

GEOLOGIC MAP OF THE CHELAN 30-MINUTE BY 60-MINUTE QUADRANGLE, WASHINGTON

**By R. W. Tabor, V. A. Frizzell, Jr., J. T. Whetten, R. B. Waitt, D. A. Swanson,
G. R. Byerly, D. B. Booth, M. J. Hetherington, and R. E. Zartman**

INTRODUCTION

Bedrock of the Chelan 1:100,000 quadrangle displays a long and varied geologic history (fig. 1). Pioneer geologic work in the quadrangle began with Bailey Willis (1887, 1903) and I. C. Russell (1893, 1900). A. C. Waters (1930, 1932, 1938) made the first definitive geologic studies in the area (fig. 2). He mapped and described the metamorphic rocks and the lavas of the Columbia River Basalt Group in the vicinity of Chelan as well as the arkoses within the Chiwaukum graben (fig. 1). B. M. Page (1939a, b) detailed much of the structure and petrology of the metamorphic and igneous rocks in the Chiwaukum Mountains, further described the arkoses, and, for the first time, defined the alpine glacial stages in the area. C. L. Willis (1950, 1953) was the first to recognize the Chiwaukum graben, one of the more significant structural features of the region.

The pre-Tertiary schists and gneisses are continuous with rocks to the north included in the Skagit Metamorphic Suite of Misch (1966, p. 102-103). Peter Misch and his students established a framework of North Cascade metamorphic geology which underlies much of our construct, especially in the western part of the quadrangle.

Our work began in 1975 and was essentially completed in 1980. The Chelan 1: 100,000 sheet is the northeast quarter of the Wenatchee 1:250,000 quadrangle and is the second release in a series of the geologic maps which will cover this larger quadrangle. A preliminary open-file report (Tabor and others, 1980) preceded this map.

Responsibility for the mapping of the Columbia River Basalt Group lies mostly with Swanson and Byerly. Tabor, Frizzell, Whetten, Booth, and Hetherington mapped most of the other bedrock area; of that, most of the Chiwaukum graben was mapped by Whetten. Waitt mapped most of the unconsolidated deposits.

A considerable number of new radiometric age analyses are included in this report. The K-Ar biotite and hornblende ages were measured by Tabor and Hetherington, and the zircon and apatite fission-track ages were determined by Frizzell. Zartman determined U-Th-Pb zircon ages in support of the field investigations, and contributed to the interpretation of all geochronologic data.

Field assistants who gave much in the line of camp and mapping duty were Jay Coburn and Ron Tal, 1976; Bill Gaum and Kim Marcus, 1977; Brett Cox, Sam Johnson, Elizabeth Lincoln Mathieson, and Nora Shew, 1978; Joe Marquez, 1979, 1980; and Jim Talpey, 1979. Dennis Sorg produced numerous clean mineral separates.

Helicopter pilots John Nelson, Doug Bucklew, the late Jack Johnson, and Tim Bormin performed miraculous feats in rough terrain and weather. We appreciate their efforts to get us to the rocks.

No geologic study proceeds without considerable discussion and argument. We are particularly indebted to M. Clark Blake, Erik Erikson, Kenneth Fox, Ronald Frost, the late Randal Gresens, Clifford Hopson, James Mattinson, Robert Miller, Victoria Todd, and Robert Yeats for keeping us thinking. We are indebted to Charles W. Naeser for unpublished fission-track ages reported in table 1.

Summary of geologic history

Pre-Tertiary metamorphic rocks in the quadrangle can be subdivided into five major tectono-stratigraphic terranes: (1) the Ingalls, equivalent to the Ingalls Tectonic Complex, (2) the Nason, (3) the Swakane, (4) the Mad River, and (5) the Chelan Mountains (fig. 1). We use the term terrane in much the same sense as Beck and others (1980, p. 454): "a fault-bounded geologic entity characterized by a distinctive stratigraphic sequence and or structural history differing markedly from those of adjoining neighbors." Although the rocks within each of the terranes described here have some unique features suggesting separate early histories, they share a similar Late Cretaceous metamorphic history.

The Ingalls Tectonic Complex is a Late Jurassic or Early Cretaceous ophiolite melange (Southwick, 1974, p. 399-401; Hopson and Mattinson, 1973, p.57; Miller, 1977, p. 468) that has been thrust onto and locally imbricated with pelitic rocks of the Chiwaukum Schist (Miller, 1980b). The numerous ultramafic pods and layers in the Chiwaukum Schist, the principal unit in the Nason terrane, suggest that the schist protolith might be a facies of the oceanic assemblage of the Ingalls but the two units could also have been part of separate terranes with different early histories.

The Swakane terrane is composed entirely of the Swakane Biotite Gneiss. The gneiss is a uniform metaclastite (?) or metadacite (?) containing rounded zircons more than 1,650 m.y. old. Its protolithic age is at least pre-Cretaceous and presumably Precambrian (Mattinson 1972, p. 3773; Tabor and others, in press). The Swakane is tectonically overlain by the Mad River terrane, a heterogeneous terrane of mostly schist, micaceous quartzite, some marble, and orthogneiss. Complex U-Th-Pb systematics of zircons from an orthogneiss suggest a mixture of Paleozoic or older zircons and Late Cretaceous or early Tertiary metamorphic zircons. The protolithic age of the metasedimentary and metavolcanic rocks of the Mad River terrane is probably Paleozoic or older. Sharply bounded alaskite and granite pegmatite dikes and sills intruded the Swakane and Mad River terranes during the Late Cretaceous metamorphism (Mattinson, 1972, p. 3778-3779).

The Tertiary Chumstick Formation and the Late Ten Peak pluton obscure most of the contact between the Mad River and Nason terranes. However, on the southern end of Chiwawa Ridge unusual low-grade schists of the

MAP

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| 1. Cashman, 1974 (fig. 2) | C |
| 2. Frost, 1977 (written commun.) | C |
| 3. Danner, 1966 (figs. 196, 201, 202, 209) | A |
| 4. Erikson, 1977 (fig. 1) | B |
| 5. Getsinger, 1978 | B |
| 6. Gualtieri and others, 1973 (plate 1) | C |
| 7. Laravie, 1976 (plate 1) | B |
| 8. Lupe, 1971 (fig. 13) | C |
| 9. McKague, 1960 (plate V) | C |
| 10. Newcomb, 1970 | C |
| 11. Oles, 1956 (plate B) | C |
| 12. Page, 1939a (fig. 3) | C |
| 13. Plummer, 1969 (plate III) | C |
| 14. Plummer, 1969 (fig. 2) | B |
| 15. Pongsapich, 1974 (plate 1) | C |
| 16. Pratt, 1958 (plate 7) | C |
| 17. Rosenberg, 1961 (plate XI) | B |
| 18. Kienle, Farooqui, Strazer, Hamill, 1978 (fig. 2) | C |
| 19. Clayton, 1977 (fig. 2R D. 8-2) | C |
| 20. Clayton, 1977 (fig. 2R D. 8-3) | C |
| 21. Southwick, 1962 (plate 1) | B |
| 22. VanDiver, 1964 (plate II) | B |
| 23. Waters, 1932 (fig. 1) | C |
| 24. Waters, 1938 (plate 1) | C |
| 25. Whetten, 1980a | A |
| 26. Whetten, 1980b | A |
| 27. Whetten, 1980c | A |
| 28. Whetten and Laravie, 1976 | A |
| 29. Whetten and Waitt, 1978 | A |
| 30. Willis, 1950 (plate 46) | C |
| 31. Hopson, 1955 (plate 2) | B |
| 32. Hopson, C. A., (1965, written commun.) | B |
| 33. Nimick, 1977 | C |
| 34. Merrill, 1966 | C |
| 35. Waters, 1933 (fig. 1) | C |
| 36. Waters, 1939 (plate 1) | C |
| 37. Kaneda, 1980 (map 2) | B |
| 38. Miller, 1980b (plate 2) | B |
| 39. Link, M. J. (U.S. Forest Service, written commun., 1980) | B |

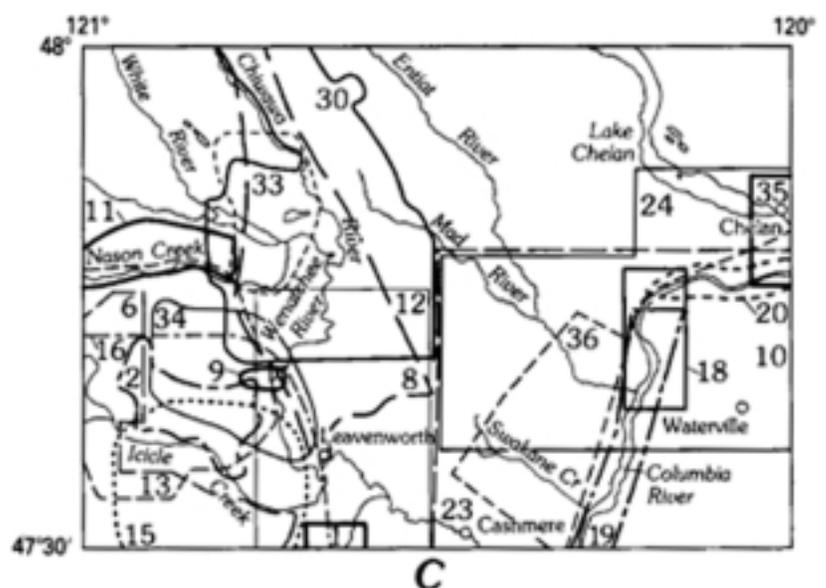
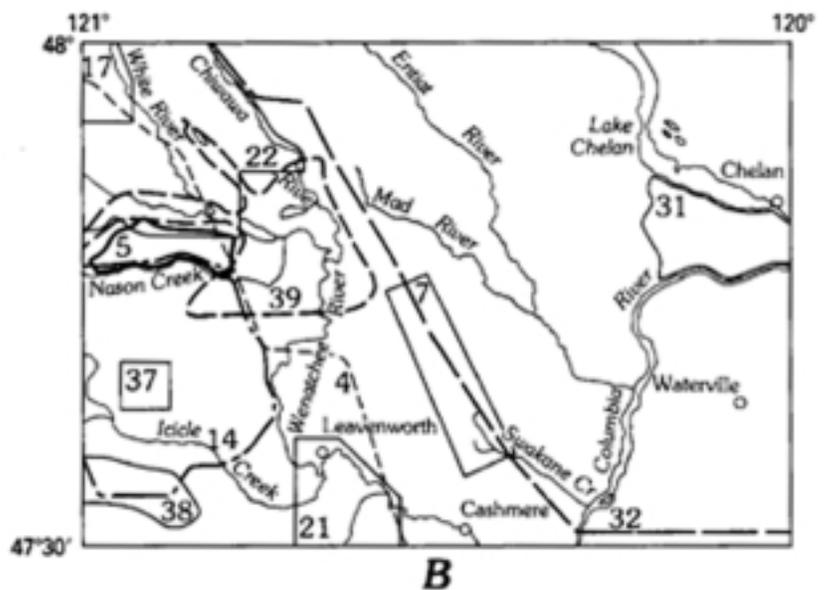
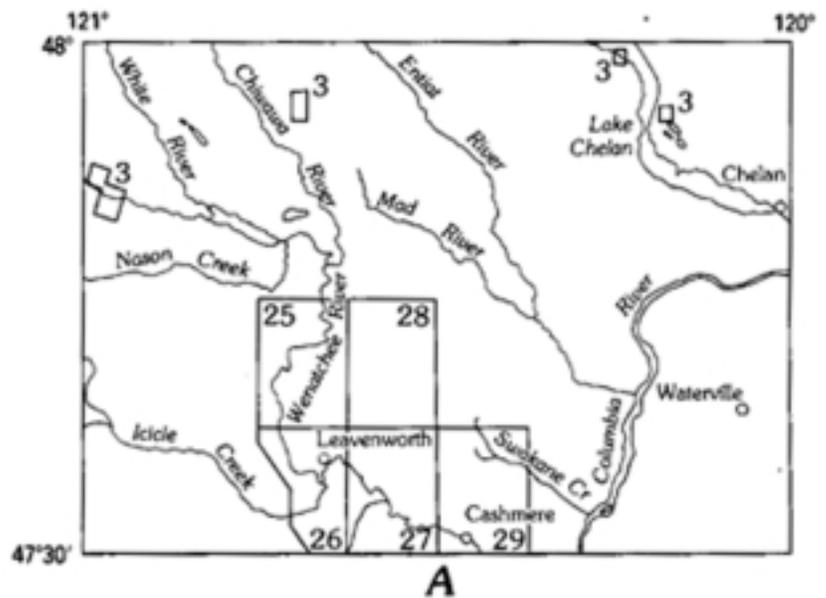


Figure 2.—Sources of map compilation data. A, Much data used with little to modest modification. B, Some data used. C, Consulted extensively but data not directly used on map.

Chiwaukum Schist are imbricated with the complexly deformed metagabbro and metadiorite body of the Mad River terrane. Numerous ultramafite pods in the schists and the sheared-out metadiorite suggest that this is the original suture between the terranes.

The Mad River terrane is separated from the highly contrasting migmatitic Chelan Complex of Hopson and Mattinson (1971) within the Chelan Mountains terrane which is also locally migmatized. Considerable mylonitization along the contact between the Mad River terrane and the Entiat pluton (within the Chelan Mountains terrane) suggests that a once deep-seated fault now separates the two terranes.

The Chelan Mountains terrane is dominated by the Chelan Complex of Hopson and Mattinson (1971) which is composed of migmatite and gneissic to tonalite of deep-seated igneous and metamorphic (Waters, 1938; Hopson, 1955; Hopson and Mattinson, 1971). The leucosome of the migmatite yields Late Cretaceous U-Pb ages (Mattinson, 1972, p. 3778-3779). The massive, possibly anatectic, tonalite is also in large part Late Cretaceous in age as shown by both U-Th-Pb and K-Ar ages, although relict zircons in several units yield a discordant pre-Cretaceous age (see Mattinson, 1972, p. 3775). The protolith of the migmatite includes both sedimentary and volcanic rocks of possible Permian age and Triassic plutons, be represented by the essentially isochemically metamorphosed rocks of Twentyfive Mile Creek and the Dumbell plutons, exposed in the Holden and Lucerne 15-minute quadrangles to the north (Cater and Crowder, 1967; Cater and Wright, 1967, Hopson and Mattinson, 1971; Mattinson, 1972, p. 3778).

During an episode of Late Cretaceous regional metamorphism, all the terranes were intruded by deep-seated tonalite to granodiorite plutons, including the Mount Stuart batholith, Ten Peak and Dirty Face plutons, and the Entiat pluton and massive granitoid rocks of the Chelan Complex of Hopson and Mattinson (1971). Tabor and others (in press) discuss probable pre-Late Cretaceous accretion of the terranes to North America.

The Duncan Hill pluton intruded rocks of the Chelan Mountains terrane about 45 to 48 m.y. ago (Cater and Wright, 1967). Accompanying this early Tertiary intrusion was local dynamothermal metamorphism, reflected in the gneissic root at the northern end of the Duncan Hill pluton exposed in the Holden quadrangle (Cater, 1982, p. 57) and zircon ages of deformed gabbro in the Mad River terrane. At about the same time fluvial arkosic sediment of the Chumstick Formation (Gresens and others, 1977, p. 100-108; Gresens and others, 1981) was deposited in a depression, now called the Chiwaukum graben (fig. 1).

The Mad River, Swakane, and Chelan Mountains terranes were riddled by 45 to 50-m.y.-old granite porphyry and rhyolite dikes related to the Duncan Hill pluton and the Cooper Mountain batholith (Barksdale, 1975, p. 17-18), which crops out just northeast of the Chelan quadrangle.

The outpouring of basalt lavas to the southeast of the quadrangle during the Miocene built up the Columbia River Basalt Group. These now slightly warped lavas lapped onto rocks of the Mad River, Swakane, and Chelan Mountains terranes.

Deformation, uplift, and erosion recorded in the rocks and deposits of the quadrangle continued into post-Miocene time. Quaternary deposits reflect advances of glaciers down the major valleys, a complicated history of catastrophic glacial floods down the Columbia River, the formation of lakes in the Columbia and Wenatchee river valleys by landslides and flood backwaters, and hillslope erosion by large and small landslides and debris flows.

Ages and Names

In the text of this report, we discuss both the metamorphic and protolith ages of metamorphic rocks and, where known, the time when tectono-stratigraphic terranes were assembled and the age of their constituent parts. The difficulty lies in placing the units in a correlation diagram and attaching meaningful map symbols. We have elected to designate almost all of the metamorphosed pre-Tertiary units with their metamorphic ages. Ghost boxes on the map sheet show their protolith age. We show the Ingalls Tectonic Complex in position of its rather indefinite assembling age rather than its Late Cretaceous metamorphic age because the unit grades rapidly to less metamorphosed rocks just south of the quadrangle.

The late metamorphic (Late Cretaceous) plutons that clearly post-date assemblage of the terranes, such as the Mount Stuart and Ten Peak, are excluded from the terrane units. Although the Late Cretaceous plutons in the Chelan Complex of the Chelan Mountains terrane are similar in principle to these other plutons, for convenience we describe them with the terrane because they are closely associated with their host rocks, show little evidence of intrusive transport, and have been strongly overprinted by the metamorphism affecting their host rocks.

Metamorphic and igneous rock names follow the standard procedure of placing the most abundant mineral name next to the rock name. Igneous rocks are named according to new IUGS recommendations (see fig. 3 and Streckeisen, 1973, 1979). Generally we use light-colored tonalite instead of trondhjemite or leucotondhjemite as have been used by workers in the migmatitic and gneissic terranes.

GENERAL DESCRIPTION OF THE UNITS

PRE-TERTIARY BEDROCK

By R. W. Tabor, V. A. Frizzell, Jr., J. T.

D. B. Booth, M. J. Hetherington, and R. E. Zartman

CHELAN MOUNTAINS TERRANE

The Chelan Mountains terrane is characterized by plutonic rocks including abundant migmatite and tonalite of the Chelan Complex of Hopson and Mattinson (1971) and minor metamorphosed supracrustal rocks. The Chelan Mountains extend from the Columbia River northwestward along the west side of Lake Chelan to Railroad Creek north of the quadrangle.

