

LATE MESOZOIC OR EARLY TERTIARY MELANGES IN THE WESTERN CASCADES OF WASHINGTON

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ABSTRACT

Marine rocks west of the Straight Creek fault, from Snoqualmie Pass to the Pilchuck River, are highly deformed and varied in lithology. These rocks, referred to various formations or stratigraphic units by earlier workers, should be considered two lithologically distinct melange belts.

Immediately west of the Straight Creek fault from the Weden Creek area to Snoqualmie Pass, the discontinuous melange of the eastern belt consists of strongly hornfelsed rocks including chert, metabasalt, graywacke, argillite, marble, and ultramafic rocks. Marble near Skykomish yields Permian fusulinids with Tethyan affinities. Zircons from tonalitic gneiss northwest of Skykomish are 190 million years old, and metagabbro near Snoqualmie Pass is 165 million years old.

In the western foothills, the melange of the western belt consists of pervasively sheared argillite and graywacke studded with outcrop- to mountain-scale steep-sided phacoids of chert, limestone, pillow basalt, gabbro, and tonalite; ultramafic rocks are rare. Strikes of bedding and cleavage and trends of small fold axes are mostly north-northwest to north-northeast. Rare macrofossils in foliated sedimentary rocks and radiolarians in chert phacoids are Late Jurassic to Early Cretaceous in age. Tonalite associated with gabbro and diabase yields zircon U-Th-Pb ages of 150 to 170 million years in the Woods Creek area, at Mount Si, at Winter Lake, and on the North Fork of the Snoqualmie River. The general Late Jurassic age of both igneous and marine sedimentary components of this western belt suggests that most of these rocks were derived from a contemporaneous, if not common, oceanic sequence; the metaplutonic rocks therein may represent the oceanic basement of the sedimentary rocks. Although we cannot rule out the possibility that the isolated tectonic blocks of metatonalite and metagabbro in the melanges are klippen of the proposed Haystack thrust to the northwest, we think that imbrication of the oceanic basement with its sedimentary and volcanic cover during formation of the melanges is a more probable origin for the blocks.

The melange belts probably were formed by accretion, perhaps after the widespread Late Cretaceous metamorphic event in the Cascades to the east, but before the Eocene. Post-Late Cretaceous right-lateral strike-slip faulting along the Straight Creek fault has moved the rocks to their present position.

INTRODUCTION

Pre-Tertiary rocks in the central and northern Cascade Range of Washington consist of weakly metamorphosed to unmetamorphosed Paleozoic and Mesozoic sedimentary and igneous rocks west of the Straight Creek fault and medium- to high-grade metamorphic rocks east of the fault (Fig. 1). Some of the medium- to high-grade rocks east of the fault are described in a companion article (Tabor and others, this volume). We describe here two lithologically distinct belts of pervasively deformed marine rocks that

crop out west of the fault more or less between the Pilchuck River and the South Fork of the Snoqualmie River.

Previously, many of these rocks have been considered to be conventional stratified formations of sedimentary rocks and assigned a variety of formation names and ages. (See table 1 of Dungan, 1974.) Because of the intense penetrative deformation and scattering of exotic components in the rocks, we consider them to be melanges. They appear to constitute two melange belts: an eastern belt that contains pre-

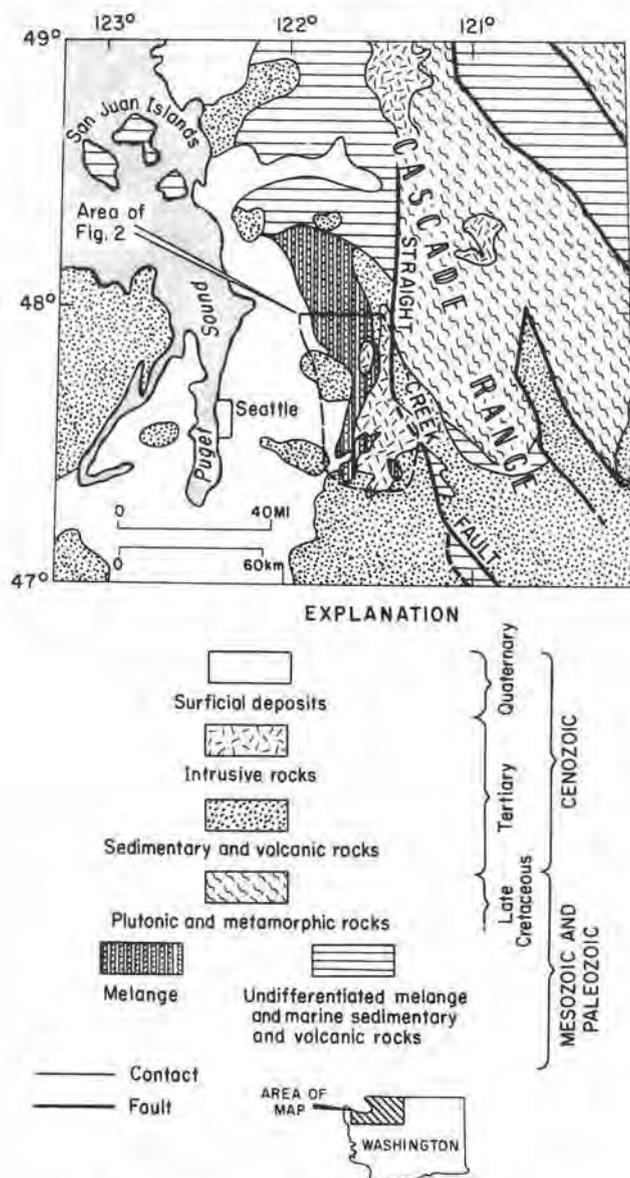


Figure 1.—Simplified geologic map of northwestern Washington, showing location of melange belts in relation to other major rock groups and structures.

dominantly chert and volcanic materials, and a western belt that contains predominantly clastic materials. Two recent studies (Fox, 1983; Jones and others, 1978) describing the distribution and composition of melanges on the West Coast omitted these rocks, even though they underlie a significant area in western Washington. In a previous report (Frizzell and others, 1982), we indicated that the melange of the eastern belt contained both Paleozoic and Mesozoic components, whereas the melange of the western belt contained mostly Late Jurassic and Early Cretaceous components. More recent field work and new fossil

data suggest that both belts contain Paleozoic and Mesozoic components.

Numerous workers, primarily students from the University of Washington, have investigated these and similar rocks; their endeavors, summarized briefly below, were of great value to us. Because much remains to be learned about the lithologies, ages, and relations of the rocks in the melanges, we refrain from using the formation names assigned by earlier workers, except when referring to work of others or correlations made by others with rocks outside the area of this report. Although Silberling and others (1984, p. C-22 - C-27) have correlated the melanges of the eastern and western belts with regional terranes called the San Juan and Olney Pass terranes, respectively, mapping and correlation of the melanges with regional terranes are still in progress. The origin of these disrupted oceanic Paleozoic and Mesozoic rocks and their subsequent accretion to the North American continent have been considered by many students of Northwest geology (for instance, Danner, 1977; Davis and others, 1978; Coney and others, 1980; Vance and others, 1980; Whetten and others, 1980).

We describe the melanges in two formats. The general description of early work, naming and correlations, overall structure, and age are discussed in the text under "Description of Melange Belts." More detailed lithologic descriptions of melange components accompany Figure 2. New radiometric ages are given in Appendix Tables A1 and A2.

"DEFINITION" OF MELANGE

The term "melange", first applied on the West Coast by Bailey (1941, p. 30) to the Franciscan assemblage "because of extreme contortion of the structures as well as the variability of the rock types," was popularized in the United States by Hsü (1968), whose work, also in the Franciscan, precipitated much discussion. Although attempts have been made to formally define the term "melange" (for example, Berkland and others, 1972, p. 2296), participants at the 1978 Penrose Conference on Melanges tried to arrive at a single comprehensive definition without complete success. However, the usage that grew from those discussions (Silver and Buetner, 1980, p. 32 [after Hsü, 1968]) is applicable to rocks discussed in this paper.

