton Schist may reflect considerable strike-slip movement along the Straight Creek fault and/or movement on other faults since Late Cretaceous time.

Monger and others (1982, p. 73) proposed that Cascadia (including the North Cascades) was sutured

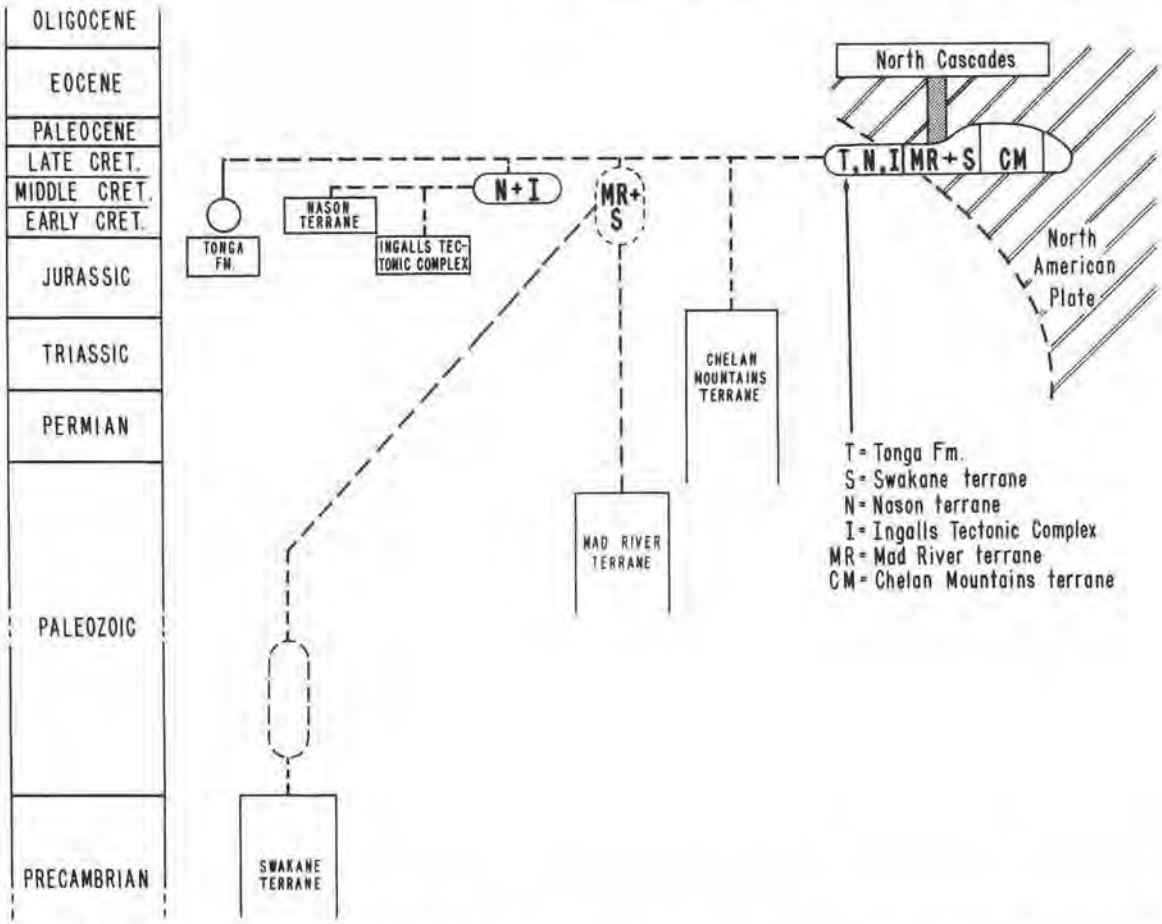

Figure 7.–Summary diagram illustrating history of tectonostratigraphic terranes and the Tonga Formation of Yeats (1958b) of the North Cascades, inspired by a diagram of Monger and others (1982, fig. 2). Positions of rectangles and ovals show ages of stratigraphic and intrusive components and probable ages of amalgamation and metamorphism, respectively. Dashed enclosures denote large uncertainly in age. Thick, shaded vertical bar denotes translation by faulting. Paleogeographic positions or structural relations between units are not shown.

to North America before the Late Cretaceous because it shares the Late Cretaceous metamorphism with the coast plutonic complex of British Columbia. They believed that the metamorphism and plutonism were caused by obduction or crustal thickening that occurred when a composite terrane including Wrangellia and the Alexander terrane crashed into North America during Late Cretaceous time. Zen and Hammarstrom (1984) supported this view with their estimates of great crustal thickness based on the pressure-temperature conditions necessary for the formation of igneous epidote.