Amphibolite and schist of Twentyfive Mile Creek

Schistose amphibolite, biotite schist, siliceous schist, and rare marble on the northern edge of the Chelan quadrangle are continuous with the younger gneissic rocks of the Holden area in the Lucerne and Holden 15-minute quadrangles to the north (Cater and Wright, 1967; Cater and Crowder, 1967). Amphibolite and siliceous schist that

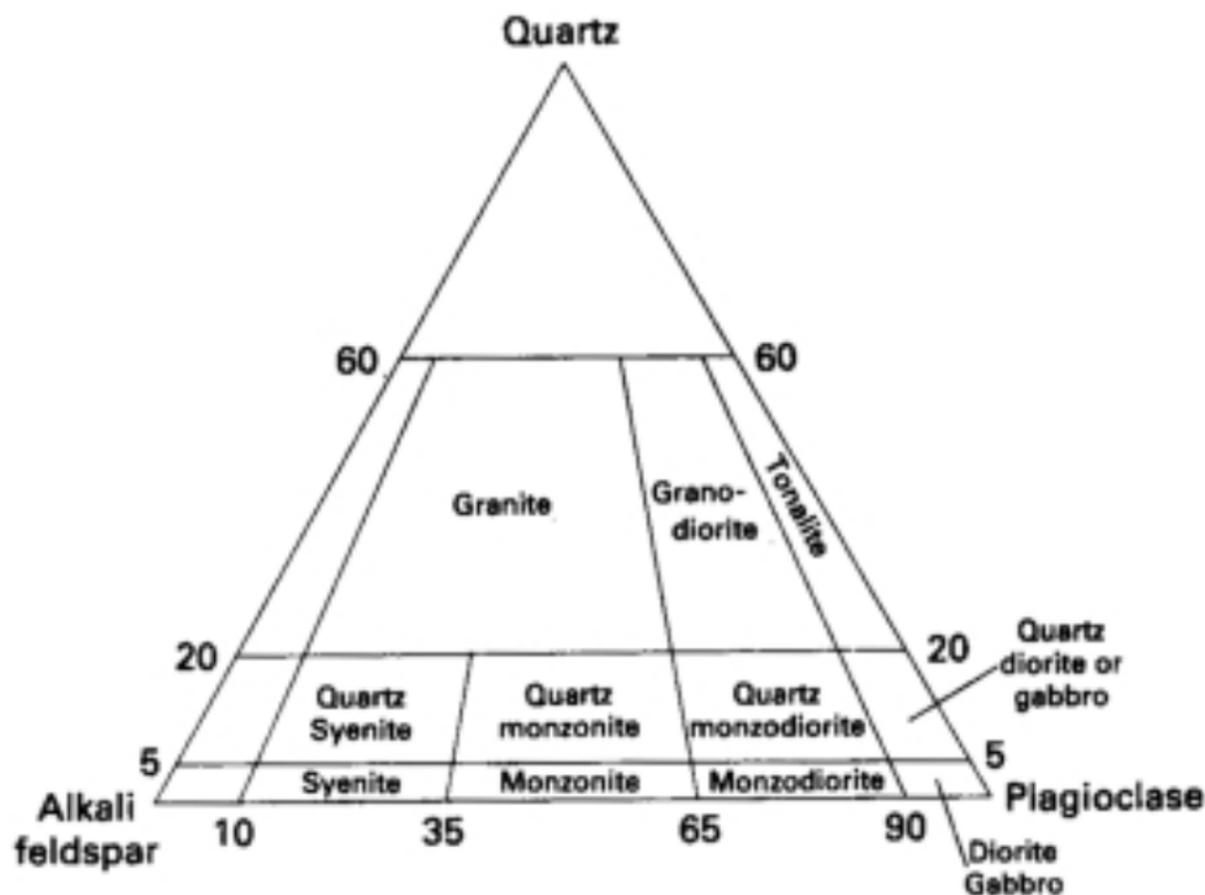


Figure 3.—Modal rock classification for plutonic rocks. After Streckeisen (1973).

we include in this unit crop out on the east side of Lake Chelan but are poorly exposed and so thoroughly penetrated by granite porphyry dikes that we saw them only in isolated outcrops.

The amphibolite and schist of Twentyfive Mile Creek grade southward into migmatite as the amounts of light-colored tonalite sills and dikes and swirled tonalitic gneiss increase. Schistose rocks of probable supracrustal derivation, especially schistose amphibolite similar to the Twentyfive Mile Creek unit, occur throughout the migmatite terrane.

A discordant U-Pb zircon analysis from biotite- quartz-oligoclase granofels in the younger gneissic rocks of the Holden area gives an older concordia intercept of 265 ± 15 m.y. if the younger intercept is assumed to be 60-90 m.y., the age bracket for the Late Cretaceous metamorphism (Mattinson, 1972, p. 3773). The granofels has been interpreted to be a metamorphosed keratophyre by C. A. Hopson (as reported in Mattinson, 1972, p. 3773) and thus the 265-m.y. (Permian) age is believed to represent the depositional age. We tentatively accept this Permian age as the protolithic age of the Twentyfive Mile Creek unit.

Chelan Complex of Hopson and Mattinson

Waters (1930, p. 58-102, 1938, p. 763-794) used the term Chelan batholith for all the granitoid rocks and migmatites of the Chelan Mountains-Lake Chelan area. Hopson and Mattinson (1971) and Mattinson (1972) referred many of these same rocks to the Chelan Complex. We include in the complex massive and gneissic tonalite and migmatite in the vicinity of the town of Chelan and along Lake Chelan, banded gneiss and banded migmatitic gneiss along the Entiat River, the distinctive Entiat pluton, and the possibly correlative hornblende tonalite gneiss of Antoine Creek. We exclude various younger Tertiary plutons, as well as large masses of strongly metamorphosed but relatively unmigmatized schist, such as those along Twentyfive Mile Creek (fig. 4).

Waters (1938) described in detail the feldspathization of supracrustal amphibolite and the production of migmatite in the Chelan area, a process he ascribed to contact metamorphism by massive of the igneous granitoid rocks of the Chelan batholith. Hopson (1955) proposed that the massive igneous rocks were derived by feldspathization and mobilization of amphibolite and schist during metamorphic processes. Later, Hopson and Mattinson (1971) emphasized that Late Cretaceous ultrametamorphism produced numerous leucotrochjemitic neosomes (light-colored tonalite) about 65-90 m.y. ago that were mobilized to a plastic mush by partial melting. Much of the mobilized material was derived from older igneous tonalite of the 220-m.y.-old (Triassic) Dumbell-Marblemount belt, which is on strike with the complex to the northwest. Building on the excellent observations of Hopson (1955) and the isotope work of Mattinson (1972), our work confirms that the ultrametamorphism led to melting but also that the massive tonalite plutons so formed in the Late Cretaceous were further metamorphosed in latest Cretaceous and earliest in Tertiary time.

Entiat Pluton-The Entiat pluton was first described by Waters (1930, p. 58-75, 1938, p. 767- 770), but was named by Hopson (1955, p. 118) who called it the Entiat facies of the Chelan batholith. The medium-grained

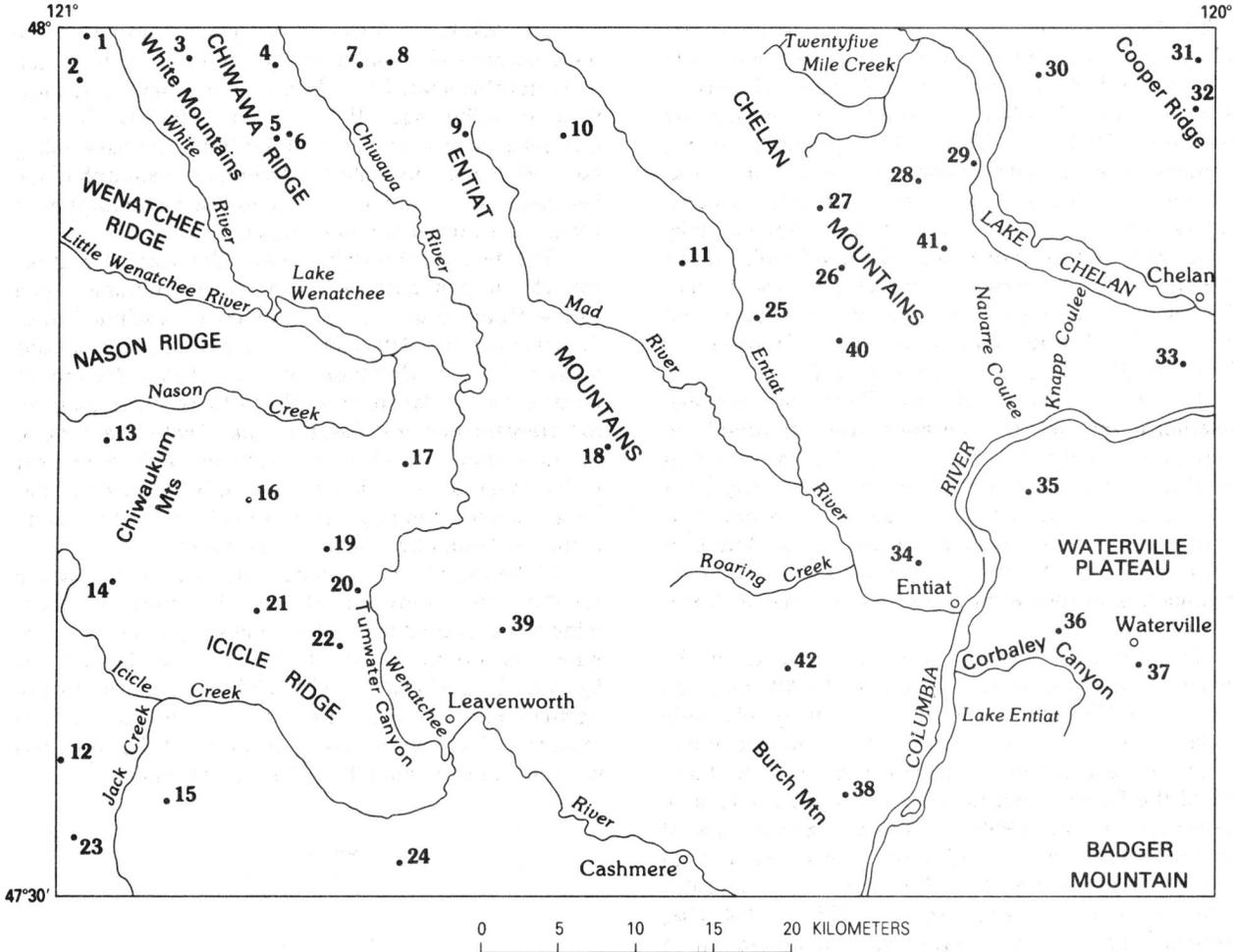


Figure 4.—Locations of obscure places in the Chelan quadrangle. Numbers indicate localities referred to in text.

granitoid rocks of the Entiat pluton extend along the Entiat Mountains from the Columbia River to the headwaters of the Entiat River. Its correlatives, the Seven Fingered Jack pluton (Cater and Wright, 1967; Cater and Crowder, 1967; Cater, 1982, p. 23-28) continue to Railroad Creek west of Holden. The Entiat pluton and its correlatives have a total length of about 80 km. In the Lucerne quadrangle directly on strike with our mapped Entiat, Cater and Wright (1967) have distinguished hornblende quartz diorite gneiss of the Seven Fingered Jack pluton from hornblende-biotite and biotite hornblende quartz diorite gneiss of the Entiat pluton to the east. We were unable to map the separate plutons within the Chelan quadrangle. Although the plutonic igneous origin is unmistakable in the texture and homogeneity of the pluton, we have not found evidence of intrusion in the Chelan quadrangle. The contact of the Entiat pluton with the heterogeneous schist and gneiss unit on the west side is marked by a strong but partially recrystallized cataclastic textures in the pluton as, for example, in Corbaley Canyon (fig. 4). Blastomylonitic micaceous quartzite is exposed along the middle part of the Mad River. An originally intrusive contact could have been obliterated by later shearing and metamorphism. Waters (1938, p. 768) and Cater (1982, p.25) ascribe the deformation along the western margin of the pluton to protoclasia, but it could represent post-intrusion deep-seated faulting and recrystallization.

At its southern end, the pluton is not intruded by the light-colored dikes that are common in the Swakane and Mad River terranes to the west (see Waters, 1932, p. 620), suggesting that the units were not adjacent when the dikes were intruded. However, to the north in the Mad Lake area [9]¹, numerous light-colored dikes and irregular bodies of tonalite do cut the pluton (see discussion of the migmatitic biotite alaskite gneiss of the Chelan Complex). On the east side, the pluton appears to grade into the migmatitic gneiss of the Chelan Complex. Hornblende tonalite gneiss identical to rocks of the pluton is a common constituent of the migmatite near the contact. Along the Entiat Mountains near the northern border of the quadrangle, the uniform tonalite gneiss of the pluton grades abruptly to banded, migmatitic hornblende gneiss rich in hornblende lenses and pods, a unit described as a contact complex by Cater and Crowder (1967) and Crowder (1959, p. 852-855) in the Holden quadrangle to the north. Waters (1930, p. 58; 1938, p. 767) considered the Entiat pluton to be part of the Chelan batholith (Chelan Complex of Hopson and Mattinson, 1971). Hopson (1955, p. 118-121) considered the Entiat to be a facies of the Chelan batholith, intrusive into hot and plastic migmatitic rocks near Chelan. Most of the rocks he studied are part of our mafic tonalite unit of the Chelan Complex. The age of the Entiat as discussed below suggests that it is indeed a facies of the massive tonalite in the Chelan Complex and represents metamorphosed anatectic tonalite derived from the same rocks as the tonalite near Chelan (see Hopson and Mattinson, 1971). However, the Entiat pluton's uniformity, higher K-feldspar content, elongate shape, and abrupt transition to migmatite distinguish it from the tonalite of the Chelan area. We include in the Entiat pluton a body of fine grained two-pyroxene gabbro in the Potato [40] and Stormy Creeks [25] area. The clinopyroxene hypersthene gabbro grades into tonalite typical of the Entiat pluton where the pyroxenes have been replaced by hornblende and biotite and where quartz becomes a constituent of the groundmass. Wide northeast southwest-trending zones of granulated and chloritized rocks cut the gabbro in Potato Creek. This body within the Entiat pluton might be a relict inclusion of (1) Triassic igneous rocks (the protolith of the Entiat pluton) or (2) a separate mafic Late Cretaceous pluton.

Banded migmatitic tonalite gneiss and mafic amphibolite-Banded migmatitic tonalite gneiss, rich in mafic amphibolite, appears to be closely associated with the Entiat pluton and gradational into it. The banded migmatite is probably continuous with migmatite of the Chelan Complex in the area of Stormy Creek [25]. Crowder (1959, p. 852-855) described in detail complexly mixed hornblende gneiss and mafic amphibolite east of the Entiat pluton just north of the quadrangle as "replacement and recrystallization migmatite." He ascribed the formation of the migmatite to extreme metamorphism with progressive granitization and anatexis changing hornblende schist into massive quartz diorite now represented by the

¹Numbers in brackets after place names refer to locations on figure 4, an aid to finding obscure places on the geologic map.

Entiat and Seven Fingered Jack plutons. Cater (1982, p. 90-91) considered the migmatite a contact zone (see also Cater and Wright, 1967) related to igneous intrusion of the pluton at catazonal depths where the distinction between metamorphism and melted rocks is not well defined (see also Buddington, 1959, p. 714). Rocks in the banded migmatite unit examined by us in the Chelan quadrangle do not display relict igneous outcrops at textures in thin section except for rare, faint euhedral oscillatory zoning in plagioclase crystals in hornblende tonalite gneiss. Except for a pendant of mostly schistose amphibolite and hornblende-biotite schist northeast of the lower Mad River, we found no relict metasedimentary or metavolcanic material in the unit.

Banded hornblende and biotite gneiss- Heterogeneous hornblende gneiss, biotite gneiss, and gneissic amphibolite in the Chelan quadrangle are continuous with rocks mapped as hornblende schist and gneiss (younger gneissic rocks of the Holden area) in mafic the Lucerne quadrangle (Cater and Wright, 1967). In a Banded hornblende-biotite gneiss grades into the banded migmatitic tonalite gneiss and mafic amphibolite unit associated with the Entiat pluton on the west and is sharply intruded by the Eocene Duncan Hill pluton on the east. The lack of supracrustal material in the banded hornblende and biotite gneiss unit within the Chelan quadrangle makes correlation with the supracrustal rocks of the Holden area doubtful (see also the discussion of the Mad River terrane). The banded gneiss and much of the banded migmatitic tonalite gneiss to the west could well be derived from plutonic igneous material of the Triassic Dumbell plutons on strike to the north (Cater and Wright, 1967), as indicated by Hopson and Mattinson (1971) and Mattinson (1972, p. 3778) for similar rocks in the Chelan Complex. The protolith of the banded gneiss is pre-Late Cretaceous, but it has been thoroughly transformed by the Late Cretaceous metamorphism and (or) earlier metamorphisms and we include it in the Chelan Complex

Migmatite-Migmatite in the Chelan Complex is characterized by extreme heterogeneity. It consists predominantly of dark-colored hornblende-rich and (or) biotite-rich gneiss, amphibolite, microdiorite, and light-colored biotite and (or) hornblende tonalite. These rocks are interlayered, mosaicked, commonly swirled, and generally mixed on all scales. Hopson (1955, p. 50- 106) describes them well. Although Hopson and Mattinson

(1971) emphasized derivation of the migmatite from the Triassic plutons exposed to the north, partial derivation from supracrustal rocks is suggested by amphibolite, hornblende schist, and especially marble on Chelan Butte [33] (Waters, 1938, p. 765; Hopson, 1955, p. 11) and by the probable gradational relations with amphibolite, marble, and quartzitic schist in the Twentyfive Mile Creek area and along Cooper Ridge (fig. 4). In general we concur with Hopson and Mattinson (1971) and Mattinson (1970; p. 3774; 1972) that the migmatite formed 60-90 m.y. ago and that older U-Th- Pb ages represent partially degraded ages of protolithic zircons.

Amphibolite and hornblendite migmatite- Dark rocks of amphibolite and hornblendite migmatite are characterized by abundant layers and pods of hornblendite. The rocks form spectacular outcrops but are otherwise similar to the migmatite unit. Contacts are gradational and approximate.

Migmatitic biotite alaskite gneiss-In outcrops at the head of Tommy Creek [10] migmatitic alaskite gneiss interfingers with and grades into the gneissic tonalite of the Entiat pluton. At the southern end of the alaskite gneiss body, a profusion of light-colored dikes and sill-like bodies of biotite alaskite gneiss cut the Entiat, but gradational contacts are rare. Pods of hornblendite are also common in the southern area. The migmatitic biotite alaskite gneiss differs from the migmatite of the Chelan area by its greater K-feldspar content, ranging from 5 to 30 percent.

Tonalite and mafic tonalite-Tonalite and mafic tonalite are characterized by relict igneous textures in a mostly crystalloblastic fabric; they appear to be metamorphosed igneous rocks. Locally, metamorphic epidote replaces plagioclase, the latter exhibiting relict euhedral oscillatory zoning. Tonalite of the Chelan Complex is poor in mafic minerals and has little or no K-feldspar. The tonalite grades into migmatite. Much material in the migmatite is similar to tonalite but is present in smaller amounts and mixed with more mafic rocks. We distinguished our tonalite and mafic tonalite units from migmatite on the basis of overall homogeneity of the tonalite over areas measured in hundreds of square meters. The tonalite contains abundant schlieren but few large mafic clots or thick layers; it is massive or gneissic and where gneissic may be swirled. It is locally cut by lighter colored tonalite, alaskite, or pegmatite dikes. Our tonalite unit includes much of Hopson's (1955, p. 19-38, pl. 2) leucocratic quartz diorite unit. A mass of mafic tonalite stretching from Knapp Coulee southwestward to the Columbia River includes rocks mapped by Hopson (1955, pl. 2) as biotite hornblende quartz diorite of the Entiat facies and mesocratic quartz diorite. Based on divergent orientation of structural elements in the tonalite compared to the host rock (such as lineation), the abrupt nature of some contacts and gradational nature of others, and the similarity of mineral facies between intrusion and host rocks, Hopson (1955, p. 119-121) concluded that the mafic tonalite intruded the migmatite while the latter was hot and plastic.