"Melange" is a general term describing a mappable (at 1:25,000 or smaller scale), internally fragmented and mixed rock body containing a variety of blocks, commonly in a pervasively deformed matrix. The term refers to rock mixtures formed by tectonic movements, sedimentary sliding,

or any combination of such processes, with no mixing process excluded. It does not imply a particular mixing process. 'Melange' is a handy descriptive term, alerting the reader to the possibility that such a body may not conform to classic principles of superposition and stratigraphic succession [our emphasis]."

We would add to this general definition that exotic blocks—tectonic inclusions detached from rocks foreign to the main body of the melange—should be present to distinguish a melange from a broken formation (Hsu, 1968, p. 1065).

DESCRIPTION OF MELANGE BELTS

Melange of the Eastern Belt

The melange of the eastern belt consists predominantly of a mixture of mafic volcanic rocks and chert with minor argillite and graywacke. Marble is conspicuous locally, and large and small pods of ultramafic rocks occur throughout the belt. Mafic migmatitic gneiss, metadiabase, and metatonalite are also present in the eastern belt. The rocks are considerably deformed, slightly metamorphosed (mostly to greenschist facies), and locally overprinted to pyroxene hornfels by Tertiary plutons. Evidence for extreme disruption of the unit is abundant. Structural orientations are highly varied, although the mean strike of bedding and foliation is NNW, and dips range from 40° to 80° to the east or west. The melange of the eastern belt, on the whole, occurs as patchy outcrops in the area of this study. It is exposed as screens or pendants in and among the Tertiary Snoqualmie, Grotto, and Index batholiths (Figs. 1 and 2).

Chert, Argillite, Mafic Volcanic Rocks, and Graywacke

Weaver (1912) first described the metasedimentary rocks in our melange of the eastern belt near the town of Skykomish (Fig. 2); he referred them to his Gunn Peak Formation and correlated them with the upper Paleozoic Cache Creek Group of Dawson (1895) in British Columbia. One of the more troublesome fossil identifications in the literature of the North Cascades followed. On the basis of a single sample from a "cherty phase of the limestone lens outcropping in Lowe Gulch," Smith (1916, p. 562; Table 1, loc. 1) assigned an Ordovician age to Weaver's Gunn Peak Formation (Smith's "Maloney Metamorphic Series"). Although no modern determi-

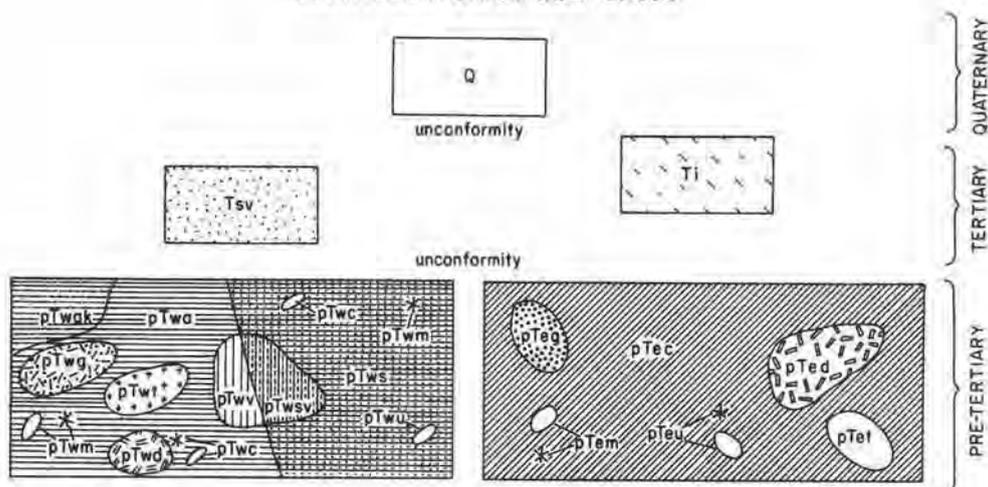
nation is possible because the specimens have apparently been lost, Duror's descriptive report on the brachiopod and trilobite fauna (*in* Smith, 1916, p. 580-582) includes line drawings. Duror concluded from her examination that the fossils were "Trenton age" and had not been recorded in rocks younger than Ordovician. Dedicated sleuthing by W. R. Danner (reported *in* Galster, 1956, p. 20-21) failed to unearth either the specimens originally described by Duror or corroborative samples from Lowe Gulch, but the line drawings indicate that the fossils were correctly identified (J. T. Dutro, Jr., personal commun., 1983).

Danner (1957) was successful, however, in producing the first truly comprehensible regional stratigraphic history of the pre-Tertiary rocks of western Washington. He (1957, p. 230-249) detailed the occurrence of marble (discussed below) in the Skykomish area that he referred to his Stillaguamish Group (Danner, 1957); he has since called these rocks the "Trafton sequence" (Danner, 1966, p. 363). Galster (1956, p. 14-22) mapped rocks southwest of Skykomish, which he also referred to the Stillaguamish (now the Trafton).

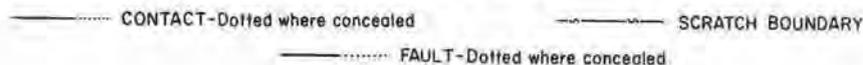
Yeats (1964, p. 556) referred a prominent area of outcrop northwest of Skykomish to his (restricted) Gunn Peak Formation and Barclay Creek Formation. The Gunn Peak Formation, composed predominantly of ribbon chert and quartzite (metachert) and containing marble pods, grades upward into the Barclay Creek Formation, which is similar to the Gunn Peak, but locally contains more banded siliceous and calcareous argillite, graywacke, and greenstone. Yeats (1958a, p. 102-114; 1958b; 1964, p. 556) geographically restricted the Gunn Peak Formation of Weaver (1912, p. 36-38), which had also included the migmatitic gneiss and parts of the Eocene nonmarine Swauk Formation, to those rocks fulfilling Weaver's original description of the unit. Although Tabor and others (1982) were not able to separate the Barclay Creek and Gunn Peak Formations in areas near Yeats' type section, his division appears to hold true to the north in the Stillaguamish River area (for example, Vance, 1957; Baum, 1968). Plummer (1964, p. 23-25) briefly described rocks that we include in the melange of the eastern belt and noted that tectonic fragmentation and shearing occurred prior to static metamorphism.

Included in the melange of the eastern belt in the Weden Creek area are rocks mapped as probable Chilliwack Group of Misch (1966) by Heath (1971, p. 90-97) and thermally metamorphosed volcanic rock and chert associated with ultramafic rocks on the north end of Silver Creek. Also included in the eastern melange belt are metabasalt and ribbon chert exposed on Mount Garfield. A roof pendant consisting mostly

CORRELATION OF MAP UNITS



EXPLANATION



DESCRIPTION OF MAP UNITS

- | | |
|---|---|
| <p> Q Surficial deposits</p> <p> Ti Tertiary intrusive rocks</p> <p> Tsv Tertiary sedimentary and volcanic rocks</p> | <p> pTcd Metadiabase Altered and thermally metamorphosed, fine-grained, ophitic to subophitic metadiabase and metagabbro, containing euhedral plagioclase, mostly urazitized subhedral clinopyroxene, and rare hypersthene. For further descriptions, see Danner (1957, p. 513-517) and Plummer (1964, p. 52-54).</p> <p> pTeg Migmatitic gneiss Fine-grained schistose amphibolite to medium- and coarse-grained massive quartz diorite, including layered hornblende gneiss, gneissose quartz diorite, trondhjemite, replacement breccia, and minor serpentized ultramafite. Mafic and less-mafic rocks commonly occur in irregular, intimately mixed layers. Cut by anastomosing shear zones; rocks cataclastically deformed prior to a late static recrystallization. Description adapted from Yeats (1958a, p. 83-99; 1964, p. 552-555).</p> <p> pTet Metatonalite Altered medium-grained metatonalite, containing subhedral plagioclase, mostly anhedral and intergranular mosaics of quartz and small amounts of altered green hornblende and perthitic potassium feldspar; locally cataclastically deformed.</p> |
|---|---|

Rocks of the eastern melange

 **pTec** **Chert, argillite, mafic volcanic rocks, and graywacke** Intensely folded, white to cream or gray ribbon chert, alternating with thin- to thick-bedded, dark- brown to black calcareous argillite that grades to phyllitic slate and is tectonically thinned and thickened into discontinuous stringers. Chert and argillite are intimately mixed with tectonized greenstone, greenschist (metabasalt and meta-andesite), and metagraywacke. Also includes subordinate chert-rich metagrit, and metaconglomerate that are mostly massive and rarely bedded. Original sedimentary and volcanic textures are largely obscured due to recrystallization by Tertiary plutons. Yeats (1964, p. 555-558; 1958a, p. 106-115) describes the rocks in the Skykomish River area in detail.