Nonetheless, rotated paleomagnetic poles of the Late Cretaceous Mount Stuart batholith suggest that the North Cascades tectonostratigraphic terranes may not have been at their present position on the margin of the North American plate by the Late Cretaceous. Paleomagnetic poles from the Mount Stuart batholith indicate as much as 29°of clockwise rotation since the Late Cretaceous, and magnetic inclination indicates considerable northward translation (Beck, 1980, p. 7117-7119; Beck and others, 1981). We know that the batholith is unconformably overlain by steeply dipping beds of lower Eocene sandstone (Yeats, 1958a, pl. 1; McDougall, 1980, appendix A) and that it probably has been uplifted more at its north end than at its south end. Although Beck (1980, p. 7125) argued persuasively on theoretical grounds that deep-seated tilt cannot have had a significant effect on the observed rotations, considerable tilting has taken place since the early Eocene. According to Beck's arguments, the batholith could not have tilted as a whole, and so it must have broken into small tilted blocks, bounded by structures as yet unidentified in the field. In contrast to Beck's conclusion, Strickler (1982, p. 59-71) believed that the Late Cretaceous Black Peak batholith (Fig. 6) probably was tilted locally, which would account for the discordant paleomagnetic poles.

We are uncertain how much of the pole discrepancy can be eliminated by tilting of the Mount Stuart batholith, but, assuming that significant rotation and northward translation of the Nason terrane has taken place since the Late Cretaceous, further studies must look to the east of the crystalline backbone of the North Cascades to find the suture with North America. From available paleomagnetic data we know that the terranes were essentially in their present position by late Eocene time, because Tertiary plutons (younger than about 43 m.y.) show no anomalous paleomagnetic pole orientations (Beck and others, 1982).

CONCLUSIONS

The major terrane units of the North Cascades crystalline core appear to be readily identified by their lithology, to some extent by their style of deformation and metamorphism, and by their probable age. Table 1 summarizes their characteristics and emphasizes the differences among terranes. There is still much to be learned about the nature of their contacts and their history before assembly. The regional extent of Late Cretaceous and earliest Tertiary metamorphism and plutonism in the North Cascades suggests that the terranes were amalgamated by Late Cretaceous time, but not earlier than the Early Cretaceous (Fig. 7).

Although considerable paleomagnetic work remains to be done, rotated paleomagnetic poles in the Late Cretaceous Mount Stuart batholith suggest that the terranes, even though assembled by Late Cretaceous, had not arrived at their present position in North America until after that time. An accretion or post-accretion event like that championed by Monger and others (1982) may be the most likely agent of metamorphism and plutonism in the North Cascades block; if so, northward translation of the terranes must have been along strike-slip faults, much in the way envisioned by Davis (1977) and Davis and others (1978). The responsible fault or faults lie to the east of the main crystalline core of the North Cascades, and examination of that episode of transport history will have to come with future studies. The coincidence of paleomagnetic poles of middle Tertiary plutons with the expected pole indicates that the terranes had arrived in their present position before the late Eocene. More details of terrane history will probably be discovered as workers continue to explore the terrane concept in the North Cascades and examine the union of these terranes with the North American plate.

ACKNOWLEDGMENTS

We owe much of our knowledge of North Cascades geology to Peter Misch and his students, who, in a pioneering, long-sustained effort, mapped and described many of the fundamental units and major tectonic elements in northwestern Washington. Our ideas here are built on their work. Our colleagues have helped to stimulate some of the ideas presented here; in particular we cherish discussions with Ken Fox, Myrl Beck, Cliff Hopson, Jim Mattinson, Bob Miller, Clark Blake, Dave Miller, and Joe Vance. We thank Art Ford and Robert Fleck for permission to quote their preliminary Rb-Sr data from the Swakane Biotite Gneiss.

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