Hornblende tonalite gneiss of Antoine [31]-A relatively uniform hornblende gneiss, locally highly sheared and strongly chloritized, crops out in the extreme northeastern corner of the quadrangle. Elongate aggregates of hornblende give the rock a pronounced lineation. Healed cataclastic zones and fibrous aggregates or

porphyroblasts of hornblende indicate that the gneiss is a metamorphosed coarsely crystalline igneous rock, although we found no relict igneous textures or structures. We observed no contacts, but in general the rock resembles the tonalite gneiss of the Entiat pluton, and we thus show it as the same age.

Radiometric ages of the Chelan Complex

Uranium-lead ages of zircons and sphenes from the Chelan Complex are predominantly discordant, revealing both the Late Cretaceous plutonic and metamorphic events and the presence of older lead the protoliths. Two samples with almost concordant ages from the center of the Entiat pluton (table 1, No. 49) and from the mafic tonalite unit east of Knapp Coulee (No. 38) record most straightforwardly reystallization of zircon about 75 to 85 m.y. ago. Because the plutons at these localities are composed of massive, uniform tonalite displaying relict igneous textures, they probably were formed from true igneous melts in the Late Cretaceous. We would argue that here especially the mineralogy and texture of the original protoiths of the plutons are long gone, and also that the superimposed effects of postcrystallization metamorphism are minimal.

An almost concordant U-Pb zircon age of 185 m.y. Jurassic) (table 1, No. 51) from a biotite-hornblende gneiss in the banded migmatite adjacent to the contact with the Entiat pluton on Tyee Mountain was interpreted by Hopson and Mattinson (1971) and Mattinson (1972, p. 3778 and written commun.) to represent the minimum age of the protolith of the banded migmatitic tonalite gneiss that they believed was intruded by the Entiat pluton. The dated biotite-hornblende gneiss, by this interpretation, is metamorphosed igneous material correlative with the Triassic Dumbell plutons on strike to the northwest.

In addition to the Tyee Mountain age, mildly discordant ages ranging from 99 to 127 m.y. of zircon from light-colored tonalite in migmatite (table 1, Nos. 42-44; Mattinson, 1972, p. 3774) also appear to reflect the influence of pre-Cretaceous protoliths. The most probable candidates for correlative units of these protoliths are the Triassic plutons exposed to the north and the Permian(?) metasedimentary and metavolcanic rocks of the Twentyfive Mile Creek unit. Zircons with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 175 m.y. from tonalite along Lake (No. 40; Mattinson, 1972, p. 3778) and $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 178 and 260 m.y. from the flaser gneiss zone of the Entiat pluton (No. 50) give evidence for the retention of older lead even in parts of the anatectic melts. The results on the Entiat flaser gneiss a minimum Permian age for detrital or primary zircon from the wall rocks intruded by, or imbricated with, the pluton--in this case, the heterogeneous schist and gneiss unit of the Mad River terrane. However, a close look at the discordant systematics suggests a considerably older--in all likelihood Precambrian--age for the inherited component of zircon. The older zircons could have been picked up from the Swakane Biotite Gneiss underlying the heterogeneous schist and gneiss unit.

Based on lithologic similarity, much of the tonalite of the Chelan Complex is thought to have formed in the Cretaceous from melts that were derived by mobilization or anatexis of older rocks as indicated by Hopson and Mattinson (1971). The gradation of the more massive plutons into migmatitic terrane and the spectrum of U-Pb

zircon ages representing mixtures of inherited and new components of this refractory mineral strongly support such an interpretation.

Metamorphic overprinting of the plutons must have occurred contemporaneously with or immediately after their emplacement if we accept the 70 to 86 m.y. U-Pb sphene ages (table 1, Nos. 42-44; Mattinson, 1972, p. 3774) and the K-Ar hornblende and biotite ages as old as 84 m.y. (No. 41) and 74 m.y. (No. 39), respectively, from the Chelan Complex as recording that event. From these maximum values for the K-Ar ages, the hornblende ranges downward to 60 m.y. and the biotite downward to 56 m.y., and where both minerals have been analyzed from a single sample, the hornblende consistently yields the greater age. This pattern is suggestive of differential uplift and cooling, with some parts of the terrane rebounding rapidly after plutonism and metamorphism, while other parts remained relatively hot for another 15 to 20 m.y. Finally, fission track ages of zircon and apatite tend to be displaced downward another 5 to 10 m.y. from the youngest K-Ar ages. The relatively tight grouping of apatite fission track ages around 50 m.y. would imply that by early Eocene time the entire terrane had cooled to 100°C. One anomalously low zircon fission track age of 33 m.y. (table 1, No. 47) possibly reflects some local, fault induced or volcanic heating event, perhaps associated with the nearby Oligocene volcanic rocks of CHIKAMIN Creek [8].

Pyroxene-hornblende gabbro and quartz gabbro

A small body of pyroxene-hornblende gabbro in the valley of upper Twentyfive Mile Creek (fig. 4) has fine-grained margins and satellitic dikes which clearly intrude biotite tonalite of the Chelan Complex. Several features suggest that metamorphism followed intrusion of the gabbro. The dikes are slightly schistose and radial clusters of hornblende have grown out into the tonalite from the walls of the dikes. Euhedral, reddish-brown hornblende commonly zoned outward to actinolitic hornblende, and plagioclase zoned from euhedral calcic cores to xenoblastic sodic rims, both indicate modification of the original igneous texture by metamorphism. Pyroxene-hornblende diorite with similar distinctive mineralogy and texture crops out in upper Potato Creek [40], in First Creek [41], and in scattered outcrops in the migmatite north of Antilon Lake [30]. Dikes with similar mineralogy but lacking foliation are scattered throughout the Chelan Mountains terrane. One dike that cuts tonalite of the Chelan Complex in Washington Creek [32] yields a K-Ar hornblende age of about 48 m.y. (table 1, No. 32), which is probably a minimum age of intrusion.

SWAKANE TERRANE

In earlier reports we (Tabor and others, 1980, 1982c) included the Swakane Biotite Gneiss and the 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 and retain the Swakane Biotite Gneiss as the sole component of the Swakane terrane (Tabor and others, in press).

Swakane Biotite Gneiss and amphibolite and hornblende schist

The Swakane Biotite Gneiss, named for exposures in the canyon of Swakane Creek (38] is exposed for at least 100 km northwestward from Wenatchee (Waters, 1932, p. 606; Chappell, 1936a, pl. 1; Page, 1939a, p. 8-16; Cater and Crowder, 1967). Waters' (1930, 1932) original description of what he mapped as the Swakane Gneiss included rocks that we have herein mapped as the heterogeneous schist and gneiss unit in the Mad River terrane. Crowder and others (1966) restricted the name to the biotite gneiss which is the predominant rock type, and we follow their usage.

North of the quadrangle, Cater and Wright (1967) and Cater (1982, p. 7) considered a biotite gneiss along Lake Chelan to be correlative with the Swakane Biotite Gneiss and to be structurally overlain by the younger gneissic rocks of the Holden area in the Chelan Mountains terrane. Their correlative biotite gneiss is continuous with rocks to the north mapped as Skagit Gneiss of Misch (1966). Important to their argument is the assumption that the younger gneissic rocks of the Holden area are equivalent to the heterogeneous gneiss and schist unit of the Mad River terrane, a correlation that we question (see below) although the latter unit does appear to structurally overlie the Swakane Biotite Gneiss. Although the Skagit Gneiss superficially resembles the Swakane, it commonly contains numerous layers of schist and amphibolite of probable supracrustal origin, and, in more uniform parts, the Skagit contains considerably more amphibole, mostly hornblende (see Tabor and others, in press, for further discussion). On the basis of overall composition, we think that the Skagit Gneiss and Swakane Biotite Gneiss are not correlative.

The Swakane is a remarkably uniform granofelsic gneiss, although it does contain rare thin layers of hornblende schist, amphibolite, and marble. Waters (1932, p. 616) and later workers (Chappell, 1936a, p. 48-49; Page, 1939a, 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. The uncertain origin makes interpretation of the ages of its protolith difficult. Based on a concordia plot of highly discordant U-Pb zircon ages (table 1, No. 66), Mattinson (1972, p. 3773) favored the hypothesis that this mineral in the Swakane originally crystallized at least 1,600 million years ago and was metamorphosed about 415 m.y. ago and again 60 to 90 m.y. ago. If the original rock was volcanic, a Middle Proterozoic or earlier age should represent its time of formation. Many of the zircons appear rounded, however, suggesting a detrital origin. Hence, the protolith could be sedimentary, and the Precambrian age of the zircons would represent only a maximum age.

We made a second attempt to establish a stratigraphic age for the Swakane by analyzing zircon from a garnet amphibolite layer in the Swakane (table 1, No. 67) thought to be originally a mafic volcanic rock. However, again the zircon is, in part, rounded and probably represents a mixed population of detrital and metamorphic components. If we pool our results on two size-fractions of zircon from the garnet amphibolite with those of Mattinson (1972) on the biotite gneiss (table 1, No. 66), the general chronologic interpretation remains the same as

for the biotite gneiss, although we find the evidence from these data to be weak for Mattinson's proposed Paleozoic metamorphism.

Rubidium-Strontium 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 (written 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.

Uranium-lead ages of zircons from pegmatite gneiss in the Swakane along the Columbia River (table 1, No. 65) and from light-colored tonalite north of the quadrangle indicate that the latest episode of regional metamorphism, which crystallized or recrystallized the Swakane, took place in the Late Cretaceous (Mattinson, 1972, p. 3779).

MAD RIVER TERRANE

The Mad River terrane structurally overlies the Swakane terrane. We include in this terrane the heterogeneous schist and gneiss unit and the probably correlative rocks of the Napeequa River area. The terrane is named for the Mad River, a major tributary of the Entiat River (fig. 4).

The heterogeneous schist and gneiss unit is continuous with metasedimentary and metavolcanic rocks in the Holden quadrangle to the north which Cater and Crowder (1967) and Cater and Wright (1967) correlated with the younger gneissic rocks of the Holden area. The former rocks are not continuous with the Holden rocks which are continuous with our Twentyfive Mile Creek unit in the Chelan Mountains terrane. For reasons more fully explained in Tabor and others (in press), we do not consider the correlation of the metasedimentary and metavolcanic rocks (our heterogeneous schist and gneiss unit) with the younger gneissic rocks of the Holden area well-established.

Rocks of the Napeequa River area [3], most extensively exposed in the Holden quadrangle to the north (Cater and Crowder, 1967), are continuous with schist and gneissic amphibolite on Chiwawa Ridge in the Chelan quadrangle. Although most rocks of the Mad River terrane are separated from the Nason terrane and Ingalls Tectonic Complex by the Chiwaukum graben in the Chelan quadrangle (fig. 1), rocks of the Napeequa River area on Chiwawa Ridge and Mount David [21 and northwest of the quadrangle are adjacent to the Chiwaukum Schist of the Nason terrane. Within the Napeequa River unit, metagabbro, metadiorite, and metadiorite gneiss are strongly mylonitic and imbricated with the Chiwaukum Schist to the south, and we interpret this contact as a fault.

On the east, rocks of the Mad River terrane are mylonitic adjacent to flaser gneiss forming the margin of the Entiat pluton of the Chelan Mountains terrane. The exact nature of the original contact is not clear, but the Entiat pluton could have intruded the Mad River terrane prior to the strong deep-seated shearing now evident along the contact.

Heterogeneous schist and gneiss

Overlying the Swakane Biotite Gneiss are hornblende schist, schistose amphibolite, micaceous quartz schist, micaceous quartzite, biotite gneiss, and rare calc-silicate schist and marble. In contrast to the broadly folded, uniform foliation in the Swakane, foliation in these rocks is isoclinally folded on an outcrop scale. The exact nature of the contact with the Swakane is enigmatic. The Swakane has rare layers of quartzitic schist, schistose amphibolite, calc silicate schist, and marble, some of which form thin beds traceable for kilometers north of the quadrangle (Cater and Crowder, 1967; Cater, 1982), and the heterogeneous schist and gneiss unit locally contains thin layers of biotite gneiss which resemble the Swakane. The abrupt contact between the two units appears to be conformable to the uniform foliation in the Swakane, but not to the folding in the schist, indicating that the contact represents a fault.

Small masses of blastoporphyratic biotite gneiss and flaser gneiss of probable igneous origin, siliceous metaporphry dikes, and small pods of ultramafic and talc-tremolite schist (locally mapped) are scattered throughout the heterogeneous schist and gneiss unit. Neither the metamorphosed intrusive rocks nor the ultramafite occur in the underlying Swakane, further evidence that the contact is a fault.

In the Holden quadrangle, the heterogeneous schist and gneiss unit is separated from the correlative rocks of the Napeequa River area to the west by an uplifted block of the Swakane Biotite Gneiss. Especially common to the correlative units are micaceous quartzite, fine-grained hornblende schist, marble, metaporphry dikes, and flaser gneiss bodies. Also common to the two units are metamorphosed ultramafic rocks. Hornblende-zoisite gneiss, probably derived from igneous masses like the protolith of the metagabbro, metadiorite, and metadiorite gneiss unit described below, is common in the rocks of the Napeequa River area, but we have not found it in the heterogeneous schist and gneiss unit east of the Chiwaukum graben.

Light-colored gneiss--Within the heterogeneous schist and gneiss unit are many unmapped masses of light-colored biotite gneiss of varying size and texture. The mapped bodies consist of heterogeneous granitoid biotite gneiss, varying from slightly gneissoid to flaseroid. These rocks are thoroughly metamorphosed with heteroblastic textures, although in some samples rounded feldspar in a mesh of mafic minerals suggests shearing and recrystallization of a previously coarse grained crystalline rock, probably of igneous granodiorite origin. We have found no evidence of intrusion.

Much of the biotite gneiss of the largest body, which stretches along the lower Entiat River, was included by Waters (1930, 1938) in the Chelan batholith and considered by him to intrude the heterogeneous schist and gneiss unit (his Swakane Gneiss, 1930, p. 63). He felt other parts were igneous bodies which were once overlain unconformably by the Swakane protolith and later metamorphosed (1932, p. 616). Laravie (1976, p. 54) concluded that the granitoid biotite gneiss bodies along the west side of the Entiat Range were metamorphosed igneous intrusions that had been emplaced in the protolith of the schist, but while we concur with his conclusion the evidence is not definitive.

Ultramafite--Small masses of ultramafite, mostly serpentine and serpentinized peridotite, crop out sporadically throughout the heterogeneous schist and gneiss unit but are most prominent in the sheared rocks between two major strands of the Entiat fault. Ultramafite does not crop out in the Swakane Biotite Gneiss. We have mapped small pods of talc-tremolite rock as ultramafite, although Waters (1932, p. 621) suggested that they could also have been impure dolomite.

Rocks of the Napeequa River area

Mica schist and gneissic amphibolite mapped with the rocks of the Napeequa River area in the Holden quadrangle (Cater and Crowder, 1967) are traceable into the Chelan quadrangle on Chiwawa Ridge (fig. 4) and west of the Napeequa River [3]. On Chiwawa Ridge the rocks of this unit are mostly fine-grained amphibolite and zoisite-amphibolite gneiss that may well be derived from the protolith of the metagabbro, metadiorite, and metadiorite gneiss unit described below. Also abundant are numerous metaporphyry dikes. Just west of the Napeequa River the unit is mostly well-recrystallized fine- and medium-grained layered amphibolite.

Schist, gneissic amphibolite, and mafic breccia—On the southeast ridges of Mount David [2], micaceous quartzite and mica schist with thick pods of mafic breccia and some amphibolite are exposed in a reentrant of the Ten Peak pluton. Where less deformed, the mafic breccia looks like volcanic breccia. Well-rounded marble clasts suggest a sedimentary origin. Although a few percent lower in silica than most mafic basalt (see Ort and Tabor, 1985, samples RW7 465A and B-79), the hornblende clasts and amphibolite matrix of the breccia are close in composition to olivine-bearing alkali basalt. Cater and Crowder (1967) described similar mafic rocks as contact breccias of the Ten Peak pluton.

Although the micaceous quartzites are most like rocks in the Napeequa River area and thus included in the Mad River terrane, their affinity with any of the tectono-stratigraphic terranes described here is uncertain. If we interpret the micaceous quartzite to be derived from chert and the mafic breccia from mafic volcanic rocks, they have some similarity to rocks in the Ingalls Tectonic Complex and could be a small remnant of overthrust Ingalls, analogous to the rocks at Windy Pass [15] to the south.

The micaceous quartzite and mafic breccia have little in common with the Chiwaukum Schist (Nason terrane), which borders them on the south. We infer that they are in fault contact with the Chiwaukum, but, except for some local mylonitization near the contact, a continuous fault is not obvious. The varied lithologies, including ultramafite pods, suggest that these rocks, and the schist of Crook Mountain (herein included in the Nason terrane), may be part of a wide zone of imbrication along the terrane junction.

Metagabbro, metadiorite, and metadiorite gneiss—Around the south end of the Ten Peak pluton, metagabbro, metadiorite, and metadiorite gneiss have undergone a complex history of deformation and metamorphism. Tabor and others (1980) once included the metagabbro and its derivatives in the Ten Peak pluton as

did Van Diver (1964, p. 117-124; 1967, p. 135), who considered them to be part of a strongly deformed and retrogressed margin of the pluton. The metagabbro and its associated rocks are lithologically distinct from the Ten Peak, however, and for reasons described below, we think they may be older.

East of Raging Creek [6] the metagabbro grades from massive to slightly gneissic granitoid rocks, made up of hornblende prisms in a matrix of mainly zoisite, to highly foliate rocks: zoisite-amphibolite gneiss, hornblende-zoisite gneiss, and schist. The progressive change from massive rocks with relict igneous texture to gneissic or schistose rocks with a completely metamorphic texture is readily observed in outcrops and thin section.

Crosscutting foliation indicates multiple deformation. Massive to gneissic zoisite amphibolite (metagabbro) grades westward across Raging Creek into alternating layers of zoisite rock (meta-anorthosite?) and zoisite amphibolite gneiss, both with northwest-trending foliation. A second episode of deformation has produced an east-west foliation cutting both the massive metagabbro and the interlayered zoisite rock and zoisite amphibolite gneiss. The east-west foliation roughly parallels the margin of the Ten Peak pluton. As proposed by Van Diver (1967), this foliation may be related to late metamorphic movement (emplacement?) of the Ten Peak pluton, and, as indicated by isotopic data, may have occurred in the Eocene (see below). However, the initial metamorphism of the gabbro, that which produced the massive hornblende-zoisite rock and the northwest layering and foliation at a marked angle to the Ten Peak contact, preceded the emplacement of the Ten Peak. Considerable zoisite in streaky amphibolite in the rocks of the Napeequa River area north of the Chelan quadrangle, well removed from the Ten Peak pluton, supports the contention that the metagabbro and its derivatives are older than the Ten Peak pluton.

Although the margin of the Ten Peak pluton is streaked with mafic inclusions, we found it difficult to tell mafic varieties of the Ten Peak from the more gneissic varieties of metagabbro. We assume that streaky inclusions of hornblende-zoisite gneiss within the Ten Peak are xenoliths of metagabbro.

The southern margin of the metagabbro is the most intensely deformed part of the body, and it is locally imbricated with fine-grained schist of the Chiwaukum Schist. We interpret this contact to be a metamorphosed fault separating the Mad River terrane from the Nason terrane.