-  **pTem** **Marble** Predominantly lenticular beds and pods of banded white to grayish, medium- to fine-grained crystalline marble intercalated with metachert and greenstone. In part with shaly laminations or graphitic impurities marking bedding planes; silicified replacement masses (Danner, 1966, p. 374; Yeats, 1958a, p. 103).
-  **pTew** **Ultramafite** Pods of highly altered pyroxenite, peridotite, and serpentinized dunite tectonically intercalated with argillite and ribbon chert, north of Skykomish (Yeats, 1958a, p. 116-119). Schistose ultramafite east of Weden Creek consists of forsterite in a mesh of antigorite. Dungan (1974, p. 40-41) described peridotite hornfels in the upper Sultan River area.
- Rocks of the western melange**
-  **pTwa** **Argillite and graywacke** Matrix mostly argillite, pervasively sheared, scaly, containing steep-sided outcrop- to mountain-sized phacoids of fine- to coarse-grained and pebbly lithofeldspathic and volcanolithic subquartzose sandstone interbedded with black argillite. The sandstone is a mixed clast type, most clasts being plagioclase, chert, volcanic rocks, and quartz. Where more strongly deformed, unstable grains are broken down into anastomosing shear zones or smeared out into indistinct chloritic matrix. Sedimentary features such as graded bedding and load casts locally are well preserved. The unit includes minor chert, polymictic and quartz-pebble conglomerate, shale-chip breccia, and phacoids of limestone, metavolcanic rock, metagabbro, and metatonalite. (On the map, chert pods are symbolized by clusters of small circles.)
-  **pTwak** **K-feldspar-bearing sandstone** Similar to sandstone in the argillite and graywacke unit, but with 2-20% K-feldspar grains and commonly relatively more plagioclase, muscovite and biotite.
-  **pTww** **Metavolcanic rocks** Mostly greenstone and metadiabase, with minor metagabbro, argillite, and sandstone. Amygdaloidal texture and pillow structures are locally preserved. Boudinaged meta-quartz-porphyry dikes cut greenstone breccia on Little Si, adjacent to Mount Si.
-  **pTww** **Metagabbro and minor gneissic amphibolite** Mostly massive to foliated, fine- to medium-grained metagabbro. Rocks in many outcrops sheared on all scales. Sheared rocks range from flaser gabbro with plagioclase cataclasts in schistose chloritic matrix to well-recrystallized greenstone, greenschist, or amphibolite, locally banded.
-  **pTwt** **Metatonalite** Fine- to coarse-grained porphyroclastic metatonalite, locally sheared into light-colored, cataclastic chlorite gneiss. Occurs mostly as small bodies generally associated with metagabbro.
-  **pTwd** **Metadiabase** Mostly metadiabase with relict intergranular to diabasic texture. Most rocks highly thermally metamorphosed and similar to unit pTed. North of Money Creek, the unit includes many hornfelsed Tertiary intrusive dikes and breccia with clasts of metadiabase in tuffaceous matrix.
-  **pTwc** **Chert** Maroon, reddish, and white chert and metachert, in part with shaly interbeds; exhibits undeformed to contorted bedding, and occurs as streamlined pods with scaly margins. Pods range from a few meters to hundreds of meters in length. Rhythmic bedding well preserved. Danner (1970, p. 535, 544-548) described interbedded cherts and marbles in detail.
-  **pTwm** **Marble** Black to bluish-gray marble and, in part, interbedded chert and shale occurring as steeply dipping pods and lenticular beds. Thin beds of green tuffaceous chert crop out in marble at Haystack quarry on Proctor Creek. Further description in Danner (1966, p. 343-358).
-  **pTwu** **Ultramafite** Serpentinized peridotite.
-  **pTws** **Slate, phyllite, and semischist** Rocks similar in composition to the argillite-and-graywacke unit, but with well-developed schistosity that is commonly parallel to bedding. Metamorphic minerals in semischist include sericite, chlorite, calcite, sodic plagioclase, and opaque minerals. Recrystallized near contacts with Tertiary plutons.
-  **pTwsv** **Slate to phyllitic greenstone** Mostly metabasalt and mafic tuff; includes foliated sandstone rich in volcanic detritus.

Table 1.—Fossils and fossil localities in the western and eastern melange belts (See Figure 2 for localities)

| Locality number | Sample number | Description | Age | Reference |
|-----------------|---------------|--|---|---|
| 1 | -- | Trilobite fragments and brachiopods | Ordovician | Duror (in Smith, 1916, p. 580-582) |
| 2 | UW-3488 | Fusulinids in limestone float | Permian | Thompson and others (1950, p. 49) and Danner (1957, p. 270-271) |
| 3 | WA-126 | Crinoid stems in limestone | Indeterminate | Danner (1957, p. 271) |
| 4 | WA-132 | "strongly sheared and deformed" <u>Aucella</u> sp. (now <u>Buchia</u> sp.) | Latest Late Jurassic | Danner (1957, p. 410-411) |
| 5 | WA-133 | "strongly sheared and deformed" <u>Aucella</u> sp. (now <u>Buchia</u> sp.) | Earliest Early Cretaceous | Danner (1957, p. 410-411) |
| 6 | JTW-76-192 | <u>Buchia concentrica</u> | Late Jurassic | D. L. Jones and J. W. Miller (personal commun., 1977) |
| 7 | RWT-145-81 | Radiolarians in chert | Late Jurassic | This report |
| 8 | VF-81-496 | Radiolarians in chert | Early Cretaceous | This report |
| 9 | VF-81-513 | Radiolarians in chert clasts of chert-pebble conglomerate | Late Triassic | This report |
| 10 | VF-81-497 | Radiolarians in chert | Jurassic | This report |
| 11 | RWT-251-78 | Radiolarians in chert | Mesozoic | This report |
| 12 | VF-79-521 | Radiolarians in chert | Late Jurassic | D. L. Jones and B. Murchey (personal commun., 1980) |
| 13 | JTW-80-109 | Radiolarians in chert | Probably Mesozoic | D. L. Jones and B. Murchey (personal commun., 1980) |
| 14 | JTW-80-107 | Radiolarians in chert | Probably Mesozoic | D. L. Jones and B. Murchey (personal commun., 1980) |
| 15 | KO-82-88 | Fragmental ichthyoliths in marble; possible tetrapod vertebrae | Probably Mississippian to Early Permian | A. G. Harris and Nicholas Hotton III (personal commun., 1983) |

of graywacke east of the headwaters of the North Fork of the Snoqualmie River could be either eastern or western melange, but Gualtieri and others (1975, p. 71) reported metavolcanic and ultramafic rocks in a nearby creek, which suggests the eastern melange belt.

Penetratively deformed and metamorphosed oceanic rocks including metamorphosed basalt, chert, limestone, and gabbro exposed near Snoqualmie Pass are also orphaned remnants of the melange of the eastern belt (Frizzell and others, 1984). Danner (1957,

p. 249-255) mapped marble in these exposures and referred the rocks to his Stillaguamish Group (now Trafton sequence) on the basis of lithology. Foster (1957; 1960, p. 111-114) included this melange in his new Denny Formation.