Blastomylonitic amphibolite schist—Along the steep hillside above the Chiwawa River [4], north northwesterly foliation in the rocks of the Napeequa River area is highly pronounced; the rocks are blastomylonites with a strong planar cleavage. Talc schist and ultramafite pods as well as metaporphyry dikes occur in the rocks. Some chlorite-clinozoisite actinolite schist and zoisite-muscovite schist in this unit are thoroughly recrystallized and could be of metasedimentary origin. The zone follows here the trend of the Leavenworth fault, which bounds the Tertiary Chumstick Formation to the south, and slickensides and low-grade alteration indicate that the blastomylonite is in a zone of continued but more brittle movement.

Radiometric ages of the Mad River terrane

Concordant to moderately discordant U-Th-Pb zircon ages obtained for two adjacent samples (table 1, No. 73) of light-colored gneiss collected along the Columbia River show systematic increase in $^{207}\text{Pb}/^{206}\text{Pb}$ ages with increasing grain size while the U-Th-Pb ages remain almost constant; this is a pattern of isotopic behavior very similar to that of the Entiat pluton (Nos. 49 and 50). As was our interpretation for that case, the light-colored granodioritic gneiss appears to contain a mixture of newly crystallized Late Cretaceous and inherited and probable Paleozoic or older zircon.

Uranium-thorium-lead analysis of zircons separated from a zoisite amphibolite gneiss from the strongly foliate margin of the metagabbro (table 1, No. 68) does not clarify the protolithic age. Although the $^{207}\text{Pb}/^{206}\text{Pb}$ ages (88-100 m.y.) seem to indicate that the metagabbro could be related to the Late Cretaceous Ten Peak pluton, the U-Pb and Th-Pb ages (45-53 M.Y.) demand considerable recrystallization during a Tertiary event. Apparently some component of the zircon was originally considerably older but had its $^{207}\text{Pb}/^{206}\text{Pb}$ -age fortuitously reduced to about 90 m.y. We do not know what young metamorphic event (50 m.y. or less) has affected these rocks, but the gneissic tonalite of the northern end of the Eocene Duncan Hill pluton, lying almost 20 km to the northeast (see below), also testifies to a Tertiary dynamothermal metamorphic event.

Potassium-argon ages of minerals from rocks of the Mad River terrane range from about 50 to 68 m.y. (table 1, Nos. 68-73), a range not unlike K-Ar ages in the adjoining Chelan Mountains terrane. Roughly concordant K-Ar ages of hornblende and muscovite of 55 and 52 m.y. (No. 72) from zoisite amphibolite in the rocks of the Napeequa River area exposed north of the quadrangle may reflect the localized Tertiary metamorphic event that recrystallized zircon in the metagabbro, metadiorite, and metadiorite gneiss unit described above. K-Ar ages of hornblende and biotite from the Ten Peak pluton of 93 and 77 m.y., respectively (table 1, No. 55) are similar to K-Ar ages of the other Late Cretaceous synkinematic plutons, but nonetheless are remarkably unaffected by the Eocene event seen in the nearby metagabbro. The remaining K-Ar ages of the Mad River terrane may reflect waning regional metamorphism and(or) uplift.

In summary, isotopic ages from rocks of the Mad River terrane reflect mostly Late Cretaceous and early Tertiary metamorphism and uplift. We infer a Paleozoic or older protolithic age for the metasedimentary and metavolcanic rocks from the old zircon components of the orthogneisses of the terrane.

INGALLS TERRANE

Most of the Ingalls Tectonic Complex crops out south of the Chelan quadrangle (see Tabor and others, 1982a) in a wide belt around the southern end of the Mount Stuart batholith. The most abundant rocks in that belt in general, and in the Chelan quadrangle in particular, are serpentinite, serpentized peridotite, and metaperidotite, but the belt also includes large and small tectonic lenses and wedges of metamorphosed sedimentary rocks, mafic and intermediate volcanic rocks, diabase, and gabbro.

Within the unit south of the quadrangle, flysch-type sandstone and argillite, radiolarian chert, pillow basalt, and ultramafic rocks are stratigraphically associated and tectonically intermixed; the assemblage is an ophiolite

complex (Hopson and Mattinson, 1973; Miller, 1977; Miller and Frost, 1977, p. 287; Miller, 1980b). Southwick (1974) proposed that the unit consists of ophiolite and island-arc materials juxtaposed during subduction. Cowen and Miller (1980) suggested that extensive serpentinized shear zones in the unit represent parts of an oceanic fracture zone.

The age of gabbro in the Ingalls is Late Jurassic according to a U-Pb date on zircon (Southwick, 1974, p. 391). Chert from south of the Chelan quadrangle contains radiolarians restricted to the Late Jurassic (E. A. Pessagno, Jr., oral commun., 1977; Tabor and others, 1982a). We accept the Late Jurassic age for most components of the complex, although it could contain tectonic slices of other ages. It was tectonically emplaced prior to intrusion of the Late Cretaceous Mount Stuart batholith.

The metamorphic grade of the Ingalls Complex varies from greenschist and prehnite-pumpellyite facies in rocks south and west of the Chelan quadrangle to pyroxene hornfels facies in rocks adjacent to the Mount Stuart batholith. The mostly static thermal metamorphism caused by the intrusion is clearly printed on an older schistose fabric. However, rocks in the southwest corner of the quadrangle, in the vicinity of Highchair Mountain [12], are less schistose and retain more original structures than rocks to the northeast at Windy Pass [15] and on the eastern margin of the batholith. We include fine-grained amphibolite, mica schist, meta-quartz-diorite, and calc-silicate rocks at Windy Pass in the Ingalls Tectonic Complex, although Miller (1980b, p. 314) concluded that the protoliths of these rocks, especially the calc-silicate rocks and metaquartz-diorite, were not to be found in the Ingalls and that they were imbricate slices of some unknown unit.

East of the Stuart Range [24], low-grade rocks, including phyllite, greenstone, and metagabbro of the Ingalls Tectonic Complex are faulted against amphibolite facies rocks adjacent to the batholith.

Schist mapped in the Ingalls Tectonic Complex is generally very fine grained, more amphibolitic than that of the Chiwaukum Schist, and characteristically has a sugary texture which, when viewed in thin section, is granoblastic. Similar rocks, however, crop out locally in the Chiwaukum Schist, especially on the ridges north of Icicle Creek (fig. 4) where they are associated with ultramafic rocks.

Miller (1980a; 1980b, p. 390-404) has suggested that the Ingalls and the Chiwaukum 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 metamorphism. On the basis of Miller's interpretation we show the contact between the two major units as a thrust fault.

We have included most small ultramafite bodies surrounded by the Chiwaukum Schist in the Chiwaukum, but they are indistinguishable from ultramafite of the Ingalls Tectonic Complex and could represent a wide zone of imbrication of the two units. According to detailed structural studies by B. R. Frost (written commun., 1977), the largest masses of ultramafite south of Hatchery Creek [20] and on the east end of Icicle Ridge appear to be metamorphosed peridotite of the Ingalls Tectonic Complex in fault contact with the Chiwaukum Schist.

The probable occurrence of ultramafite of the Ingalls Tectonic Complex as tectonic slivers and (or) pre-metamorphic serpentinite slide blocks (see below) within the Chiwaukum Schist suggests that the two units may

have been adjacent prior to initial metamorphism of the Chiwaukum. Miller (1980b, p. 404) thought that the Ingalls had been obducted onto the Chiwaukum Schist.

NASON TERRANE

The Chiwaukum Schist and banded gneiss derived from the Chiwaukum Schist by metamorphic and igneous processes are the major units of the Nason terrane. Intruded into the terrane are the Late Cretaceous Ten Peak pluton, the Dirty Face pluton, and the Mount Stuart batholith. We have named the terrane for Nason Creek (near U.S. Route 2) where rocks typifying the terrane are exposed. The terrane is bounded by the Leavenworth fault on the east and, west of the quadrangle, the Straight Creek fault zone (see Yeats and others, 1977, p. 267-171; and Tabor and others, 1982b). The main constituent of the Nason terrane, the Chiwaukum Schist, is in probable thrust contact with the Ingalls Tectonic Complex as previously described.

Pre-Eocene clockwise rotation and (or) considerable northward translation of all or part of the Nason terrane and of the Ingalls Tectonic Complex has been suggested by Beck and Nason (1971) and Beck and others (1981), on the basis of anomalous Late Cretaceous magnetic pole directions in the Mount Stuart batholith. Post-intrusion, in situ tilt of the batholith can eliminate some of the pole discordance, but not all of it.

The Chiwaukum Schist and the associated banded gneiss unit appear to be derived from predominantly pelitic rocks with only a small amount of mafic volcanic rocks. The Chiwaukum has undergone at least two dynamothermal metamorphic episodes. An early regional metamorphic episode was in the almandine amphibolite facies (Plummer, 1980, p. 386). In pelitic rocks of the Chiwaukum Schist a regional gradient progresses from garnet zone through staurolite to kyanite and back to at least staurolite and probably subtauroilite from south-southwest to north-northeast across the quadrangle. The second episode of metamorphism was also in the almandine amphibolite facies but graded from cordierite subfacies in pelitic rocks of the Ingalls Tectonic Complex on the south to sillimanite subfacies in the Chiwaukum Schist to the north and locally associated with the late synkinematic Mount Stuart batholith (see Plummer, 1980, p. 386; Kaneda, 1980).

The overall structure in the Nason terrane (shown in cross section A-A', map sheet) reflects a complex antiform on the south pushed up (?) and intruded by the Mount Stuart batholith, a synform in the Chiwaukum Mountains area, and an antiform in the Little Wenatchee River area. These generalized structures do not continue to the west (Tabor and others, 1982b). The major antiform in the Little Wenatchee River area (fig. 4) brings up the banded gneiss unit and the light-colored gneiss of Wenatchee Ridge, rocks characterized by metasomatic and intrusive material that has transformed the Chiwaukum Schist into light-colored tonalite gneiss. The uplifted rocks also bear kyanite in contrast to staurolite on the limbs of the antiform, indicating the core formed at greater depths. Local blastomylonite zones in the gneiss suggest pronounced movement parallel to foliation. The antiform might be viewed as an incipient gneiss dome (see, for instance, Fox and Rinehart, 1971).

Chiwaukum Schist

Page (1939a, p. 15-16; 1940) named the Chiwaukum Schist for exposures along Chiwaukum Creek [19] and its main tributary, Glacier Creek. The predominant rocks in the Chiwaukum Schist are garnet and staurolite-bearing graphite-biotite-quartz schist and subordinate muscovite paragneiss, siliceous hornblende schist, amphibolite, calc-silicate schist, and marble. We correlate with the Chiwaukum Schist all the predominantly pelitic schist and interbedded calcic rocks more or less continuous with the type area in the vicinity of Chiwaukum Creek as well as pelitic schists both in the vicinity of Panther Creek [1] (see cross section A-A', map sheet) and stretching northward into the Glacier Peak quadrangle (including the schists of the central schist belt) except for the rocks of the Napeequa River area [3] as shown by Crowder and others (1966). Rosenberg (1961, p. 21-34), Van Diver (1964, p. 15-38), Plummer (1969, p. 54-124; 1980), Getsinger (1978, p. 13-67), and Heath (1971, p. 12-26) described rocks in the Chiwaukum Schist. Also correlated with the Chiwaukum are more or less isochemical, predominantly pelitic, schists mapped by Oles (1956, p. 41- 86), Yeats (1958, p. 17-40) and Tabor and others (1982b) west of the quadrangle. The Chiwaukum Schist includes at least part of the rocks originally mapped as the Green Mountain unit by Bryant (1955) and Ford (1959, p. 27-111).

We have mapped two major informal units in the Chiwaukum Schist: a structurally and probably stratigraphically upper, predominantly pelitic biotite schist and a lower pelitic biotite schist with subordinate amphibolite, hornblende- mica schist, calc-silicate schist, and marble. Minor mapped units are marble and a poikiloblastic mica schist of limited extent and unknown stratigraphic significance.

Most of the above-mentioned workers considered the Chiwaukum a metamorphosed sandy to argillaceous deposit with local mafic igneous rocks and carbonate rocks. Plummer (1980) and Kaneda (1980) documented the metamorphic history recorded by the growth of aluminum silicate minerals in the Chiwaukum type area, showing that the rocks underwent a progressive regional synkinematic metamorphism of Barovian-type (high pressure) followed by a Buchan type metamorphism (lower pressure) associated with continued deformation during intrusion of the Mount Stuart batholith. The isograds shown on the map, which probably express both episodes, are derived from a number of sources: the garnet-staurolite and staurolite-kyanite isograds are from Plummer (1980, p. 387), and the less well controlled northern kyanite-staurolite isograd is from scattered data in Rosenberg (1961) and Van Diver (1964), our reconnaissance work, and projection from the staurolite-kyanite isograd in the Glacier Peak quadrangle (Crowder and others, 1966).

Closely associated with the micaceous schist of the Chiwaukum are banded gneiss, biotite gneiss, light-colored tonalite gneiss, and alaskitic gneiss derived from the schists by addition of quartzofeldspathic materials by metasomatism and intrusion. We exclude this banded gneiss from the Chiwaukum Schist although the contact between the two units is very gradational and layers of the Chiwaukum Schist are prominent in much of the gneiss.

Biotite schist and amphibolite—Biotite-quartz schist, hornblende-biotite schist, and subordinate gneissic amphibolite, hornblende schist, calc-silicate schist, and marble appear to underlie the biotite schist unit (cross section A- A', map sheet). The contact between the predominantly biotite schist unit and the biotite schist and

amphibolite unit is marked only by the pronounced increase in the amount of hornblende in mica schist and in amphibolite and other calcium-bearing schists. We think the contact was primarily depositional and time transgressive.

Good descriptions of the biotite schist and amphibolite unit are in Rosenberg (1961, p. 31-34), Van Diver (1964, p. 26-36), Getsinger (1978, p. 3456), and Kaneda (1980, p. 15-35). The unit grades into the banded gneiss unit which is exposed in an antiform roughly aligned along the Little Wenatchee River (cross-section A-A', map sheet). The biotite schist and amphibolite unit exposed on the northeastern limb of the antiform was previously termed the Whittier Peak unit by Rosenberg (1961, p. 57-74) and Van Diver (1964, p. 15-41).

Marble—Marble crops out sporadically throughout the lower part of the Chiwaukum Schist but is most abundant in a zone along the Little Wenatchee River where it has been quarried for cement.

Biotite schist—Overlying and interfingering with the biotite schist and amphibolite unit and exposed in a complex synform (cross section A-A', map sheet) is biotite-quartz schist. This unit is rich in quartz segregations and veinlets and is isoclinally folded and crinkled. Original bedding, shown by graphite, mica, and quartz laminae, is locally preserved (see Plummer, 1980, part 11, p. 1634). Relict graded beds are preserved in the cliffs west of lower Chiwaukum Creek [19].

Poikiloblastic mica schist—Along the northern margin of the Chiwaukum Schist from Chiwawa Ridge to Mount David [2] is the poikiloblastic mica schist unit characterized by considerable chlorite, muscovite, and microscopic albite poikiloblasts. Van Diver (1964, p. 86-88; 1967, p. 147-148) considered these rocks to be retrograded Chiwaukum. He attributed the widespread albite micropoikiloblasts in the mica schists to sodium metasomatism associated with the shearing of the Ten Peak pluton along a deep-seated fault (see discussion of Mad River terrane and the Ten Peak pluton). The replacement of plagioclase by zoisite in metagabbro in the Mad River terrane to the north might indeed free enough sodium to produce the sodic plagioclase in the schist. However, much of the chlorite in schist of this unit is in equilibrium with biotite; the rocks have a low degree of metamorphic segregation and bear no relict aluminum silicates. Evidence for strong retrogression is generally lacking. On the south end of Chiwawa Ridge, just north of the poikiloblastic mica schist unit, rocks in the schist of Crook Mountain are clearly low grade and only show a progressive metamorphic history.

Schist of Crook Mountain [5]

On the southern end of Chiwawa Ridge, the poikiloblastic mica schist unit grades over a short distance into a variety of phyllitic schist with predominantly high-rank greenschist to low-rank almandine amphibolite facies mineralogy. Many of these rocks look like phyllites, and most do not show textural evidence of retrogression. In the schist are metaporphyritic dikes and some ultramafite pods, rock types common to both the micaceous quartzite and

mica schist on Mount David and the heterogeneous gneiss and schist unit of the Mad River terrane farther north on Chiwawa Ridge. We are uncertain about the terrane affinity of this unit, but its position on the Nason Creek terrane side of the tectonized rocks along the contact with the metagabbro, metadiorite, and metadiorite gneiss unit suggests that it is a low-grade part of the Nason terrane. Its low degree of metamorphism is enigmatic.

Ultramafite

Small pods and layers of ultramafite in the Chiwaukum Schist above Icicle Creek may well be pieces of the Ingalls Tectonic Complex, tectonically imbricated in the Chiwaukum or emplaced as serpentinite landslides into the Chiwaukum shales prior to metamorphism. A conspicuous, discontinuous layer of metaperidotite (0.15-0.20 km thick by 5 km long) paralleling compositional layering in the biotite schist and amphibolite unit near Frosty Pass [14] and adjacent to the Mount Stuart batholith is the best candidate for an ultramafic body having a sedimentary origin. Although Kaneda (1980, p.37) indicates that the present contact mineral assemblage is consistent with derivation from a serpentinite parent, no other structures suggesting landslide origin have been found.

Banded gneiss

On a regional scale, the Chiwaukum Schist grades into a unit characterized by conspicuous thin to very thick layers and irregular masses of light-colored tonalite gneiss and light-colored fine-grained to pegmatitic dikes and sills. Much of the unit is alternating mica schist, hornblende schist, amphibolite, and light-colored tonalite gneiss. Gneiss and light-colored dike rocks range from 10 to 99 percent of the terrane. Contacts between the gneiss and schist are sharp in some places and gradational in others. The rocks are locally migmatitic especially where light-colored gneiss predominates. Foliation is generally conspicuous everywhere; massive rocks are rare. The occurrence of kyanite in siliceous mica schist layers, evidence of higher temperature and pressure, more or less coincides with the extent of the banded gneiss.

The banded gneiss unit corresponds to Rosenberg's (1961, p. 35-54) Poe Mountain subdivision of the Green Mountain-White Chuck unit and is continuous with rocks in the Glacier Peak quadrangle mapped as biotite gneiss by Crowder and others (1966).

Early workers (Oles, 1956, p. 86-111; Rosenberg, 1961, p. 35-54; Van Diver, 1964, p. 51-66; 1967, p. 146-147) considered much of the gneiss to be of replacement origin, although Rosenberg (1961, p. 2130) considered some of the fine-grained gneiss to have a sedimentary protolith that was metamorphosed without appreciable metasomatism. Gradational contacts, relics of schist and mafic schlieren, and textural evidence of replacement all suggest that the banded gneiss unit formed from the Chiwaukum Schist; the schist was at least partially replaced by quartzofeldspathic (leucotrochhemitic) material during regional synkinematic metamorphism. Considerable material has been added to the banded gneiss unit by intrusion as shown by crosscutting contacts, dilation, and relict igneous textures in some gneiss. Layers of porphyroblastic mylonitic gneiss throughout the unit suggest considerable

local deep-seated shearing. Getsinger (1978, p. 135-136) concluded that the light-colored gneiss on Nason Ridge was emplaced late in the regional metamorphism. If so it may be related to the intrusion of the Mount Stuart batholith.