Migmatitic Gneiss

The summit areas northwest of Skykomish are composed mostly of migmatitic hornblende gneiss ranging in composition from amphibolite to trondhjemite. A tonalitic phase of the migmatitic gneiss yields U-Th-Pb ages of about 190 m.y. (Table 2, loc. A). Detailed petrographic studies by Yeats (1958a, p. 83-98; 1964, p. 551-555) indicate that the gneiss underwent regional dynamic metamorphism and granulitization, cataclastic deformation, and, finally, static recrystallization associated with the intrusion of Tertiary batholiths.

Metadiabase, Metatonalite, and Metagabbro

Danner (1957, p. 513-517) and Plummer (1964, p. 54-56) interpreted a large mass of metadiabase southwest of Skykomish as intrusive into the pre-Tertiary sedimentary and volcanic rocks. Erikson (1969, pl. 1) included the metadiabase in his early phase of the Miocene Snoqualmie batholith. The uralitized metadiabase is similar in appearance to that in the melange of the western belt and locally displays steeply dipping crude foliation. It has a sharp contact against sheared pods of metatonalite associated with highly deformed argillite and graywacke, but the relative ages of the metadiabase and metatonalite are unknown. Zircons from the metatonalite in this area (Table 2, loc. B) have $^{207}\text{Pb}/^{206}\text{Pb}$ ages of around 60 to 70 m.y. and are probably partially reset. The U-Pb and Th-Pb ages from these zircons reflect new growth during intrusion of the nearby Tertiary batholiths. Zircons recovered from metatonalite on Mount Garfield have a similar history (Table 2, loc. C).

Zircons from highly sheared metagabbro associated with the metabasalt and marble at Cave Ridge near Snoqualmie Pass yield U-Th-Pb ages of about 165 m.y. (Table 2, loc. D).

Marble

West of Skykomish, south of the Skykomish River, Danner (*in* Thompson and others, 1950, p. 49; Danner, 1966, p. 362-363) found Permian fusulinids and crinoid stems of indeterminate age in a float block of limestone (Table 1, locs. 2 and 3). Although marble lenses are abundant in the melange in this area, the

find was significant, considering that most of the limy rocks in outcrop are well-recrystallized marble, barren of diagnostic fossils. On the basis of fossil evidence and overall lithologic similarity, Danner (1966, p. 363) assigned these rocks to his Trafton sequence exposed in a belt farther north. The rocks contain fusulinids with Tethyan affinities, indicating that the sequence did not become a part of North America until at least the middle Mesozoic (Danner, 1977, p. 497, 500). The Trafton contains fossils ranging from Devonian to Middle Jurassic in age (Danner, 1977, p. 492-493; Whetten and Jones, 1981). Perhaps continued work will push the age of the Trafton back even further and corroborate Smith's (1916) troublesome report of an Ordovician fauna in the melange.

Ultramafite

Pods of ultramafic rocks are both scattered throughout the melange of the eastern belt and concentrated along the sheared and imbricated margin of the migmatitic gneiss. A conspicuous belt of hornfelsed peridotite cuts across the upper Sultan River (Fig. 2). In the Skykomish area, most of the ultramafic rocks are partially or completely altered to tremolite and serpentine minerals, but relict pyrogenic minerals indicate that the pods were originally pyroxenite, peridotite, and dunite (Yeats, 1958a, p. 116-118). On the east side of Weden Creek, a large sliver of dunite appears to be faulted against Tertiary sedimentary and volcanic rocks. Dungan (1974, p. 98-100) argued from detailed petrologic and petrochemical data that correlative ultramafic pods on strike to the north were once a coherent petrologic suite, the later-named Stillaguamish ophiolite of Vance and others (1980). Many of the mafic and ultramafic rocks of the melange of the eastern belt are likely correlative with the dismembered Stillaguamish ophiolite (Vance and others, 1980, p. 362-363, fig. 3). On the basis of regional correlations, Vance and others (1980, p. 378) suggested a Middle and Late Jurassic age of original igneous crystallization and infer a Middle Jurassic to middle Cretaceous age for the associated volcanic rocks.

Melange of the Western Belt

The melange of the western belt is predominantly argillite and graywacke (subquartzose sandstone), with generally lesser amounts of mafic volcanic rocks, conglomerate, chert, and marble compared to the melange of the eastern belt. Ultramafite, mostly serpentinite or metaperidotite, is present, but very scarce. Outcrop-to mountain-sized phacoids of metagabbro, metadiabase, and metatonalite are ubiquitous. In the general area of the lower Sultan Basin (Fig. 2), the disrupted rocks of

Table 2.—U-Th-Pb and K-Ar¹ ages of rocks in the eastern and western melange belts, central Washington (See Figure 2 for localities)

| Location | Sample number | Method | Material | Melange belt | Unit Symbol (Fig. 2) | Age ¹ (m.y.) | Reference |
|----------|---------------|---------|------------|---|------------------------|--|--------------------------|
| A | JTW-79-214 | U-Th-Pb | Zircon | Eastern | pTeg | (-100,+150) 186.0; 187.0; 199.0; 195.0 (-200,+270) 191.0; 193.0; 211.0; 196.0 | Whetten and others, 1980 |
| B | RWT-513-80 | U-Th-Pb | Zircon | Eastern | pTet, (hornfelsed) | (-150,+200) 39.4; 39.8; 62.6; 37.9 (-250,+325) 47.0; 47.5; 72.0; 46.6 | This report (Appendix 1) |
| C | JTW-79-215 | U-Th-Pb | Zircon | Eastern | pTet, (hornfelsed) | (-200,+250) 28.1; 28.7; 77.2; 29.1 (-250,+325) 34.4; 34.9; 73.3; 33.0 | This report (Appendix 1) |
| D | JTW-79-216 | U-Th-Pb | Zircon | Eastern | pTeg | (-150,+200) 164.7; 165.9; 183.4; 173.4 (-250,+325) 158.2; 159.3; 175.9; 152.3 | This report (Appendix 1) |
| E | JTW-77-215 | U-Th-Pb | Zircon | Western | pTwg | (-100,+150) 152.0; 152.0; 159.0; 146.0 (-200,+270) 148.0; 149.0; 169.0; 144.0 | Whetten and others, 1980 |
| F | JTW-76-191 | U-Th-Pb | Zircon | Western | pTwt | (-100,+150) 164.0; 164.0; 168.0; 165.0 (-270,+325) 172.0; 173.0; 184.0; 167.0 | Whetten and others, 1980 |
| G | RWT-278-80 | U-Th-Pb | Zircon | Western | pTwt | (-150,+200) 148.1; 149.0; 163.3; 146.7 (-250,+325) 149.8; 150.5; 161.1; 149.8 | This report (Appendix 1) |
| H | RWT-316-80 | U-Th-Pb | Zircon | Western | pTwg | (-150,+200) 148.2; 149.1; 162.3; 143.0 (-250,+325) 148.1; 149.5; 172.0; 145.9 | This report (Appendix 1) |
| I | RWT-70-81 | K-Ar | Hornblende | Western | pTwg | 96.3±3.9 | |
| J | JTW-76-199 | K-Ar | Hornblende | Western | pTwg | 118.0±8.3 | |
| K | RWT-372-80 | K-Ar | Sericite | Western | pTws, (metamorphic) | 47.8±0.7 | |
| L | DB-81-568 | K-Ar | Hornblende | Granodiorite intrusive into western melange | | 46.6±2.1 | |

¹K-Ar ages calculated on the basis of 1976 IUGS decay and abundance constants. Errors in K-Ar ages based on variation taken from an empirical curve relating the coefficient of variation in the age to percent radiogenic argon (Tabor and others, 1985). U-Th-Pb isotope ages reported in the following order: ²⁰⁶Pb/²³⁸U; ²⁰⁷Pb/²³⁵U; ²⁰⁷Pb/²⁰⁶Pb; ²⁰⁸Pb/²³²Th. Two data sets are reported.

the western melange grade into one mappable unit of slate, phyllite, and semischist, with minor greenschist and chert. This greater development of metamorphism and schistosity is enigmatically limited to a zone between less-metamorphosed melange of the western belt and the at least less-schistose melange of the eastern belt. In the upper Sultan Basin area, where Tertiary contact metamorphism obscures the penetrative shearing in the phyllitic rocks of the western melange, we distinguished the eastern melange from the phyllitic unit of the western melange on the basis of greater amounts of greenstone, chert, and ultramafite in the eastern melange. Our contact lies west of a wide fault zone mapped by Heath (1971, pl. 1); the disruption in Heath's fault zone seems to us to be no more severe than disruption elsewhere in the melange belts, and we prefer to place the fault contact between the two melange belts on the basis of lithology.

Although the strike of bedding and foliation is generally within 45° of north, some discernible variations do exist (Tabor and others, 1982). The strike of both bedding and foliation south of the Skykomish River averages NNE; dips for both range between 20° and 80° , mostly dipping to the east. Attitudes in argillite and graywacke north of the Skykomish have NNW strikes, and most dip east. Foliation in the schistose unit generally parallels bedding and has a mean NNW strike with 70° dips to the east; these attitudes gradually change from north-south near the Sultan Basin to WNW south of the Pilchuck River.

Argillite and Graywacke

Fuller (1925, p. 29-39, 46-55) first recognized the extreme deformation in rocks we consider part of the western melange belt. Near their southern outcrop limit, he described "angular blocks of quartzite (graywacke) imbedded in a contorted mass of argillite that exhibits marked lines of flowage," structures similar to those illustrated in Figure 3. Bethel (1951, p. 22-55) mapped feldspathic graywackes and argillite in the southern part of the Snoqualmie River area and referred them to his Calligan unit (the later-named Calligan Formation of Grant, 1969), ignoring the name applied earlier by Fuller. Although Bethel (1951, p. 43-56) incorrectly correlated the graywacke with the Swauk Formation, Kremer (1959, p. 40-62) expanding upon the work of Danner farther north, correctly assigned the graywacke (and associated argillite and chert) a Late Jurassic and Early Cretaceous age.

Danner (1957, p. 330-363) considered the oceanic sedimentary rocks predominating in this belt to be



Figure 3.—Streamlined phacoids of graywacke enclosed in argillaceous matrix. West side of Proctor Creek, near locality H of Figure 2.

correlative lithologically and in age with the relatively undeformed Upper Jurassic-Lower Cretaceous Nooksack Formation described by Misch (1966, p. 118; Sondergaard, 1979) and exposed about 120 km to the north. Danner referred the rocks to his Sultan unit and Olo Mountain unit (the later-named Olo Mountain Formation of Miller, 1979). Less tectonized rocks of similar age and lithology on the San Juan Islands (Fig. 1) make up the Decatur terrane of Cowan and others (1977, p. 324) and the earlier described Constitution Formation of McKee (1972) and the Lummi Formation of Russell (1975; also, Vance, 1975, p. 13; 1977, p. 182, 186).

K-Feldspar-bearing Sandstone

Some outcrops of sandstone in the western melange belt contain 2 to 20 percent K-feldspar grains and, commonly, more plagioclase, muscovite, and biotite than the balance of the western belt. Rocks of the westernmost part of the western melange belt appear to be richer in K-feldspar than rocks farther east, but the sampling is sparse. The eastern melange is notably poor in K-feldspar. Whether these differences in K-feldspar content are due to differences in provenance of the detrital materials or loss of K-feldspar during metamorphism, especially in the eastern melange, is unknown.

Slate, Phyllite, and Semischist

Sandstone and argillite in the western melange grade eastward into slaty argillite, slate, phyllite, and semischist in the upper Sultan drainage area; the change from nonfoliate to foliate rocks takes place over a few hundred meters. Greenstone and tuffaceous rocks grade eastward into phyllitic greenstone or fine-grained greenschist.

This more metamorphosed unit is continuous with rocks that Danner (1957, p. 423-455) included in his phyllitic Olo Mountain unit, north of the area considered here (Fig. 1). Danner felt that his Olo Mountain unit was also lithologically similar to the Nooksack Formation, except that the Olo Mountain contained more chert. Similarly, our phyllitic unit contains more chert than the less foliated rocks to the west. However, we include in our phyllitic unit Upper Jurassic and Lower Cretaceous fossiliferous rocks on the north side of the Sultan River, which Danner included in his Sultan unit (Table 1, locs. 4-6).

Non-schistose and Schistose Metavolcanic Rocks

Most metavolcanic rocks in the western melange belt crop out in the area south of the Sultan River. The protoliths were mostly mafic, probably basaltic, and some were deposited in submarine environments, as shown by pillow structures. Boudinaged and penetratively sheared quartz-porphry dikes intruding the metavolcanic rocks in the Mount Si area (Fig. 2) are the only silicic metavolcanic rocks that we have found.

Chert

Chert occurs as isolated blocks and thick lenses throughout the melange of the western belt. Individual blocks range from a few meters to more than a kilometer long. Chert blocks are commonly highly broken and boudinaged, but may be relatively undeformed or crinkle folded. Several recently sampled radiolarian chert bodies from the western melange—with the exception of a block of chert pebble conglomerate—range in age from Late Jurassic (Kimmeridgian/Tithonian) to Early Cretaceous (Berriasian/early Valanginian). A Late Jurassic (Kimmeridgian/Tithonian) radiolarian assemblage (Table 1, loc. 7) contained *Archaeodictyomitra* (?) *rigida* Pessagno, *Hsuum* sp., *Mirifusus* (?) *mediodilatata* (Rust), *Pantanelium* (?) *riedeli* Pessagno, and *Parvicingula* sp. An early Cretaceous (Berriasian/early Valanginian) fauna from the western melange belt (Table 1, loc. 8) contained *Mirifusus baileyi* Pessagno, *M. mediodilatata* (Rust), and *Tripocyclus blakei* Pessagno.

(For a review of Jurassic and Early Cretaceous radiolarian biostratigraphy for the western United States, refer to Pessagno, 1977a,b; Pessagno and Blome, 1980; Pessagno and Whalen, 1982; and Pessagno and others, 1984). Other chert bodies generally contain a Jurassic radiolarian fauna (Table 1, locs. 10-14).

Chert clasts from the chert-pebble conglomerate (Fig. 4; Table 1, loc. 9) contain Late Triassic (Carnian/Norian) radiolarian taxa, including *Canesium lentum* Blome, ?*Canoptum* sp., *Castrum perornatum* Blome, *Triassocampe* sp., and *Xenorum* sp. The same radiolarian faunal assemblage has been reported from Baja California (Pessagno and others, 1979), northern California (Blome and Irwin, 1983), and eastern Oregon and British Columbia (Pessagno and Blome, 1980; Blome, 1983, 1984).

Marble

A quarry in the Proctor Creek area exposes thin layers of altered mafic volcanic ash interbedded with marble. Samples from this quarry yield ichthyoliths (fish teeth) and tetrapod (early amphibian) vertebrae (Table 1, loc. 15), indicating a probable Mississippian to Early Permian age. Limestone bodies in rocks typical of the western melange north of the study area yield Permian fusulinids (Danner, 1966, p. 72). These older components of the melange of the western belt make the age range of its protolith comparable to the ages reported from the melange of the eastern belt.

Metagabbro, Metadiabase, and Metatonalite

Metagabbro in the western melange belt ranges from slightly uralitized ophitic gabbro through cataclasite to well-recrystallized gneissic amphibolite.