Light-colored gneiss of Wenatchee Ridge—First described and named by Van Diver (1964, p. 51-62), the light-colored gneiss of Wenatchee Ridge is a mappable concentration of the light-colored tonalite gneiss and dikes characteristic of the banded gneiss unit elsewhere. Pegmatite is an important constituent. The gneiss is very heterogeneous and appears to be made up of numerous dikes, sills, and irregular masses with only a few percent of relict schist in pods and layers.

Biotite granodiorite gneiss—The biotite granodiorite gneiss unit is similar to some layers of light-colored gneiss in the banded gneiss unit but is in larger, relatively uniform masses. Only small parts of the gneiss bodies are exposed on the extreme western edge of the Chelan quadrangle. Contacts are generally conformable and interlayered with schist but are locally crosscutting. Textural features such as euhedrally zoned plagioclase cores and intergranular quartz suggest an early igneous history, although the bodies are predominantly crystalloblastic now and in mineral equilibrium with the surrounding schist.

Age of the Nason terrane

We have no isotopic data to help establish the protolithic age of the Chiwaukum Schist. If, as we suggest, some of the ultramafite in the schist was originally emplaced as serpentine slides from the Late Jurassic Ingalls Tectonic Complex, then the Chiwaukum sediments were deposited between the Late Jurassic and the Late Cretaceous.

Although we have no direct isotopic age for the initial metamorphism of the Chiwaukum, the last episode of metamorphism took place in the Late Cretaceous during intrusion of the late- to postkinematic Mount Stuart batholith and the Ten Peak pluton. The Ten Peak pluton intrudes both the Mad River and Nason terranes, indicating that the two terranes were sutured by then (Tabor and others, in press). The age of the early metamorphism described by Plummer (1980, part 11, p. 1631-1660) is not known, although his data do not preclude a lengthy, multiphased, metamorphic event in the Late Cretaceous.

Potassium-argon ages of biotite and muscovite from schist, gneiss, and pegmatite of the Nason terrane are in a fairly tight cluster around 81 to 85 m. y. (table 1, Nos. 35-37). Hornblende from an amphibolite, and biotite and muscovite from the biotite granodiorite gneiss unit west of the Chelan quadrangle also yield ages in the Late Cretaceous range (Tabor and others, 1982c). The near concordancy of different minerals from a variety of rocks suggests that the ages represent the last episode of regional metamorphism and(or) the intrusion of the youngest synkinematic Mount Stuart pluton dated at about 85 to 87 m.y. (see below). Of particular importance is the uniformity of the K-Ar ages in the Nason terrane in contrast to the wide range of ages in the terranes to the

east. In particular, there is no evidence of the 50 m.y. event demonstrated by both K-Ar and zircon ages of meta-igneous rocks in the Mad River terrane. The isotopic evidence suggests that the metamorphism and intrusion in the Nason terrane was followed by rapid uplift and cooling of the rocks now exposed at the surface. In the Mad River, Swakane, and Chelan Mountains terranes, the rocks stayed hot longer and were more widely affected by Tertiary thermal events.

PLUTONS INTRUDED AFTER TERRANES WERE ASSEMBLED

Mount Stuart batholith

The Mount Stuart batholith crops out in the southwestern part of the quadrangle and underlies Mount Stuart (elevation 2,870 m), located just south of the quadrangle. The batholith consists of two major irregular elongate plutons and several smaller bodies (see Tabor and others, 1982c, fig. 1). Much of the elongate eastern pluton crops out in the Chelan quadrangle, but only a small part of the major western pluton is exposed as the tonalite of Harding Mountain [23]. The overall length of the batholith is about 62 km.

Since its early description by Russell (1900, p. 105107) considerable work has focused on the batholith (Smith, 1904, p. 4, 5; Page, 1939a; Pratt, 1958, p. 4649; Plummer, 1969, p. 10-34; Engels and Crowder, 1-971; Pongsapich, 1974; and Erikson, 1977).

Tonalite, granodiorite, diorite, gabbro, and contact complex—Discordant K-Ar ages of hornblende and biotite and fission-track ages of allanite require some interpretation. In general, we believe the oldest hornblende age to be closest to the true age. The oldest hornblende age from the major eastern tonalite pluton, exposed northwest of the Chelan quadrangle, is 95 ± 3.3 m.y. (Engels and Crowder, 1971, p. D42; this and following ages are corrected for new constants). An amphibole from diorite and gabbro, presumably an early phase exposed in Icicle Creek, is about 95 m.y. (table 1, No. 52). Biotite ages of these same rocks are 82 and 90 m.y., respectively, and were probably degraded by prolonged burial or later events. A best estimate based on all ages for the eastern pluton is about 93 ± 3 m.y. Tonalite dominates the batholith, but a considerable amount of granodiorite crops out to the south in the Wenatchee quadrangle (Erikson, 1977, fig. 2), where Smith (1904, p. 4) called it the Mount Stuart Granodiorite. Much of the batholith is roughly concordant with the country rock foliation and shows a consistent faint planar alignment of minerals roughly parallel to its contacts. Although the south end of the batholith is surrounded by ultramafic rocks of the Ingalls Tectonic Complex, the contact is marked by a discontinuous selvage of biotite and hornblende schists and metaporphry dike rocks, in part tectonically mixed with the ultramafite. This selvage, or contact complex (contact schist of Smith, 1904), is strongly overprinted by static thermal metamorphism. Schistose metaporphry in the contact complex and elsewhere in the Chiwaukum Schist may be early intrusive phases of the batholith (Tabor and others, 1982a; Miller, 1980b, p. 227-230).

Metagabbro and metadiorite-In our preliminary report (Tabor and others, 1980), we concurred with Getsinger (1978, p. 137-139) that some of the metagabbro and metadiorite bodies cropping out in the Nason terrane and common along the margins of the Mount Stuart batholith were emplaced prior to intrusion of the main batholith. In contrast, Plummer (1969, p. 31-36), with some reservation, and Erickson (1977, p. 184) concluded that most of these masses were early mafic phases of the Mount Stuart. Although several lines of evidence suggest that the metadiorite and metagabbro are older than the Mount Stuart and its early mafic phase on Icicle Ridge, isotopic ages indicate that they are only slightly older or the same age as the main-phase tonalite of the eastern pluton of the batholith.

The metagabbro and metadiorite bodies have textures suggesting static recrystallization, a metamorphism probably associated with the intrusion of the main part of the batholith. Based on analysis of structures in schists adjacent to two bodies of metagabbro exposed on Big Jim Mountain [21] and to the south, B. R. Frost (written commun., 1977) concluded that the metagabbro was involved in the folding of the Chiwaukum Schist. The tonalite of the batholith is not folded, and hence it must be younger than the metagabbro and metadiorite bodies. In general, the metagabbro and metadiorite bodies show little evidence of internal deformation, but rocks in the area of mixed metagabbro, ultramafite, and schist may be tectonically imbricated. Getsinger (1978, p. 138) suggested that small metadiorite bodies on Nason Ridge might be emplaced tectonically into the schist.

Substantiating evidence for deformation following emplacement of the metagabbro-metadiorite body at Big Jim Mountain are schistose metatonalite porphyry dikes that intrude nonschistose metagabbro. Inclusions of metadiorite in the dikes are strongly strung out by shearing indicating that although penetrative deformation was resisted by the massive metagabbro-metadiorite body, strong shearing was taken up along the dikes. Tabor and others (1980) assumed that these dikes were derived from the tonalite magma of the Mount Stuart and deformed in the waning stages of regional metamorphism accompanying the intrusion of the batholith.

Despite the field indications for an age difference between the tonalite of Mount Stuart and the metagabbro and metadiorite bodies, isotopic ages only marginally substantiate such a distinction. Concordant U-Th-Pb ages of 96 m.y. from zircon of the metagabbro-metadiorite of Big Jim Mountain (table 1, No. 55) are within the error range (93 ± 3 m.y.) of K-Ar ages for the main phase of the eastern pluton.

Tonalite of Harding Mountain-The tonalite of Harding Mountain [23] forms the southeastern end of the major southwestern pluton of the batholith. The body is separated from the northeastern pluton by a screen of Chiwaukum Schist and metamorphosed rocks of the Ingalls Tectonic Complex (see Tabor and others, 1982c). The major part of this southwestern pluton exposed to the northwest is composed of tonalite and granodiorite which look like typical Mount Stuart rocks (see Erickson, 1977, fig. 1). Tonalite exposed on Harding Mountain, however, has a slightly different outcrop appearance than most rocks of the Mount Stuart and is somewhat less mafic than parts of the eastern pluton. It does not differ much otherwise, petrographically or chemically (see Erickson, 1977). Outcrops are highly jointed and disintegrate into

extensive talus mostly of small blocks 10-20 cm on a side. The Mount Stuart rocks elsewhere form bold cliffs commonly broken by widely spaced joints, and talus is typically built of larger blocks.

The western pluton appears to be slightly younger than the eastern pluton. Hornblende and biotite from the tonalite of Harding Mountain exposed on The Cradle, just west of the quadrangle, yield concordant K-Ar ages of 87.5 ± 3.0 and 87.7 ± 0.2 m.y. (Tabor and others, 1982c), respectively. Hornblende and biotite from the main part of the southwestern pluton near Scenic (west of the quadrangle) yield 85.5 ± 2.7 and 84.6 ± 2.5 m.y., respectively (Engels and Crowder, 1971).

Ten Peak and Dirty Face plutons

Granitoid rocks of the Ten Peak and Dirty Face plutons along the White River were first described by I. C. Russell (1900, p. 107). The Ten Peak pluton north of the Chelan quadrangle was first mapped by Cater and Crowder (1956) and by Ford (1959, p. 149). Van Diver (1964, p. 89; 1965; 1967) formalized names used by the earlier workers to White River Orthogneiss, but the name White River is preempted. In the Glacier Peak and Holden quadrangles to the north, Crowder and others (1966) and Cater and Crowder (1967) referred to granitoid gneiss exposed in the White Mountains and along Chiwawa Ridge as the Ten Peak pluton, and the adjacent eastern body as the White Mountain pluton. The composition, texture, and structure indicate that the two bodies, separated by a thin screen of amphibolite and faults, are essentially products of the same igneous and metamorphic events. The name White Mountain is also preempted, and as the two bodies are essentially continuous we will refer to both as the Ten Peak pluton. The Ten Peak pluton yields a K-Ar hornblende age of about 93 m.y. (table 1, No. 56).

The smaller Dirty Face pluton is very similar to the Ten Peak although overall somewhat more mafic. The Dirty Face pluton was first described briefly by Willis (1950, p. 30) as part of his Chelan facies of the metamorphic complex. Van Diver (1964, p. 42-46) determined that the gneissic tonalite of the Dirty Face pluton is an orthogneiss. The description that we give below for the Ten Peak suffices for the Dirty Face pluton as well.

Based on the metamorphic minerals in the Ten Peak pluton, its general conformity with the regional foliation, lack of thermal aureole, and interlayering of country rock and tonalite, early workers (see Ford, 1959, p. 163) concluded that the pluton was metasomatically derived from the surrounding schist, that is, formed by granitization. However, the gneissic to massive quartz diorite, tonalite, and gabbro of the Ten Peak and Dirty Face plutons display ample field evidence of igneous origin: disoriented inclusions, crosscutting contacts, and apophyses and dikes that intrude the enclosing schist.

The common occurrence of relict euhedral oscillatory zoning and synneusis twins in plagioclase as well as healed cataclastic texture and flaser structure in the Ten Peak pluton led Van Diver (1964, p. 89-124; 1967) to conclude that the pluton was of igneous origin but metamorphosed during the main Cascade metamorphism (Misch, 1966, p. 102). Van Diver (1967) concluded that blastomylonites and healed cataclastic textures especially concentrated along the southern and western margins of the pluton and locally within it, as well as retrogressive polymetamorphic textures in the adjacent schist, indicated major paracrystalline (synmetamorphic) fault movement.

Some of the blastomylonitic rocks described by Van Diver are in our metagabbro, metadiorite, and metadiorite gneiss unit of the Mad River terrane. On the other hand, Cater (1982, p. 82) considered all the signs of brittle deformation and subsequent healing in the Ten Peak pluton to be evidence of protoclasia which occurred during deep-seated igneous intrusion of the pluton some time after regional synkinematic metamorphism had ceased.

In spite of evidence of igneous intrusive origin and of shearing of the crystalline mass, textures and mineralogy indicate that the pluton and surrounding metamorphic rocks reached approximate equilibrium within the almandine amphibolite facies of regional metamorphism. The pluton appears to have intruded at mesozonal to catazonal depths (Buddington, 1959, p. 708) during regional metamorphism. Based on the occurrence of igneous epidote in the Ten Peak pluton, Zen and Hammarstrom (1984) estimate a crystallization depth of about 30 km. The Late Cretaceous age of the Ten Peak pluton (93 m.y.) and the Mount Stuart batholith (85-96 m.y.) is somewhat older than the Entiat and Chelan plutons (about 80 m.y.) of the Chelan Mountains terrane, but still about the same age as North Cascade metamorphism, between 60 to 90 m.y. as determined by Mattinson (1972, p. 3774).

TERTIARY BEDROCK

Pre-Miocene bedrock

DIABASE

Small bodies of partially altered diabase intrude quartz diorite of the Mount Stuart batholith on Icicle Ridge. The nearest rocks of similar composition are the diabase and gabbro dikes and sills intruding the Chumstick Formation in the Camus Land area south of the quadrangle (Tabor and others, 1982a).

DUNCAN HILL PLUTON AND ASSOCIATED DIKE SWARMS

Intruding rocks of the Chelan Mountains terrane, the Duncan Hill pluton forms an elongate teardrop-shaped body stretching about 48 km from Baldy Mountain west of Lake Chelan to the upper Entiat River in the Holden quadrangle (Cater and Crowder, 1967; Cater and Wright, 1967). The pluton grades from its north end, where it is tonalitic, "protoclastic," gneissic, and bordered by a complex of deep-seated contact rocks, to its south end where it is granitic, sharply crosscutting, miarolitic, and granophyric. Hopson and others (1970) and Cater (1982, p. 61) suggested that the pluton has the shape of a blade or teardrop in cross section; now that the pluton is exposed by differential (tilted) uplift and erosion, its teardrop shape appears much elongated. We are seeing along its length a vertical cross section ranging from a deep-seated catazonal root at the north end to an epizonal top at the south end (see also Hopson and Mattinson, 1971).

Granophyric granite and rhyolite dikes are abundant at the southern end of the Duncan Hill pluton, and some grade into the granite mass just west of Lake Chelan. The dikes also cut the granite and the granodiorite phase of the pluton to the northwest. Both the pluton and dikes yield hornblende and biotite ages ranging from 43 to 48 m.y. (table 1, Nos. 26-28). Zircons from the gneissic northwestern end yield roughly concordant U-Th-Pb ages of

about 48 m.y. (No. 25). Granite porphyry dikes form a northeast-southwest-trending swarm between the granite on Slide Peak [28] and the Cooper Mountain pluton of Barksdale (1975, p. 17) just out of the quadrangle to the northwest. The age of the granodiorite of the Cooper Mountain pluton is also about 48 m.y. (table 1, No. 31). On strike to the northwest of the Cooper Mountain pluton is the granodiorite of the Railroad Creek pluton (Cater and Wright, 1967), which yields K-Ar ages on hornblende of 43.7 ± 2 m.y. and on biotite of 44.8 ± 1.3 m.y. (Engels and others, 1976). Schist in upper Washington Creek [32] in the northeast part of the quadrangle is hornfelsed, suggesting that the subsurface extension of the Cooper Mountain pluton may not be far below.

CHUMSTICK FORMATION

In the area of the Chelan quadrangle, the rocks of the terrestrial middle Eocene Chumstick Formation (Gresens and others, 1981; see also Buza, 1976, 1979; Whetten, 1976; Whetten and Laravie, 1976; and Gresens and others, 1977) are white to gray micaceous feldspathic sandstone with varying but lesser amounts of interbedded pebbly sandstone, conglomerate, minor shale, and rare siliceous tuff. The formation includes rocks originally called the Camas Sandstone by Russell (1900, p. 118) and Alexander (1956, p. 16) but, until recently, usually referred to the Swauk Formation (for instance, Waters, 1930; Chappell, 1936b; Page, 1939a; Willis, 1953; Cater and Crowder, 1967), which is now known to be early Eocene (Gresens and others, 1977; Tabor and others, 1982a, 1984) and thus older than the Chumstick.

The Chumstick Formation crops out in the Chiwaukum graben, an erosional lowland developed on the downdropped sandstone and shale. The graben is bounded on the east by the Entiat fault, which is well exposed in a fault line scarp along the west side of the Entiat Mountains, and on the west by the Leavenworth fault, which is well exposed along logging roads east of Tumwater Canyon.

Along the eastern side of the graben, concurrent with downfaulting of the graben, the Chumstick was deposited upon the Swakane Biotite Gneiss (Whetten, 1976). Transport directions were dominantly to the southwest across the graben (Buza, 1979), but detritus in fanglomerates along the Leavenworth fault shows that the Mount Stuart block was a local source as well (Cashman, 1974; Frizzell and Tabor, 1977). Modal data on the feldspathic to lithofeldspathic sandstone, although indicating a significant proportion of volcanic rocks in the source terrane, are compatible with a crystalline source such as the Swakane Biotite Gneiss (Frizzell, 1979). The Chumstick originated as fans and flood-plain deposits and as such is extremely variable in thickness. Whetten (1976) estimated that there was at least 5,800 m of sedimentary rocks in the Chumstick Formation in its type area.

Tuffs in the Chumstick yield middle Eocene (about 45 m.y.) fission-track ages on zircon (table 1, Nos. 1523~ Whetten, 1976, p. 420; Gresens and others, 1981). Palynomorphs indicate a late Eocene age for the Chumstick (Newman, 1971, p. 397-398; 1975, p. 158). Fossil leaves have led most workers to correlate the unit with the Swauk (Brown in Waters, 1930, p. 157; LaMotte in Chappell, 1936a, p. 70), although the LaMotte collection was a composite from both the Swauk and the Chumstick. A modern determination based upon leaves has not been attempted.

Redbed fanglomerate member—Basal deposits commonly exposed on underlying Swakane Biotite Gneiss consist of angular to subrounded clasts of biotite gneiss and minor vein quartz in a matrix of reddish sandstone. Gresens and others (1981) interpreted these redbed fanglomerates to be the remains of alluvial fans that mantled knobs and ridges of basement rock within the graben.

Conglomerate and conglomeratic sandstone member and monolithologic fanglomerate breccia member—Lenses of conglomerate crop out discontinuously along the eastern side of the graben adjacent to the Entiat fault. Similar lenses occur less commonly along the Leavenworth fault in the Chelan quadrangle where they locally become almost monolithologic fanglomerates (Gresens and others, 1981, part 11). A very coarse monolithologic fanglomerate breccia of tonalite derived from the Mount Stuart batholith forms a thick lense east of Tumwater Canyon, but similar rocks are more common farther south along the fault in the Wenatchee quadrangle (Cashman, 1974; Cashman and Whetten, 1976; Frizzell and Tabor, 1977; and Tabor and others, 1982a). The conglomerate interfingers with the sandstone and shale units in the Chumstick.

Nahahum Canyon Member—Shaly sandstone and shale conformably rest on the lower beds of the Chumstick and interfinger and grade into the conglomerate and sandstone members. The evenly and finely laminated shale and interbedded thin turbidite sandstone indicate deposition in shallow lakes (Gresens and others, 1981).