The mode of formation and emplacement of the gabbroic masses is important to the interpretation of the origin of the melange unit. Danner (1957, p. 535-541) considered the metagabbro bodies (his Woods Creek intrusive bodies) to be intrusive into the sedimentary rocks. Sparse, but relatively widespread, strongly uralitized metadiabase dikes are clearly intrusive into the sedimentary rocks. Similar rocks, presumably dikes, occur in the overlying Tertiary volcanic rocks as well. However, most contacts between the larger metagabbro bodies (Fig. 5) and other metagabbro bodies and adjoining rocks are strongly sheared (Fig. 6); hornfelsic textures are not found in the sedimentary rocks adjacent to the meta-intrusives. Small knockers of metagabbro are clearly tectonically emplaced within the sedimentary rocks. The most compelling evidence that the metagabbro and metatonalite masses are exotic in the melange, and not intrusive, are the 150- to 170-m.y. U-Th-Pb ages



A



B

Figure 4.—A. Phacoid of chert-pebble conglomerate (locality 9 of Fig. 2 and Table 1), surrounded by scaly argillaceous matrix of the melange of the western belt; the chert pebbles contain a Late Triassic radiolarian fauna similar to those found at widely separated localities from Baja California to British Columbia. B. Detail of phacoid/matrix boundary.

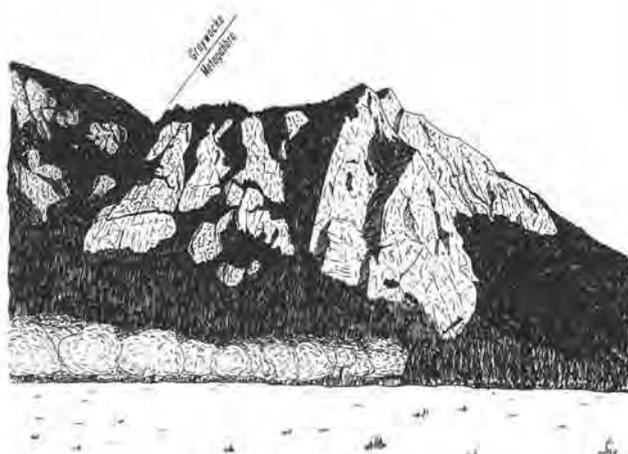


Figure 5.—Large knocker of metagabbro that forms Mount Si, near North Bend.

(Table 2, locs. E-H), which are older than the enclosing Late Jurassic and Early Cretaceous sedimentary rocks (152-131 m.y.; time scale of Palmer, 1983). We interpret the conventional K-Ar ages on hornblende from the uralitic metagabbro (96 and 118 m.y.; Table 2, locs. I-J) as most likely reflecting uplift or as-yet-unrecognized post-Late Jurassic thermal events.

Vance and others (1980) studied the trace-element composition of some metagabbro, metabasalt, and ultramafite from disrupted ophiolites in western Washington. Their data, although pooled with data on rocks correlative with the eastern melange, support an oceanic origin for the mafic and ultramafic rocks of the melange of the western belt.

DISCUSSION

The penetrative deformation, the range in the ages of the components, and the mixing of rocks of contrasting origin indicate that the rocks herein described are melanges. Of the questions arising from such an assertion, perhaps the most important are these: (1) How did the melanges form?, and (2) When were the melanges emplaced?

Formation of the Melanges

Chaotic mixtures of materials such as those in the melanges of the eastern and western belts have been attributed to olistostromal or tectonic processes involving large-scale submarine sliding or mixing in subduction, triple-junction, or transform-fault settings.

There is little doubt that native and exotic blocks of various sizes are present in a generally pervasively deformed matrix in the melanges. Evidence for small-scale olistostromal or surficial sedimentary mixing is



Figure 6.—Sheared contact between metatonalite and argillaceous matrix. Locality H of Figure 2 and Table 2.

evident in a few outcrops, but large-scale olistostromal mixing has not been documented. Phacoids of various sizes that range to as large as several kilometers in length appear to have deformed in a brittle manner and are surrounded by an argillaceous matrix that deformed in a more plastic manner. Thus, if the rocks had originally been deposited as olistostromes, they were subsequently deformed under some confining pressure. Furthermore, the phyllitic rocks in the Sultan Basin area are proof of a tectonic process at least late in the formation of the melanges.

Northwest of Skykomish, the formation of a chaotic zone by shearing at the base of a major overthrust fault has been suggested by Yeats (1964). On the basis of a regional correlation with similar sheared metamorphic complexes that apparently overlie unmetamorphosed rocks as described by Misch (1960, 1963, 1966, p. 106), Yeats (1964, p. 559) considered the migmatitic gneiss in the melange of the eastern belt to be klippen of basement rocks thrust over younger rocks. The evidence that the migmatitic gneiss is in tectonic contact with the sedimentary and volcanic rocks is, indeed, convincing. However, argillite is

imbricated with the gneiss in many places (Yeats, 1964, p. 556), and even lies atop the gneiss on Baring Mountain. Furthermore, some contacts presumed to be horizontal are in fact steep (such as those on the south side of Gunn Peak), and some of the high-angle faults mapped by Yeats (1958a, pl. II; 1964, pl. 1) may be the same age as the faults emplacing the gneiss against the sedimentary and volcanic rocks. Our alternative interpretation to the overthrust hypothesis is that the gneiss masses are exotic blocks in the eastern melange, an interpretation alluded to by Yeats (1964, p. 558) when he referred to the klippen as megabreccia.

Uranium-thorium-lead ages from the tonalite phase of the migmatitic gneiss, the plutonic rocks at Woods Creek, and the Mount Si massif led Whetten and others (1980, p. 365-366) to suggest that these rocks were in-folded or in-faulted klippen of their regional Haystack thrust plate that was emplaced between latest middle Cretaceous and earliest Late Cretaceous time. However, as an alternative hypothesis, they suggested a large-scale melange in which exotic blocks of Mesozoic ophiolite were imbricated over a wide area in western Washington, including our eastern and western melanges. The appearance of imbrication at the base of a large thrust sheet, such as the proposed Haystack thrust, probably would differ little from the appearance of imbrication and tectonic mixing at the upper surface of an oceanic plate during subduction (for example, Hamilton, 1969, p. 2415-2416). Perhaps we are looking at different parts of the same elephant.

That melanges are formed in subduction zones has become nearly an article of faith for many workers. Fox (1983), however, pointed out the poor correlation between known past subduction zones and melanges of the same age. He proposed that melanges form at Humboldt-type triple junctions as they migrate along the continental margin (Fox, 1976) and asserted that reasonable chronological and spatial ties exist between melanges and the transit of these features (Fox, 1983, fig. 11). He envisioned melanges formed as huge tectonic gravity slides off upwarded oceanic floor and emplaced on the continental margin as huge nappe-like thrusts. We find Fox's model appealing and are intrigued by the idea that a Humboldt-type triple junction may have existed in the vicinity of Seattle about 100 m.y. ago (Fox, 1983, fig. 11).

On the other hand, other workers (for example, Asahiko Taira, personal commun., 1983) suggest that some melange zones, such as the Kurosegawa belt in Japan, formed in strike-slip regimes; M. C. Blake, Jr. (personal commun., 1983) has indicated that this may be the process that formed the central-belt melanges of the Franciscan assemblage in northern California.

Timing of Melange Formation and Emplacement

The age of emplacement of the melange belts is uncertain, but some constraints do exist. Although we have inferred that the melanges of the eastern and western belts are in fault contact, they both could have been accreted to North America at about the same time. Formation and emplacement of the melange of the western belt must have taken place after the Early Cretaceous, because the melange contains fossils of that age. Combining petrologic, chemical, and isotopic evidence, Dungan (1974, p. 216-217), Vance and Dungan (1977), and Johnson and others (1977) considered the northern parts of the Stillaguamish ophiolite to have been regionally metamorphosed to at least the middle amphibolite facies prior to faulting that emplaced it adjacent to the unmetamorphosed rocks. This regional metamorphism occurred in the Middle or Late Jurassic (Vance and others, 1980, p. 378-384), and accretion of the ophiolite and its sedimentary and volcanic cover took place in the early Late Cretaceous (Vance and others, 1980, p. 362), an age in agreement with that of the proposed emplacement of the Haystack thrust of Whetten and others (1980). These early events could have been amalgamation processes west of the continental margin, with the actual docking of the rocks to North America occurring later.

Monger and others (1982, fig. 2) indicated that the approximate time of accretion of these melange belts was the latest Late Cretaceous, after the Late Cretaceous metamorphic event—documented as about 60 to 90 m. y. ago by Mattinson (1972, p. 3778) in the North Cascades—that they ascribe to “tectonic overlap and (or) compressional thickening of crustal rocks” during collision between North America and Wrangellia and other terranes that form a composite terrane unit inboard of their Pacific Rim terrane that is in part correlated with our melange belts (Monger and others, 1982, p. 70, fig. 2). If this model is correct, it could help explain the absence of Late Cretaceous high-grade thermal metamorphism in most of the melange. It is possible, though, that the melange belts were accreted in pre-Late Cretaceous time, but juxtaposed with the higher-grade metamorphic rocks by strike-slip and/or vertical movement on the Straight Creek fault. The Straight Creek fault, first noted by Vance (1957, p. 302), is one of the major structures of the Cascade Range. Interpretations concerning the existence, importance, and timing of movement vary, but it is clear that considerable right-lateral post-Late Cretaceous strike-slip movement has occurred, indicated by correlation of offset alumina-rich schist and associated Late Cretaceous intrusions (Misch, 1977, p. 36; Friz-

zell, 1979, p. 92; Vance and Miller, 1981; Kleinspehn, 1982).

Although lower Eocene rocks are clearly cut and deformed by the Straight Creek fault and northwest-trending dikes of the Teanaway Formation south of Mount Stuart indicate that the fault was still active in the middle Eocene, large amounts of movement probably have not occurred after late Eocene time (Tabor and Frizzell, 1979) and definitely not since Oligocene time (Vance and Miller, 1981).

The timing and origin of the metamorphism that produced the phyllitic part of the western belt are important footnotes to this story. If the phyllitic rocks represent the highest metamorphic grade attained during imbrication of the Mesozoic rocks during amalgamation or plate collision, a K-Ar age of about 48 m.y. (Table 2, loc. K) on newly formed sericite in a well-recrystallized phyllite from the western belt is a minimum age for one or the other of these events. The presence of the eastern phyllitic part of the western belt between unmetamorphosed eastern belt and the relatively unmetamorphosed part of the western belt may indicate that the two belts also were juxtaposed by high-angle faulting after the accretion that presumably formed them.

On Fuller Mountain, near North Bend, the melange of the western belt is intruded by granodiorite which yielded a K-Ar age of 47 m.y. (Table 2, loc. L). This early Eocene episode of intrusion may also be represented by the 44-m.y.-old Granite Falls and 49-m.y.-old Mount Pilchuck stocks, both north of the report area (Yeats and Engels, 1971). We know that the stocks were intruded after formation of the melange, but we do not know whether the stocks were intruded before or after accretion of the melange. These lower Eocene plutons are generally assigned to the root zone of the northwest-trending Challis arc, most clearly defined and continuous about 70 km to the northeast and generally east of the younger Cascade arc (Yeats and Engels, 1971; Vance, 1982; Frizzell and Vance, 1983). The plutons seem out of place, but preliminary paleomagnetic work on the Granite Falls stock indicates no significant rotation or northwest translation since its cooling (Beck and others, 1982, p. 516).

We know that the melange of the eastern belt was emplaced at least by the early Eocene because it is overlain by the lower Eocene Silver Pass Volcanic Member of the Swauk Formation. The rocks in the melange of the western belt are overlain unconformably by the gently folded volcanic rocks of Mount Persis that are probably late Eocene in age (Tabor and others, 1982). The Mount Persis volcanic rocks overlie

Table 3.—Summary of lithologies, fossil ages, isotopic determination, and contact relationships of melange belts in the western Cascades of Washington

| Characteristic | Eastern melange belt | Western melange belt |
|-------------------------|--|--|
| Lithology | Predominantly a mixture of mafic volcanic rocks, chert, and lesser amounts of argillite and graywacke; conspicuous marble and ultramafite; includes blocks of migmatitic gneiss, metadiabase, and metatonalite | Predominantly argillite and graywacke, in part K-feldspar rich, with phacoids of metavolcanic rocks, chert, marble, metagabbro, metadiabase, and metatonalite; very rare ultramafite; includes a phyllitic unit consisting of slate, phyllite, semischist, and greenschist |
| Fossil ages | Permian fusulinids and fragments of Ordovician brachiopods and trilobites in marble | Late Jurassic to Early Cretaceous <i>Buchia</i> sp. in metasandstone in the phyllitic unit; Late Jurassic to Early Cretaceous radiolarians in chert pods; Triassic radiolarians in chert-pebble conglomerate; fragments of ichthyoliths and tetrapods in pods of marble indicate probable Mississippian to early Permian Age |
| Isotopic determinations | 190-m.y. U-Th-Pb age on zircon from tonalitic gneiss 165-m.y. U-Th-Pb age on zircon from metagabbro | 150- to 170-m.y. U-Th-Pb ages on zircons from metagabbro and metatonalite; 47-m.y. K-Ar age on granodiorite intruding melange |
| Contacts | Probable fault contact with melange of the western belt; unconformably overlain by lower Eocene sedimentary and volcanic rocks; intruded by 33- to 18-m.y.-old granitic rocks | In fault contact with lower and middle Eocene sedimentary and volcanic rocks; unconformably overlain by upper Eocene volcanic rocks; intruded by 47-m.y. old granodiorite as well as younger (33- to 18-m.y.-old) granitic rocks |

only rocks of the western melange belt, and their lack of deformation contrasts with the strong deformation in contemporaneous rocks of the adjoining Puget Group to the south (Vine, 1969; Tabor and others, 1982; Turner and others, 1983; Frizzell and others, 1984). This suggests that the entire block—melange of the western belt overlain by the volcanic rocks of Mount Persis—is in fault contact with the Puget Group.

Thus, the lithologically distinguishable melanges (Table 3) formed during or after the Early Cretaceous and before the early Eocene. They were concurrently or subsequently accreted to North America, but before the early Eocene. The melanges were translated northward, relative to North America, by movement on the Straight Creek and equivalent faults, and they have probably been near their current relative position since late Eocene time.

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REFERENCES CITED

- Bailey, E. H., 1941, Mineralogy, petrology, and geology of Santa Catalina Island, California: Stanford University Doctor of Philosophy thesis, 194 p., 1 pl.
 Baum, L. F., 1968, Geology and mineral deposits, Vesper Peak stock area, Snohomish County, Washington: University of Washington Master of Science thesis, 75 p., 1 pl.