Tuff and tuffaceous sandstone interbeds—Thin beds of siliceous tuff and tuffaceous sandstone occur throughout the Chumstick Formation. Many are mappable and some have been traced for many kilometers. The remarkable continuity of these beds is evidence that the graben filled with little erosional interruption. Fission-track ages on zircon from tuff samples cluster at about 45 m.y., which we take to be the age of deposition of the formation. The average ages of tuff interbeds are statistically the same and cannot be separated, which suggests very rapid sedimentation. The source of the tuffs is not known, but we speculate that it may have been a volcano related to the Duncan Hill or Cooper Mountain and Railroad Creek plutons, northeast of the graben, because the ages of the tuffs and the plutons are about the same (see description of the Duncan Hill pluton). On the other hand, volcanoes feeding the late Eocene and Oligocene (?) Naches Formation about 75 km to the west may have been the sources (Tabor and others, 1984).

PORPHYRITIC DACITE OF BASALT PEAK

A flatlying thick lens of biotite-hornblende porphyritic dacite intrudes the Chumstick Formation and the Swakane Biotite Gneiss on Basalt Peak [7]. Several smaller bodies of the same rock type intrude nearby. Willis (1950, p. 114117) described the body as a laccolith, and although it appears to be parallel to the bedding in the Chumstick, the general structure of the sandstone is too poorly known to be sure of the concordance.

Cater (1982) believed the porphyritic dacite of Basalt Peak to be identical to the dacite of the Old Gib volcanic neck exposed a few kilometers north of the Chelan quadrangle (see Cater and Crowder, 1967). Biotite from Old Gib yields an age of 45 ± 1.5 m. y. (Cater and Crowder, 1967; Engels and others, 1976, No. 119). The

hornblende K-Ar age of a porphyritic andesite dike cutting the Chumstick in Walker Canyon [39] is about 42 m.y. (table 1, No. 5) and thus may be related to the Old Gib igneous activity.

WENATCHEE FORMATION

In the Wenatchee quadrangle (Tabor and others, 1982a) to the south, the Wenatchee Formation (Gresens and others, 1981) generally overlies the Chumstick Formation along an unconformity first recognized by Chappell (1936a, p. 93-94). Included in the unit in the area of Wenatchee are quartzose sandstone, tuffaceous shale, and a conglomerate that contain clasts of felsic volcanic rocks and vein quartz. Shales have red oxidized horizons. In the Chelan quadrangle, rocks referred to the Wenatchee Formation unconformably overlie the Swakane Biotite Gneiss on Burch Mountain, north of Swakane Creek [38], and south of the Columbia River. No coherent stratigraphy is present within these patches but rock types are the same as those closer to Wenatchee. Tuffaceous beds in the Wenatchee Formation yield zircons with fission track ages of about 34 m.y., or early Oligocene (Gresens and others, 1977, p. 109, and 1981, p. 236). Palynomorph assemblages also yield an Oligocene age (Newman, 1975, p. 158).

VOLCANIC ROCKS OF CHIKAMIN CREEK

Massive rhyolite grading downward into crystal-lithic rhyolite tuff and breccia apparently overlies tonalite gneiss of the Entiat pluton in a small area of upper Chikamin Creek [8]. The age of the volcanic rocks is 23 to 28 m.y. based on fission-track ages of zircon and apatite (table 1, Nos. 3 and 4), surprisingly old because the deposit appears to lie within the present valley. The nearest known eruptive rocks of early Oligocene age are the 34 m.y.-old tuffs in the Wenatchee Formation near Wenatchee, south of the quadrangle.

COLUMBIA RIVER BASALT GROUP

By **D. A. Swanson and G. R. Byerly**

Yakima Basalt Subgroup

The Columbia River Basalt Group underlies the southeast half of the quadrangle and extends far to the east and south, where it forms the Columbia Plateau of eastern Washington and adjacent Oregon and Idaho. The Yakima Basalt Subgroup (Swanson and others, 1979) is divided into three formations: in ascending order, the Grande Ronde, Wanapum, and Saddle Mountains Basalts. Only the Grande Ronde and Wanapum Basalts occur within the Chelan quadrangle. Evaluation of K-Ar ages and paleontologic evidence relative to the late Cenozoic time scale of Berggren and Van Couvering (1974) indicates that the Grande Ronde Basalt is mostly or wholly of late early Miocene age (16.5 to 14 m.y.) and the Wanapum Basalt is of middle Miocene age (Swanson and others, 1979).

The basalt covered an erosional surface of moderate local relief in the quadrangle. Map relations along the northern part of the Waterville Plateau (fig. 4) imply a prebasalt hill at least 240 m high (Waters, 1930). Relief along

the prebasalt surface near the junction of Pine [36] and Corbaley Canyons is about 120 m. The basalt flows lapped up against a highland not far beyond the present-day margin of the field, as shown by a general northwestward thinning of stratigraphic units and the southeast current directions indicated by sandstones interbedded with the basalt. Exposures on Dick Mesa [34] and at the head of Mills Canyon [42] are probably remnants of flows that advanced up broad valleys near the margin of the highland.

All of the basalt apparently was erupted from vents farther east or southeast of the quadrangle. Flow directions, measured chiefly on foreset-bedded lava deltas, consistently show west and northwest movement.

Grande Ronde Basalt

In the Chelan quadrangle, the Grande Ronde Basalt consists of tholeiitic basalt flows that are nonporphyritic or contain only rare, small plagioclase phenocrysts. Many of the flows invade sedimentary rocks, which were not lithified at the time; such invasive flows are described later in this summary.

Lithologic similarity and discontinuous jointing habits make field identification and correlation of single flows across unexposed areas difficult. Elsewhere on the Columbia Plateau we have determined remanent magnetic polarity with a portable fluxgate magnetometer and on this basis have been able to subdivide the formation in the field (Swanson and Wright, 1976; Swanson and others, 1979;

Hooper and others, 1976). We found that the most reliable measurements are obtained on oxidized parts of a flow, especially the base and top where magnetic stability is highest (Swanson and Wright, 1976). Few flows with such oxidized zones were found in the quadrangle; however, we discovered that samples with glassy rinds, formed by rapid quenching against water or sediment, give far more reliable results than unoxidized and unquenched samples, presumably because the small grain size of magnetic minerals favors magnetic stability. Samples with glassy margins yield less consistent results than oxidized ones but are nonetheless usable with care. As glassy material is abundant throughout most of the quadrangle, we were able to establish a magnetostratigraphy in the field with reasonably consistent measurements; anomalous readings were easily identified and corrected. The map units, flows of reversed magnetic polarity and flows of normal magnetic polarity, are the result of this work and are time correlatives with similar units mapped on the Columbia Plateau (the R₂ and N₂ magnetostratigraphic units of Swanson and others, 1979, 1980).

Flows of reversed magnetic polarity (R₂)—The flows of reversed polarity form a unit recognizable along the cliffs forming the western and northern margins of the Waterville Plateau. The thickest measured section is along U.S. Route 2 in Pine Canyon [36] where three flows along with interbedded and invaded sedimentary rocks total 145 m in thickness. Elsewhere, the unit is generally less than 75 m thick, although landslide debris commonly covers the base of the unit and the true thickness could be substantially greater. The variability in thickness reflects relief on the prebasalt surface.

We mapped one subunit, the invasive flow of Hammond (Hammond sill of Hoyt, 1961), within the R₂ unit in places where it is readily separated in the field from other reversely magnetized flows. The Hammond is chemically distinctive, unusually thick (more than 50 m in most places), and has a flat, nonscoriaceous upper surface chilled against sedimentary rocks of the Ellensburg Formation. The Hammond occurs all along the margin of the plateau in the quadrangle, although it is mostly obscured by landslide debris south of Corbaley Canyon. Correlations of many isolated outcrops have been confirmed by chemical analyses.

Flows of normal magnetic polarity (N₂)-The flows of normal polarity form the most widespread and thickest unit of basalt in the quadrangle. No more than six flows were recognized in any section, although complex pillowed flows and hyaloclastites make identification of many flow contacts rather arbitrary.

We divided the flows of normal magnetic polarity into two subunits in the southeast part of the quadrangle: the invasive basalt flows of Keane Ranch, which contain several flows of at least three different chemical compositions, and the basalt flow of Beaver Creek. Both subunits are best defined in the Wenatchee quadrangle to the south (Swanson and others in Tabor and others, 1982a, p. 13-14).

Toward the crest of Badger Mountain (fig. 4), the invasive flows of Keane Ranch become normal subaerial flows, lose their distinctiveness, and cannot be separated from the rest of the normally polarized Grande Ronde Basalt. Likewise, the flow of Beaver Creek becomes increasingly more difficult to distinguish from underlying flows toward the north, and we mapped it only on the lower slope of the Badger Mountain anticline.

Interactions of basalt with water and sediment

The Grande Ronde Basalt in the Chelan quadrangle differs notably from correlative flows in most other areas on the Columbia Plateau because of the extensive development of pillow basalt, hyaloclastite, lava deltas, and invasive flows, all of which record interaction of lava with water or wet sediment. Pillows and intermixed hyaloclastite formed when lava was quenched as it entered water, probably shallow lakes created as the advancing flows dammed rivers draining the ancestral Cascade Range.

Foreset-bedded lava deltas (Moore and others, 1973) formed in places where current action was slight and water depth was several meters or more. The best example of such a delta is exposed in roadcuts along rerouted U.S. Route 2, in the SE₁/₄, sec. 26, T. 25 N., R. 22 E. The foreset layers in lava deltas dip in the direction of flow.

Basalt flows that burrowed into fine-grained, unconsolidated sediments, forming local peperites (chaotic mixtures of sediment and invasive volcanic rock) and rather extensive sills, are termed invasive flows (Byerly and Swanson, 1978). Peperites, best described by Schmincke (1967) for rocks of the Columbia Plateau, are recognized by the presence of abundant basalt fragments, many with chilled, glassy margins, mixed with disturbed, fine-grained sediment (now weakly lithified). Fragments range in size from a few millimeters to tens of centimeters across. Such peperites are commonly associated with pillow basalt and hyaloclastite and probably were formed as basalt flows encountered sediment on lake or stream bottoms or on broad alluvial plains.

Sills formed by invasive flows in the quadrangle are 5 to 60 m thick and cover areas of less than 1 km² to more than 100 km²; the invasive flow of Hammond probably covers more than 300 km². Some of the best exposures of invasive flows occur along U.S. Route 2 in Pine Canyon [36]. Another good example is on the promontory in the SW1/4, sec. 16, T. 24 N., R. 21 E. The upper surfaces of invasive flows are nearly planar to slightly wavy, contain few vesicles, and have glass selvages indicating rapid quenching against the enclosing sediment. Locally, thin dikes and sills sprout from the top and intrude the host sediments. In places, dikes rooted in an invasive flow extend several meters above the top of the flow, providing an estimate of the minimum thickness of sediment invaded by the flow. Sediments are commonly slightly baked along the upper contact of the invasive flow, and a thin layer closest to the flow is black owing to reduction during baking.

The sill-like bodies are interpreted to be invasive flows, not sills in the classic sense, for the following reasons: (1) all these bodies cut only sediments; (2) no feeder dikes have been found; (3) the bodies represent the distal portions of a complex lithic assemblage that changes from normal flows southeast of the quadrangle through pillowed flows and hyaloclastites into peperites and sills; (4) flow directional data in lava deltas show that lava moved toward the sill-like bodies, not away from them; (5) microprobe analyses of glassy selvages by G. R. Byerly in 1978 confirm correlations made across the facies changes and show that the invasive flows fit perfectly into the chemical stratigraphy established in nearby areas; and (6) the magnetic stratigraphy shows no anomalies; all flows and interlayered invasive flows in the upper part of the section have normal polarity, and all in the lower part have reversed polarity. We thus interpret each invasive flow to have formed, before the next higher flow, as lava advanced across low areas and burrowed into unconsolidated sediments.

Wanapum Basalt, Priest Rapids Member

The Priest Rapids Member occurs only on the Waterville Plateau (fig. 4). It overlies a subfeldspathic sandstone interbed of the Ellensburg Formation resting depositionally on the weathered top of the Grande Ronde Basalt. Only one flow occurs within the quadrangle, but several flows characterize the unit to the east. We recognized the Priest Rapids in the field chiefly by its reversed magnetic polarity (Rietman, 1966), and confirmed its identity by several chemical analyses which show that the flows are of Rosalia chemical type (Swanson and others, 1979). The Priest Rapids Member presumably once covered most of the Waterville Plateau, but rapid erosion of the Ellensburg Formation beneath the unit caused its rapid disintegration and has reduced it to scattered outcrops. Waters (1930, 1939) first recognized this relation correcting Willis' (1903) inference that the remnants indicate widespread peneplanation.

Ellensburg Formation

The Ellensburg Formation consists of weakly indurated sedimentary rocks and is extensively interbedded with the Yakima Basalt Subgroup in the quadrangle. The formation generally is poorly exposed because of extensive

landsliding and vegetation. Some of the best exposures are: (1) roadcuts along U.S. Route 2 in Pine Canyon (361; (2) cuts along a powerline road in the east half of SE1/4 sec. 35, T. 25 N., R. 21 E.; and (3) on Standpipe Hill [37] south of Waterville.

Most interbeds are less than a few meters thick; only those thicker than 5 to 8 m are shown on the map. Thick interbeds, as much as 20 m, occur in a few places, such as in Pine Canyon and on Standpipe Hill.

The Ellensburg in the Chelan quadrangle consists dominantly of micaceous feldspathic sandstone and siltstone; altered fine-grained tuffaceous deposits are of secondary importance. The subfeldspathic detritus was derived from erosion of older crystalline rocks in the area. Sedimentary structures such as crossbeds and imbrication suggest that the subfeldspathic facies of the Ellensburg was deposited on broad floodplains by small streams flowing southward and southeastward. We found no coarse conglomerate in the feldspathic facies that would indicate channel deposition. Swamps and lakes existed in places, as attested by carbonaceous, leaf-bearing claystone and paper-thin continuous bedding.

The Ellensburg thins eastward and southward in the quadrangle. Interbeds are less common in the Badger Mountain area than along the margin of the Waterville Plateau, possibly because of structural relief at the time of deposition. Lack of remnants of the Priest Rapids Member on Badger Mountain is supporting evidence for earlier uplift of the Grande Ronde Basalt.

MIOCENE AND YOUNGER VOLCANIC ROCKS

Andesite and dacite of Burch Mountain

On Burch Mountain (fig. 4) several plug and dike-like bodies of andesite (Whetten and Waitt, 1978) intrude biotite gneiss as well as sandstone of the Wenatchee Formation. East of Swakane Creek [38] a plug of hornblende-pyroxene dacite has radial, outward-dipping pyroxene lineation and platy joints heavily slickensided suggesting solid movement of the mass, such as in a volcanic vent.

Hornblende from the andesite at Burch Mountain has a K-Ar age of about 11 m.y. and from the plug east of Swakane Creek, 10 m.y. (table 1, Nos. 1 and 2).

Andesite of Sugarloaf Peak

A small mass of pyroxene andesite crops out at the summit of Sugarloaf Peak [18]. Vertical columnar joints exposed under the lookout tower suggest a thick flow as indicated by Page (1939a, p. 73) and Willis (1950, p. 117), but the contact is not exposed and the rock is holocrystalline; it could be a hypabyssal intrusion. The age is uncertain but if the andesite is a flow it may be post Columbia River Basalt Group because it is an erosional monadnock above the Entiat surface, which developed according to arguments developed below-after eruption of the lavas. It may be

about the same age as the andesite at Burch Mountain. Isolated outcrops of fragmented andesite which cap ridges within the Chiwaukum graben (Page, 1939a, p. 74) are actually debris-flow deposits (upland gravel and diamictite) derived from Sugarloaf.

EROSIONAL LANDSCAPE AND SURFICIAL DEPOSITS

By Richard B. Waitt

Erosional surfaces and incision of drainage basins

In the lower Entiat drainage basin, a low-relief upland surface, the Entiat surface of Willis (1903) and Waters (1939) is manifested by broad-crested spurs of accordant height. Modern streams are incised several hundred meters below the surface; the Chelan and Entiat Mountains rise steeply above the surface on the north and west. The surface is beveled across upturned belts of several rock types and is overlain only discontinuously and thinly by colluvium, loess, tephra, and rare patches of gravel: it is an erosion surface, as Willis (1903) inferred.

The age of the Entiat erosion surface has been controversial. Willis (1903) had inferred that the surface postdates the Columbia River Basalt Group, whereas Waters (1939) contended that it was exhumed from beneath a carapace of the basalt and intercalated strata of the Ellensburg Formation. A conspicuous outlier of basalt west of the Columbia River, Dick Mesa [34], is perched upon a surface that Waters inferred to be continuous with the Entiat surface immediately to the west; the erosion surface also might appear to pass beneath the basalt on the east side of the river. The Entiat surface, however, equally appears to pass over the surface of the basalt. The erosion surface extends many tens of kilometers north and northwest of the basalt margin (Waters, 1939; Waitt, 1972) and extends west to beyond the Leavenworth fault. In none of these areas is there evidence that the surface has been covered by basalt. In fact, the westernmost outliers of the Columbia River Basalt Group in the Entiat Mountains have steep basal contacts against the pre-Miocene crystalline rocks: the basalt evidently was emplaced in incised tributary valleys and probably never extended much beyond its present limits. After reviewing the Entiat surface and other similar upland surfaces broadly surrounding the Columbia 'Plateau' (actually a plain), I conclude that most of this regional surface developed during and after emplacement of the basalt. The cause was an interval perhaps as long as 10 m.y. when the great regional dam of basalt held base levels high in adjacent highlands.

Dammed by the Columbia River Basalt field, drainage from the highlands collected as lakes in the natural gutter between the high-relief east-sloping pre-basalt surface and the gently west-sloping basalt surface (Willis, 1887; Waters, 1939; Mackin and Cary, 1965). As this waterway in Saddle Mountains Basalt time (Mackin, 1961) became integrated into a drainage line seeking the Pacific Ocean, the Columbia River was born. Since emplacement of the basalt, the Columbia River and its major tributaries (Wenatchee, Entiat, and Chelan Rivers) and their tributaries have incised hundreds of meters below the levels of the Entiat erosion surface and the basalt. Pliocene uplift of the Cascade Range not only caused this incision, but also tilted the basalt surface eastward. The Cascades

were eventually raised enough to nourish valley glaciers during the global climatic coolings of the Pleistocene Epoch.

Conglomerate of Brays Landing

The Columbia River valley contains only a meager depositional record of the incision of the river below the basalt. Near Brays Landing [35], sparsely exposed but locally distinct pebble conglomerate consists of subangular to rounded clasts of gneiss, quartzite, granite, basalt, dike rocks, and chert conglomerate, which together reveal the Columbia River provenance of the deposit. The conglomerate is distinguished from Pleistocene outwash and catastrophic-flood deposits by its finer clast size, moderate sorting, and lithified state. The conglomerate lies within the Columbia River valley some 240 m above the present river. The incision of the Columbia River and tributaries some 500 m below the surface of the Grande Ronde Basalt before the conglomerate accumulated within the valley indicates that the conglomerate is much younger than the early to middle Miocene basalt. The lithified state of the conglomerate and the postdepositional incision of the Columbia River and its tributary network by 240 m indicate that the unit is at least as old as Pliocene.

Upland gravel and diamictite

Nonlithified to weakly lithified gravel and diamictite caps ridges as high as 600 m above modern drainages. Accumulations as thick as 1.5 m of angular to subrounded clasts of vein quartz locally derived from the Swakane Biotite Gneiss mantle ridges on Burch Mountain. This gravel being loose, it probably postdates the Grande Ronde Basalt.

A diamictite composed of angular clasts of andesite and dacite—the Summit Conglomerate (Page, 1939a, p. 76-78; see also Houghland, 1932)—constitutes the upper 60 m of Natapoc Mountain [17], some 600 m above the Wenatchee River. The clasts composing the diamictite are not identical to any nearby bedrock source, but their angularity, poor sorting, and dimensions as large as 3 m indicate a source nearby. Some clasts in the diamictite are glassy hypersthene-pyroxene andesite mineralogically similar to andesite at Sugarloaf Peak [18] 14 km to the east. The diamictite probably was shed from a volcanic edifice at Sugarloaf that was later stripped from the underlying intrusive rock that now crops out there. The volcanic-clast diamictite on Natapoc Mountain overlies weathered cobble conglomerate nearly identical to modern Wenatchee River gravel. This relation signifies that the ancient Sugarloaf volcano had shed mudflows and sidestream alluvium onto the floor of the ancient Wenatchee River valley. Conglomerate similar to that of the Natapoc Mountain but containing volcanic clasts of smaller maximum size, more rounded, and mixed with crystalline-rock clasts, is sparsely exposed on Pole Ridge 13 km north-northwest of Natapoc Mountain, and on McCue Ridge [16] 10 km to the west. The volcanic-rich deposits, broadly spread across an area of now-bald relief, reveal a former time of less relief, a landscape that probably predates the episode of incision that formed the modern valleys.

Patches of boulder diamictite, consisting of clasts identical to andesite and dacite of the Eagle Rocks area of Burch Mountain, cap ridges within the Chiwaukum lowland a few kilometers southwest of Burch Mountain, relations mapped in more detail by Whetten and Waitt (1978). Surviving lateral levees indicate that at least some of the ancient diamictite originated as debris flows. Some of this diamictite overlies rounded, mixed-lithology Wenatchee River gravel that crops out high on valley sides-similar to the gravel that underlies the diamictite atop Natapoc Mountain (see above). Since emplacement of the diamictite, which must have flowed down old tributary valleys, the entire drainage basin has incised and the former topography become inverted: the deposits are now isolated on ridge crests as high as 500 m above present tributaries.

Patches of diamictite composed of Grande Ronde Basalt clasts cap summits southwest of the Entiat Mountain escarpment at the heads of Wenatchee River tributaries. These ancient debris-flow deposits from the Entiat Mountains indicate that the digitate margin of valley-filling basalt in the Entiat Mountains formerly extended a few kilometers northwest of the most distal outliers. A similar diamictite of Grande Ronde Basalt clasts overlies Wenatchee River gravel atop a knoll above the Wenatchee River southeast of Cashmere (fig. 4). An ancient debris flow must have coursed some 25 km down Mission Creek from the Mission Peak area (off map to south). Since then the entire drainage network has incised 250 m, inverting the earlier topography.

These various upland gravel and diamictite deposits thus cap ridges 100 to 600 m above present drainages. Because the tributaries as well as the trunk streams have incised hundreds of meters into bedrock and thereby inverted the former topography, the upland deposits must be at least as old as Pliocene. These deposits sparsely record incision of the Wenatchee drainage network below the Entiat erosion surface. Some of these deposits probably correlate with the conglomerate of Brays Landing.

In many places debris from summit-capping diamictite has been redeposited downslope as the drainage networks have progressively incised and developed. Successively lower spurs, terraces, and tributary valley floors are mantled by successively younger diamictons. The occurrence of diamictite clasts on some modern fans and valley floors indicates that the process of redistribution is still in progress.

Deposits of catastrophic floods in the Columbia River valley

Catastrophic floods from glacial Lake Missoula swept down the northwestern segment of the Columbia River valley (Waitt, 1977a, 1977b, 1980a, 1980b). Most floodbedload sediment is cobble gravel, but this sediment has several attributes abnormal to ordinary river discharge and river terraces. Among the abnormal features are (1) exceptionally large boulders sparsely entrained in the moderately sorted cobble gravel, (2) sediment bodies in the form of enormous bars rather than terraces, (3) scour depressions around bedrock knolls that protrude from the surface of bars, (4) fields of large boulders littering some bars, especially immediately downvalley of bedrock salients, (5) giant current dunes embellishing the surface of some bars, and (6) internal structure of tall deltalike foreset bedding.

In the wide segment of valley above Entiat, each of five discrete, nested, barlike surfaces in point-bar positions on the east side of the valley is marked by giant current dunes or boulder fields. Because compared to lower altitude deposits, the surface of the 'great' terrace (of Russell, 1900) is only modestly flood scoured or overlain by flood deposits, the floodwaters must not have deeply submerged this terrace surface, which is 200 to 245 m above the modern Columbia River. Conspicuous giant current dunes on two surfaces above the level of the 'great' terrace therefore must have been caused by floods predating the accumulation of the 'great'-terrace deposits. Evidence of flooding higher than 220 m above the Columbia River, moreover, can be traced from below Wenatchee (off map to south) upvalley to the glacial limit but not beyond. These relations indicate that the largest floods predated the blocking of the Columbia River by Cordilleran ice.

The highest three flood-swept surfaces—the surface of the 'great' terrace and two higher bars—are overlain by the 11,250 yr.-B.P. Glacier Peak tephra (layer B of Porter, 1978). The flood deposits capping the 'great' terrace and the higher bars therefore predate the tephra. Two sets of flood bars below the 'great' terrace, however, are barren of the Glacier Peak tephra and must postdate the tephra. The lower bars therefore were produced by relatively young discrete floods, not by the waning phase of the larger flood(s).

Glacial drift and related deposits

Deposits related to alpine glaciers-Glacial drift, which on slopes consists largely of till, and on valley floors consists of outwash gravel grading downvalley from moraines, dominates the broader segments of the Wenatchee drainage basin near Lake Wenatchee and near Leavenworth and parts of the Entiat drainage basin. Most of these deposits have ill-developed soils and sparsely weathered surface stones, and therefore date from the last glaciation. Tracts beyond and above the last-glacial limits near Lake Wenatchee and Leavenworth are underlain by older weathered drift. Nested lateral moraines that become successively more weathered and eroded upslope reveal as many as three separate glaciations of Icicle Creek (Page, 1939b; Porter, 1969; Waitt, 1977b). I have deciphered a similar threefold glacial sequence in the upper Wenatchee drainage basin (see also Nimick, 1977). These sequences are similar to the sequence in the Peshastin and Yakima drainages off the map to the south (Hopkins, 1966; Porter, 1976; Waitt, 1979; Waitt in Tabor and others, 1982). Although these deposits are not dated within the Chelan quadrangle, the last two alpine glaciations of the region probably occurred between 150,000 and 130,000 years ago and between 20,000 and 13,000 years ago (Porter, 1976; Waitt, 1979). During the last glaciation as many as four recessional moraines formed behind the terminal moraine.

Deposits related to Cordilleran icesheet-The lower Chelan basin contains a thick, complex glaciogenic accumulation ranging from boulder till to lacustrine silt that was deposited by the Okanogan lobe of Cordilleran ice, which during the last glaciation ascended the lower end of the valley from the east. This glacier merged with contemporaneous Cordilleran ice that invaded over high divides far to the north and descended into and along the Chelan trough (Waitt, 1972; Waitt and Thorson, 1983). At the maximum stand of the Cordilleran icesheet the

surface sloped south-southeastward from altitude 2,600 m at the United States-Canada boundary to below 300 m near Chelan. In the area affected by Cordilleran ice there are no known glacial deposits predating the last glacial episode.

The sharply defined limit of icesheet drift on the northwestern part of the Columbia Plateau (off map to east) descends steeply along the east side of the Columbia River valley, where it merges with a prominent fill terrace, the 'great' terrace of Russell (1900). The conspicuous and persistent 'great' terrace is therefore the outwash train of the Cordilleran icesheet. A similar drift limit and the upper limit of ice-marginal channels slopes down the west side of the Columbia River valley and from there up the lower Chelan valley. Meltwater confined between the Cordilleran icesheet and the high-relief landscape escaped over low parts of the southern divide of the lower Chelan valley and into Columbia River tributaries, which were thus modified into deep, steep-walled coulees (Waters, 1933; Waitt, 1972). After deglaciation, glaciolacustrine sediment accumulated in the lower Chelan valley—the consequence of Lake Brewster in the Columbia River valley behind the highest part of the 'great' terrace at the head of the outwash train (Waitt and Thorson, 1983). The 'great' terrace slopes gently downvalley for many kilometers until below Entiat it becomes too modified by younger catastrophic-flood deposits to be traced farther. The outwash consists of downvalley-fining gravel ranging from moderately sorted boulder gravel at the head of the valley train to well-sorted cobble gravel near Entiat. As high as 215 m above the present river level the 'great' terrace has been swept by catastrophic floods.

Talus and cirque-glacier deposits-Angular rubble accumulated as talus at the base of steep slopes throughout the region. In many places it grades downslope into alluvial fans that in turn grade into stream alluvium. In many cirques, crested moraines or rock glaciers largely of angular gravel-size fragments are lithologically identical to talus deposits. The age of some of these bouldery accumulations can be bracketed by tephrochronology and ¹⁴C dating. The Mazama ash (about 6,900 yr. B.P.) extends upvalley to the cirque floors, which indicates that the cirques have been deglaciated since before that time. Southeast of Glacier Peak, even Glacier Peak tephra (11,250 m.y. old, Mehringer and others, 1984; layer G of Porter, 1978) blankets the cirque floors, which indicates that the cirques were deglaciated before 13,000 yr. B.P. (Waitt and others, 1982).

Two sets of moraines occur on or above many cirque floors, an older moraine predating the 6,900-yr.-B.P. Mazama ash and younger moraines postdating the 450-yr. B.P. Mount St. Helens Wn tephra. Weathering and lichen data suggest that the older moraine is early Holocene or possibly very late Wisconsin age, and that the younger moraines are less than a century old. This sequence is dissimilar to that reported at large glaciers on Mount Rainier volcano and elsewhere in the western Cascade Range, but is consistent with reinterpreted sequences of small-glacier moraines in several parts of the Pacific Northwest (Waitt and others, 1982). In addition, many cirques in and just north of the Chelan quadrangle have moraine sets overlain by the Yn (3,400 yr. B.P.) tephra but not by the Mazama ash (6,900 yr. B.P.) or moraines overlain by the Wn (450 yr. B.P.) tephra but not by the Yn (R. B. Waitt and P. T. Davis, unpub. data).

Alluvium

Alluvium constitutes a continuum from poorly sorted gravelly sand or sandy gravel of steep fans, to well-sorted pebble gravel and cobble gravel of trunk rivers. The fans are common not only beyond the glacial limit, but also constitute most of the lower slopes of glacially eroded troughs at the northern and western map boundary. Four successive tephra layers- Glacier Peak layer B (about 11,250 yr. B.P.), Mazama ash (about 6,900 yr. B.P.), Mount St. Helens layer Yn (3,400 yr. B.P.) and Mount St. Helens layer Wn (450 yr. B.P.)-commonly are all deposited near the surface of fans in the northwestern part of the map area, whereas along the Columbia River in the east, the Glacier Peak and Mazama tephras generally occur well below the surface of the fan and are separated from each other by considerable sediment. From these relations it appears that most slope material accumulated along the sides of the forested glacial troughs soon after deglaciation, before vegetation became fully established, whereas in the sparsely vegetated eastern areas, fans aggraded at similar rates throughout late Wisconsin and Holocene times.

Landslide Deposits

Landslide deposits, recognized by their angular nonsorted locally derived clasts, their hummocky topography, and by an arcuate scarp upslope, are widespread throughout the quadrangle. Individual slides range in area from a few tens of square meters to a square kilometer, in form from hummocky to subdued, and in age from modern to preglacial.

Slides are particularly abundant and voluminous along the regional escarpment of the Columbia River Basalt Group, where coherent columnar-jointed basalt overlies weakly lithified sandstone and shale of the Ellensburg and Wenatchee Formations. Water that percolates readily through the columnar-jointed basalt accumulates in the less permeable sedimentary beds, further weakening coherence of the sedimentary beds. Because the thick, resistant, permeable caprock of basalt insures steep erosional slopes—generally a scarp—the topography as well as the stratigraphy encourages backwasting of slopes by landsliding.

A second concentration of landslides occurs within the outcrop area of the Chumstick Formation, whose sandstone and shale are less coherent than most of the crystalline rocks of the quadrangle. The slides are not particularly concentrated at any stratigraphic horizon or structural setting within the Chumstick Formation, but they are about twice as abundant southeast of the glacial limit as behind it, which suggests that half of the slides in the nonglaciated area predate the last two glaciations.

A third area of voluminous slides is Wenatchee Ridge, where weakly coherent talc-tremolite schist and serpentinitized ultramafite crop out on steep glaciated slopes. In this area most landslides postdate glaciation.

Incipient blockslides are deeply crevassed and fractured bedrock areas. The largest incipient blockslide, some 4 km² in area, is in the northern Entiat Mountains, a region of coherent crystalline bedrock. This huge block, although situated along a broad divide area of generally moderate slopes, has slid in various directions toward

incised glaciated valleys and the Entiat escarpment. The occurrence of small linear scarps throughout the area suggests that this sliding has occurred largely since the latest glaciation (since about 13,000 yr. B.P.).

Potentially hazardous incipient blockslides

Two incipient blockslides perched on steep slopes 1,000 m above water bodies—one above Lake Chelan, another above Lake Wenatchee—could be severe hazards during future large earthquakes. Although both of these incipient slides may have been in their present form and positions during the largest historic earthquake of the region (in 1872), it is only a matter of time before they will fail and descend to the lakes either gradually or swiftly. The existence of ancient slide deposits in Lake Chelan is suggested by the lake-bottom topography (Whetten, 1967) and the narrowing of the lake (see cross section B-B', map sheet). During the 1872 earthquake a small slide 6 km north of Entiat swept into the Columbia River (Russell, 1900, p. 202). The Columbia is now a series of reservoirs, and any future slides will descend into lake water, where displacement could be locally devastating. Should one of the incipient blockslides above Lake Chelan or Lake Wenatchee suddenly detach, it would probably acquire great speed and momentum on its descent. When the slide enters a lake, water would be suddenly displaced to generate a wave that could devastate the shoreline area for many meters if not tens of meters above lake level. Water thus catastrophically displaced by landslides has devastated shoreline areas in Norway, Japan, and Alaska (Miller, 1960), and at Vaiont Reservoir, Italy (Kiersch, 1964).

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Table 1.—Summary of fission track (FT) and isotope analysis of rocks in the Chelan Quadrangle and vicinity

[For analytical data see Tables 2, 3, and 4; all isotopic ages calculated or converted with decay constants recommended by the IUGS Subcommission on Geochronology (Steiger and Jäger, 1977). K-Ar ages from Engels and others (1976) and earlier reports corrected by use of table in Dalrymple (1979). Uranium-thorium-lead isotope ages reported in the following order: $^{206}\text{Pb}/^{238}\text{U}$;

$^{207}\text{Pb}/^{235}\text{U}$, $^{207}\text{Pb}/^{206}\text{Pb}$; $^{208}\text{Pb}/^{232}\text{Th}$. Multiple data sets are reported preceded by the mesh size or identification of fraction in parenthesis. For Mattinson samples: cg = coarse grained; mg = medium-grained; fg = fine-grained; m = magnetic; nm = non-magnetic. Map numbers refer to locations shown on the map except where noted under comments.]

Map no.	Field sample no.	Method	Mineral	Location		Unit or formation	Age (m.y.)	Comments	Source
				Lat N.	Long W.				
1	RWT 141-77	K-Ar	Hornblende	47°35.3'	120°19.1'	Burch Mountain unit	10.0 ± 0.3		Table 2
2	JE 17-67	K-Ar	Hornblende	47°33.0'	120°22.2'	do.	11.1 ± 1.7		Engels and others (1976)
3	VF 79-629	FT	Zircon	47°59.1'	120°42.6'	Volcanic rocks of Chickamin Creek	23.0 ± 0.7	rhyolite breccia	Table 3
4	78-239	FT	Zircon	47°58.7'	120°43.3'	do.	27.1 ± 1.3		C. W. Naeser (written comm. 1981)
		FT	Apatite				28.9 ± 5.1		Table 2
5	75-207	K-Ar	Hornblende	47°38.9'	120°36.8'	Dike cutting Chumstick Formation	41.5 ± 2.6		
6	78-226	FT	Apatite	47°33.9'	120°25.3'	Chumstick Formation	38.3 ± 6.7	detrital grains	C. W. Naeser (written comm. 1981)
7	78-227	FT	Zircon	47°32.4'	120°26.7'	do.	45.4 ± 2.0	do.	do.
		FT	Apatite				32.1 ± 5.1	do.	do.
8	78-235	FT	Zircon	47°39.4'	120°31.0'	do.	46.6 ± 2.4	do.	do.
9	78-236	FT	Zircon	47°44.9'	120°38.4'	do.	47.5 ± 2.3	do.	do.
		FT	Apatite				34.7 ± 6.1	do.	do.
10	78-237	FT	Apatite	47°49.9'	120°46.7'	do.	47.7 ± 7.7	do.	do.
11	78-232	FT	Zircon	47°34.9'	120°37.0'	do.	50.2 ± 2.4	do.	do.
		FT	Apatite				29.0 ± 4.8	do.	do.
12	78-228	FT	Zircon	47°30.1'	120°26.0'	do.	50.6 ± 2.2	do.	do.
		FT	Apatite				38.8 ± 6.3	do.	do.
13	78-233	FT	Zircon	47°39.1'	120°34.2'	do.	56.6 ± 5.2	do.	do.
14	78-242	FT	Zircon	47°50.6'	120°39.9'	do.	66.4 ± 3.9	do.	do.
		FT	Apatite				43.9 ± 7.4	do.	do.
15	75-211	FT	Zircon	47°40.4'	120°36.6'	Tuff in Chumstick Formation	41.9 ± 6.8	"zeolite tuff"	Gresens and others (1981)
16	75-209	FT	Zircon	47°38.2'	120°36.2'	do.	42.7 ± 5.1	do.	do.
17	75-202	FT	Zircon	47°37.2'	120°35.2'	do.	48.8 ± 7.2	do.	do.
18	75-200	FT	Zircon	47°38.5'	120°32.5'	do.	42.7 ± 1.5	"Eagle Creek tuff"	do.
19	75-124C	FT	Zircon	47°40.7'	120°34.8'	do.	46.1 ± 1.9	do.	do.
20	75-x	FT	Zircon	47°40.1'	120°34.0'	do.	46.2 ± 1.8	do.	do.
21	75-191	FT	Zircon	47°40.8'	120°36.0'	do.	46.4 ± 1.9	unnamed tuff	do.
22	75-215	FT	Zircon	47°41.5'	120°36.1'	do.	42.7 ± 3.7	do.	do.
23	76-67	FT	Zircon	47°30.4'	120°29.0'	do.	44.4 ± 2.6	do.	do.
24	TLW 252	Pb-a	Zircon	48°00.6'	120°34.4'	Duncan Hill pluton	40.0 ± 10.0	just north of quad	Engels and others (1976)
25	RWT 221-80	U-Th-Pb	Zircon	48°06.5'	120°41.4'	do.	(-150 + 200)45.8; 45.8; 47.8; 47.1 (-250 + 325)46.4; 46.5; 49.7; 45.3	Borealis Ridge, north of quad	Table 4
26	FWC252B	K-Ar K-Ar K-Ar	Hornblende Hornblende Biotite	47°56.7'	120°31.2'	do.	48.3 ± 4.5 48.3 ± 5.0 43.0 ± 2.2		Table 2 Cater and Crowder (1967)
27	C-685-1	K-Ar	Hornblende	47°56.3'	120°31.2'	do.	44.9 ± 1.8		do.
28	C-541-1	K-Ar	Biotite	47°54.0'	120°21.1'	do.	46.2 ± 1.4		do.
29	RWT 32-79	K-Ar	Hornblende	47°59.6'	120°09.9'	Granite porphyry dike swarm	44.2 ± 2.0		Table 2
30	RWT 455-78	K-Ar K-Ar	Biotite Biotite	47°53.5'	120°34.7'	do.	46.9 ± 0.6 47.0 ± 0.7		do.
31	RWT 40-79	K-Ar	Biotite	48°01.4'	120°11.5'	Cooper Mountain pluton dike	48.1 ± 4.5	just off quad	do.
32	RWT 91-79	K-Ar	Hornblende	47°56.7'	119°59.9'	do.	48.1 ± 1.3	cuts Chelan Complex just off quad	do.
33	KA-4	K-Ar	Biotite	47°38.1'	120°10.6'	Porphyritic dacite dike	47.8 ± 1.9	Corbaley Canyon swarm	R. Gresens (written comm., 1978)
34	C101	K-Ar	Hornblende	47°38.2'	120°10.2'	Hornblende gabbro dike	48.4 ± 2.2	do.	do.
35	RWT 21-82	K-Ar	Muscovite	47°50.3'	120°50.5'	Light-colored gneiss of Wenatchee Ridge	80.8 ± 1.6	pegmatite	Table 2
	RWT 23-82	K-Ar	Biotite	47°50.3'	120°50.5'	do.	81.2 ± 1.0	do.	do.
		K-Ar	Muscovite				83.7 ± 1.1		
36	RWT 16-82	K-Ar	Biotite	47°36.2'	120°53.0'	Chiwaukum Schist	85.3 ± 1.4	mica schist	do.
37	RWT 483-79	K-Ar K-Ar	Biotite Muscovite	47°41.1'	120°45.2'	do.	83.5 ± 0.7 86.0 ± 0.6	do.	do.
38	RWT 183-77	FT K-Ar K-Ar U-Th-Pb	Apatite Hornblende Biotite Zircon	47°46.9'	120°01.6'	Chelan Complex of Hopson and Mattinson (1971)	54.7 ± 1.6 63.9 ± 1.2 58.4 ± 0.6 (-100 + 150)78.7; 78.7; 80.5; 77.1 (-200 + 250)77.1; 77.3; 84.6; 77.2	mafic tonalite unit	Table 3 do. Table 4
39	RWT 84-79	K-Ar	Biotite	47°57.7'	120°01.6'	do.	74.3 ± 1.0	tonalite unit	Table 2
40	68-1	U-Th-Pb	Zircon	47°51.7'	120°09.3'	do.	(cg) 131; ---; ---; --- (mg) ---; ---; 175 ± 10; --- (fg) 110; ---; ---; ---	do.	Mattinson (1972)
41	C-724-1	K-Ar	Hornblende	47°50.4'	120°12.2'	Migmatite of Chelan Complex	84.3 ± 2.5 62.0 ± 2.0		Engels and others (1976)
42	68-2	K-Ar	Biotite	47°45.7'	120°05.9'	do.	120; 121; 127 ± 10; --- 70; ---; ---; ---	leucosome	Mattinson (1972)
43	68-4	U-Pb	Zircon	47°49.0'	119°59.0'	do.	112; 111; 100 ± 25; --- 81; ---; ---; ---	do.; just off quad to east	do.
44	68-5	U-Pb	Zircon	47°48.8'	119°58.1'	do.	(m) 99; ---; ---; --- (nm) 100; ---; ---; --- (mg) ---; ---; 104 ± 15; --- 86; ---; ---; ---	do.; just off quad to east	do.
45	C-679-1	U-Pb K-Ar	Sphene Hornblende	47°46.6'	120°08.7'	do.	69.4 ± 2.0		Engels and others (1976)

Table 1—(continued)

Map no.	Field sample no.	Method	Mineral	Location		Unit or formation	Age (m.y.)	Comments	Source
				Lat N.	Lat W.				
46	78-249	FT FT	Zircon Apatite	47°42.5'	120°13.0'	Entiat pluton	57.6 ± 4.1 50.2 ± 7.7	gneissic tonalite	C. W. Naeser (written comm., 1981)
47	78-241	FT	Zircon	47°58.8'	120°42.6'	do.	32.8 ± 1.8	do.	do.
48	C-677-1	K-Ar	Hornblende	47°42.8'	120°13.2'	do.	65.9 ± 1.9	do.	Engels and others (1976)
49	RWT 151-79	K-Ar	Biotite	47°48.2'	120°26.2'	do.	62.8 ± 1.8	massive tonalite	Table 3
		FT	Apatite				48.1 ± 1.2		
50	RWT 123-79	K-Ar	Hornblende	47°38.6'	120°09.7'	do.	73.2 ± 1.5	flaser gneiss	Table 2
		K-Ar	Biotite				60.3 ± 0.8		
		K-Ar	Hornblende				(-100 + 150)/76.5; 77.1; 96.6; 84.2		
		FT	Apatite				(-200 + 250)/74.5; 75.4; 102.5; 90.0		
		U-Th-Pb	Zircon				60.0 ± 1.2		
51	69-3	U-Pb	Zircon	47°51.8'	120°28.2'	Banded migmatitic tonalite gneiss and mafic amphibolite	56.2 ± 0.8	Mattinson (1972)	
							50.4 ± 1.2		
							(-100 + 150)/72.8; 76.0; 178.2; 72.0		
							(-200 + 250)/73.8; 79.7; 259.7; 79.1		
							181; 182; 185 ± 10; ---		
52	JE 21-67	K-Ar	Amphibole	47°32.7'	120°44.3'	Mount Stuart batholith	95.0 ± 10.7	Eastern pluton	Engels and Crowder (1971)
		K-Ar	Hornblende				90.0 ± 2.7		
		K-Ar	Biotite				90.0 ± 2.6		
53	JE 22-67	FT	Apatite	47°32.8'	120°41.8'	do.	65 ± 4	do.	do.
		K-Ar	Biotite				87.7 ± 2.6		
		K-Ar	Hornblende				90.0 ± 2.9		
54	JE 20-67	FT	Apatite	47°33.7'	120°46.6'	do.	57 ± 4	do.	do.
		K-Ar	Biotite				94 ± 9		
		K-Ar	Hornblende				87.7 ± 2.6		
55	JH 79-191	FT	Allanite	47°39.6'	120°49.1'	do.	93.2 ± 3.1	pluton of Big Jim Mtn.	Table 4
		K-Ar	Biotite				71 ± 5		
		K-Ar	Hornblende				(-100 + 150)/95.0; 95.0; 96.3; 95.5		
							(-200 + 250)/95.9; 95.9; 97.2; 95.9		
56	JE 16-67	K-Ar	Biotite	47°57.2'	120°56.2'	Ten Peak pluton	77.3 ± 2.4		Engels and others (1976)
		K-Ar	Hornblende				92.8 ± 3.1		
57	78-243	FT	Zircon	47°39.0'	120°27.0'	Swakane Biotite	36.6 ± 2.3		C. W. Naeser (written comm., 1981)
58	78-244	FT	Zircon	47°38.4'	120°26.6'	Gneiss	37.8 ± 2.0		
59	78-246	FT	Zircon	47°35.4'	120°22.7'	do.	40.5 ± 2.1		do.
60	78-245	FT	Zircon	47°36.0'	120°24.8'	do.	42.3 ± 8.2		do.
61	78-248	FT	Zircon	47°32.7'	120°17.4'	do.	43.9 ± 2.5		do.
62	78-247	FT	Zircon	47°33.7'	120°19.4'	do.	44.8 ± 2.3		do.
63	78-225	FT	Zircon	47°31.9'	120°27.1'	do.	46.4 ± 3.0		do.
64	78-234	FT	Zircon	47°39.2'	120°33.9'	do.	52.9 ± 2.4		do.
65	69-4	U-Pb	Zircon	47°35.8'	120°14.9'	Swakane Biotite	68; ---; ---; ---	pegmatite	Mattinson (1972)
66	68-8	U-Pb	Zircon	47°32.2'	120°17.9'	Gneiss			do.
							(cg) 421; 611; 1398 ± 10; --		
							(fg) 400; 612; 1486 ± 10; --		
							(m) 467; 638; 1304 ± 10; --		
67	RWT 3-81	U-Th-Pb	Zircon	47°36.5'	120°14.5'	do.	(+250)/312.8; 488.3; 1427.0; 344.3	amphib. layer	Table 4
		K-Ar	Hornblende				(-250)/315.6; 499.5; 1464.5; 329.1		
							50.8 ± 1.4		
68	RWT 203-80	U-Th-Pb	Zircon	47°56.3'	120°49.9'	Mad River terrane	(-150 + 200)/52.1; 53.0; 91.4; 45.1	metagabbro	Table 4
		U-Th-Pb	Zircon				(-250 + 325)/50.1; 50.9; 88.4; 48.9		
							(-400)/51.4; 52.4; 99.8; 47.4		
69	RWT 9-81	K-Ar	Hornblende	47°44.2'	120°27.9'	Heterogeneous schist and gneiss unit	67.1 ± 3.0	amphibolite layer	Table 2
70	RWT 39-80	K-Ar	Hornblende	47°43.9'	120°29.1'	do.	63.0 ± 3.5	do.	do.
71	RWT 8-81	K-Ar	Hornblende	47°44.2'	120°26.7'	do.	71.5 ± 1.4	do.	do.
72	RWT 201-80	K-Ar	Muscovite	47°56.3'	120°49.8'	Rocks of the Napeequa River area	52.4 ± 0.9	zoisite amphib.; north of quad.	do.
		K-Ar	Hornblende				54.9 ± 4.0		
73	VF 80-179	K-Ar	Biotite	47°38.9'	120°13.9'	Light-colored gneiss unit	53.2 ± 1.0		Table 2
	U-Th-Pb	Zircon	(-150 + 250)/77.2; 78.5; 104.2; 90.9						
	VF 81-1091	U-Th-Pb	Zircon	do.	do.	do.	(-150 + 200)/75.9; 77.5; 127.0; 75.1		do.
							(-250 + 325)/76.4; 76.9; 92.3; 74.3		
							(-400)/73.5; 73.6; 75.5; 71.3		

Table 2.—Data for new ages obtained by the K-Ar method for this report

[All USGS K-Ar ages calculated on the basis of 1976 IUGS decay and abundance constants; errors based on variation in replicate K₂O and argon analyses or expected variation from an empirically derived curve relating coefficient of variation in the age to percent radiogenic argon (Tabor and others, 1985) K₂O was determined by flame photometry by analysts P. Klock, B. Lai, S. Neil, J. Marinenko, and D. Vivit]

Map number	Sample number	Mineral	K ₂ O percent	Ar ⁴⁰ Rad moles/gm x 10 ⁻¹⁰	Ar ⁴⁰ Rad percent	Age (m. y.)
1	RWT 141-77	Hornblende	0.184, 0.184, 0.178, 0.180	0.0251, 0.0263	5.8, 3.5	10.0 ± 0.3
25	RWT 221-80	Hornblende	0.304, 0.300, 0.304, 0.298	0.212	12.9	48.3 ± 5.3
29	RWT 32-79	Hornblende	0.763, 0.750, 0.762, 0.757	0.488	33.1	44.2 ± 3.9
		Biotite	8.47, 8.57	5.83	89.9	46.9 ± 0.5
30	RWT 455-78	Biotite	7.92, 7.92	5.43	87.3	47.0 ± 0.4
31	RWT 40-79	Biotite	9.13, 9.14	6.41	17.4	48.1 ± 5.4
32	RWT 91-79	Hornblende	0.769, 0.769, 0.762, 0.768	0.538	53.6	48.1 ± 2.4
35	RWT 21-82	Muscovite	10.63, 10.62	12.649	74.4	80.8 ± 2.4
	RWT 23-82	Biotite	8.96, 9.01	10.745	92.3	81.2 ± 1.6
		Muscovite	10.44, 10.54	12.942	91.0	83.7 ± 1.7
36	RWT 16-82	Biotite	8.45, 8.46	10.639	86.3	85.3 ± 1.4
37	RWT 483-79	Biotite	8.03, 8.03	9.96, 9.80	80.6, 78.2	83.5 ± 0.7
		Muscovite	8.30	10.53	47.6	86.0 ± 3.4
38	RWT 183-77	Hornblende	1.040, 1.019, 1.027, 1.037	0.0966	75.7	63.9 ± 1.5
		Biotite	7.84, 7.86	6.54, 6.66	86.1, 89.1	58.4 ± 0.6
39	RWT 84-79	Biotite	9.22, 9.24	10.1	89.1	74.3 ± 0.6
49	RWT 151-79	Hornblende	1.222, 1.242, 1.210, 1.222	1.32	69.0	73.2 ± 2.3
		Biotite	9.83, 9.71	8.62	90.0	60.3 ± 0.6
50	RWT 123-79	Hornblende	1.287, 1.288, 1.287, 1.286	1.13	73.0	60.0 ± 1.4
		Biotite	8.71, 8.74	7.17	86.6	56.2 ± 0.4
67	RWT 3-81	Hornblende	0.756, 0.788, 0.786, 0.790	0.585	56.5	50.8 ± 1.5
69	RWT 9-81	Hornblende	0.467, 0.466, 0.467, 0.467	0.459	32.9	67.1 ± 3.0
70	RWT 39-80	Hornblende	0.401, 0.396, 0.399, 0.396	0.367	23.9	63.0 ± 6.5
71	RWT 8-81	Hornblende	1.562, 1.560	1.639	78.2	71.5 ± 1.4
72	RWT 201-80	Muscovite	9.59, 9.60	7.337	79.3	52.4 ± 1.2
		Hornblende	0.250, 0.248, 0.276, 0.268	0.209	17.5	54.9 ± 5.6
73	VF 80-179	Biotite	7.83, 7.88	6.11	75.3	53.2 ± 1.3

Table 3.—Data for new zircon and apatite fission track ages of this report
[all fission track ages calculated with $F = 7.03 \times 10^{17} \text{ yr}^{-1}$]

Map number	Sample number	Mineral	$\rho_s \times 10^6/\text{cm}^2$	$\rho_l \times 10^6/\text{cm}^2$	$\phi_x \times 10^{15}$ neutrons/cm ²	Age (m. y.)	U (ppm)	Grains counted
3	VF 79-629	Zircon	1.91	4.68	0.94	23.0 ± 0.7	150	7
38	RWT 183-77	Apatite	1.07	1.24	1.06	54.7 ± 1.6	35	100
49	RWT 151-79	Apatite	1.46	1.90	1.05	48.1 ± 1.2	55	100
50	RWT 123-79	Apatite	1.73	2.10	1.03	50.4 ± 1.2	61	100

Table 4.—Uranium-thorium-lead isotopic ages of zircon from rocks of the Northern Cascades, Washington

Constants: $^{238}\text{U} = 1.55125 \times 10^{-10} \text{ yr}^{-1}$; $^{235}\text{U} = 9.8485 \times 10^{-10} \text{ yr}^{-1}$; $^{232}\text{Th} = 4.9475 \times 10^{-11} \text{ yr}^{-1}$; $^{238}\text{U}/^{235}\text{U} = 137.88$.

Isotopic composition of common lead assumed to be $^{204}\text{Pb}:^{206}\text{Pb}:^{207}\text{Pb}:^{208}\text{Pb} = 1:18.60:15.60:38.40$

Map number	Mesh size	Isotopic composition of lead								Age, in millions of years			
		Concentration (ppm)			(atom percent)				^{206}Pb	^{207}Pb	^{207}Pb	^{208}Pb	
		U	Th	Pb	^{204}Pb	^{206}Pb	^{207}Pb	^{208}Pb	^{206}Pb ^{238}U	^{207}Pb ^{235}U	^{207}Pb ^{206}Pb	^{208}Pb ^{232}Th	
25		RWT 221–80 Duncan Hill pluton (gneissic tonalite of Borealis Ridge)											
	–150+200	399.8	125.6	2.937	0.0513	84.38	4.719	10.85	45.8	45.8	47.8	47.1	
	–250+325	426.8	140.9	3.216	0.0658	83.75	4.906	11.28	46.4	46.5	49.7	45.3	
38		RWT 183–77 Tonalite of the Chelan Complex of Hopson and Mattinson (1971)											
	–100+150	145.6	54.3	1.947	0.1045	81.01	5.396	13.49	78.7	78.8	80.5	77.1	
	–200+250	165.6	58.8	2.148	0.0966	81.57	5.313	13.02	77.1	77.3	84.6	77.2	
49		RWT 151–79 Entiat pluton											
	–100+150	286.4	125.8	3.845	0.1043	78.45	5.296	16.15	76.5	77.1	96.6	84.2	
	–200+250	307.6	155.0	4.067	0.0754	77.10	4.815	18.01	74.5	75.4	102.5	90.0	
50		RWT 123–79 Entiat pluton (flaser gneiss)											
	–100+150	266.1	174.8	3.851	0.1991	71.29	6.462	22.05	72.8	76.0	178.2	72.0	
	–200+250	320.5	145.4	4.993	0.3119	69.41	8.138	22.14	73.8	79.7	259.7	79.1	
55		JH 79–191 Mt. Stuart batholith (mafic pluton of Big Jim Mtn.)											
	–100+150	234.8	92.3	3.683	0.0563	82.45	4.781	12.71	95.0	95.0	96.3	95.5	
	–200+250	330.6	122.3	5.183	0.0526	83.20	4.764	11.99	95.9	95.9	97.2	95.9	
67		RWT 3–81 Garnet amphibolite in the Swakane Biotite Gneiss											
	+250	402.3	109.4	27.683	0.3603	68.72	11.19	19.73	312.8	488.3	1427.0	344.3	
	–250	410.4	113.5	28.911	0.3765	68.11	11.47	20.04	315.6	499.5	1464.5	329.1	
68		RWT 203–80 Metagabbro of the Mad River terrane											
	–150+200	370.7	134.7	3.100	0.0531	84.46	4.822	10.66	52.1	53.0	91.4	45.1	
	–250+325	422.3	140.4	3.365	0.0369	84.97	4.603	10.39	50.1	50.9	88.4	48.9	
	–400	457.9	162.4	3.862	0.0767	83.08	5.119	11.73	51.4	52.4	99.8	47.4	
73		VF 80–179 Light-colored gneiss of the Mad River terrane											
	–150+250	727.4	77.8	8.454	0.0313	90.28	4.803	4.885	77.7	78.5	104.2	90.9	
73		VF 81–1091 Light-colored gneiss of the Mad River terrane											
	–150+200	619.8	109.6	7.177	0.0373	88.63	4.853	6.479	75.9	77.5	127.0	75.1	
	–250+325	576.4	88.9	6.671	0.0381	89.31	4.835	5.822	76.4	76.9	92.3	74.3	
	–400	616.6	72.3	6.936	0.0665	88.91	5.204	5.823	73.5	73.6	75.5	71.3	