- Beck, M. E., Jr.; Burmester, R. F.; Schooner, Ruth, 1982, Tertiary paleomagnetism of the north Cascade Range, Washington: *Geophysical Research Letters*, v. 9, no. 5, p. 515-518.
- Berkland, J. O.; Raymond, L. A.; Kramer, J. C.; Moores, E. M.; O'Day, Michael, 1972, What is Franciscan?: *American Association of Petroleum Geologists Bulletin*, v. 56, no. 12, p. 2295-2302.
- Bethel, H. L., 1951, Geology of the southeastern part of the Sultan quadrangle, King County, Washington: University of Washington Doctor of Philosophy thesis, 244 p., 1 pl.
- Blome, C. D., 1983, Upper Triassic Capnuchoisphaeridae and Capnodocinae (Radiolaria) from east-central Oregon: *Micropaleontology*, v. 29, no. 1, p. 11-49.
- Blome, C. D., 1984, Upper Triassic Radiolaria and radiolarian zonation from western North America: *Bulletins of American Paleontology*, v. 85, no. 318, 88 p.
- Blome, C. D.; Irwin, W. P., 1983, Tectonic significance of Late Paleozoic to Jurassic radiolarians from the North Fork terrane, Klamath Mountains, California. In Stevens, C. H., editor, 1983, *Pre-Jurassic rocks in western North American suspect terranes*: Society of Economic Paleontologists and Mineralogists, Pacific Section, p. 77-89.
- Coney, P. J.; Jones, D. L.; Monger, J. W. H., 1980, Cordilleran suspect terranes: *Nature*, v. 288, no. 5789, p. 329-333.
- Cowan, D. S.; Whetten, J. T.; Brown, E. H., 1977, Geology of the southern San Juan Islands. In Brown, E. H.; Ellis, R. C., editors, 1977, *Geological excursions in the Pacific Northwest*: Geological Society of America meeting, 1977: Western Washington University, p. 309-338.
- Danner, W. R., 1957, A stratigraphic reconnaissance in the northwestern Cascade Mountains and San Juan Islands of Washington State: University of Washington Doctor of Philosophy thesis, 561 p., 7 pl.
- Danner, W. R., 1966, Limestone resources of western Washington: Washington Division of Mines and Geology Bulletin 52, 474 p.
- Danner, W. R., 1970, Cherts and jaspers of the western cordilleran eugeosyncline of southwestern British Columbia and northwestern Washington: *West Commemoration Volume 1970*, p. 533-553.
- Danner, W. R., 1977, Paleozoic rocks of northwest Washington and adjacent parts of British Columbia. In Stewart, J. H.; Stevens, C. H.; Fritsche, A. E., editors, 1977, *Paleozoic paleogeography of the western United States*; Pacific Coast Paleogeography Symposium 1: Society of Economic Paleontologists and Mineralogists Pacific Section, p. 481-502.
- Davis, G. A.; Monger, J. W. H.; Burchfiel, B. C., 1978, Mesozoic construction of the cordilleran "collage", central British Columbia to central California. In Howell, D. G.; McDougall, K. A., editors, 1978, *Mesozoic paleogeography of the western United States*; Pacific Coast Paleogeography Symposium 2: Society of Economic Paleontologists and Mineralogists Pacific Section, p. 1-32.
- Dawson, G. M., 1895, Report on the area of the Kamloops map sheet, British Columbia: Geological Survey of Canada Annual Report, new series 7, p. 39B-49B.
- Dungan, M. A., 1974, The origin, emplacement, and metamorphism of the Sultan mafic-ultramafic complex, north Cascades, Snohomish County, Washington: University of Washington Doctor of Philosophy thesis, 227 p., 3 pl.
- Erikson, E. H., Jr., 1969, Petrology of the composite Snoqualmie batholith, central Cascade Mountains, Washington: *Geological Society of America Bulletin*, v. 80, no. 11, p. 2213-2236.
- Foster, R. J., 1957, Tertiary geology of a portion of the central Cascade Mountains, Washington [abstract]: *Dissertation Abstracts*, v. 17, no. 9, p. 1982.
- Foster, R. J., 1960, Tertiary geology of a portion of the central Cascade Mountains, Washington: *Geological Society of America Bulletin*, v. 71, no. 2, p. 99-125.
- Fox, K. F., Jr., 1976, Melanges in the Franciscan Complex, a product of triple-junction tectonics: *Geology*, v. 4, no. 12, p. 737-740.
- Fox, K. F., Jr., 1983, Melanges and their bearing on late Mesozoic and Tertiary subduction and interplate translation at the west edge of the North American plate: U.S. Geological Survey Professional Paper 1198, 40 p.
- Frizzell, V. A., Jr., 1979, Petrology and stratigraphy of Paleogene non-marine sandstones, Cascade Range, Washington: U.S. Geological Survey Open-File Report 79-1149, 151 p., 3 pl.
- Frizzell, V. A., Jr.; Tabor, R. W.; Booth, D. B.; Ort, K. M.; Waite, R. B., Jr., 1984, Preliminary geologic map of the Snoqualmie Pass 1:100,000 quadrangle, Washington: U.S. Geological Survey Open-File Report 84-693, 43 p., 1 pl.
- Frizzell, V. A., Jr.; Tabor, R. W.; Zartman, R. E.; Jones, D. L., 1982, Mesozoic melanges in the western Cascades of Washington [abstract]: *Geological Society of America Abstracts with Programs*, v. 14, no. 4, p. 164.
- Frizzell, V. A., Jr.; Vance, J. A., 1983, Cenozoic volcanism of the Cascades of Washington [abstract]: *EOS (American Geophysical Union Transactions)*, v. 64, no. 45, p. 886.
- Fuller, R. E., 1925, The geology of the northeastern part of Cedar Lake quadrangle, with special reference to the de-roofed Snoqualmie batholith: University of Washington Master of Science thesis, 96 p., 4 pl.
- Galster, R. W., 1956, Geology of the Miller-Foss River area, King County, Washington: University of Washington Master of Science thesis, 96 p., 2 pl.
- Grant, A. R., 1969, Chemical and physical controls for base metal deposition in the Cascade Range of Washington: Washington Division of Mines and Geology Bulletin 58, 107 p.
- Gualtieri, J. L.; Thurber, H. K.; Miller, M. S.; McMahan, A. B.; Federspiel, F. F., 1975, Mineral resources of additions to the Alpine Lakes study area, Chelan, King, and Kittitas Counties, Washington: U.S. Geological Survey Open-File Report 75-3, 161 p., 2 pl.
- Hamilton, Warren, 1969, Mesozoic California and the underflow of Pacific mantle: *Geological Society of America Bulletin*, v. 80, no. 12, p. 2409-2429.
- Heath, M. T., 1971, Bedrock geology of the Monte Cristo area, northern Cascades, Washington: University of Washington Doctor of Philosophy thesis, 164 p., 1 pl.
- Hsu, K. J., 1968, Principles of melanges and their bearing on the Franciscan-Knoxville paradox: *Geological Society of America Bulletin*, v. 79, no. 8, p. 1063-1074.
- Johnson, Michael; Dungan, M. A.; Vance, J. A., 1977, Stable isotope compositions of olivine and dolomite in peridotites formed by deserpentinization, Darrington area, north Cascades, Washington: *Geochimica et Cosmochimica Acta*, v. 41, no. 3, p. 431-435.
- Jones, D. L.; Blake, M. C., Jr.; Bailey, E. H.; McLaughlin, R. J., 1978, Distribution and character of upper Mesozoic subduction complexes along the west coast of North America: *Tectonophysics*, v. 47, no. 3-4, p. 207-222.
- Kleinspehn, K. L., 1982, Cretaceous sedimentation and tectonics, Tyaughton-Methow Basin, southwestern British Columbia: Princeton University Doctor of Philosophy thesis, 198 p.
- Kremer, D. E., 1959, The geology of the Preston-Mt. Si Area: University of Washington Master of Science thesis, 103 p., 1 pl.
- Mattinson, J. M., 1972, Ages of zircons from the northern Cascade Mountains, Washington: *Geological Society of America Bulletin*, v. 83, no. 12, p. 3769-3783.

- McKee, Bates, 1972, *Cascadia—The geologic evolution of the Pacific Northwest*: McGraw-Hill Book Company, 368 p.
- Miller, G. M., 1979, Western extent of the Shuksan and Church Mountain thrust plates in Whatcom, Skagit, and Snohomish Counties, Washington: *Northwest Science*, v. 53, no. 4, p. 229-240.
- Misch, Peter, 1960, Large overthrusts in the northwestern Cascades near the 49th parallel (Whatcom and Skagit Counties, Washington, and lower Tomyhoi Creek area, British Columbia) [abstract]: *Geological Society of America Bulletin*, v. 71, no. 12, part 2, p. 2069.
- Misch, Peter, 1963, Crystalline basement complex in northern Cascades of Washington [abstract]: *Geological Society of America Special Paper* 76, p. 213-214.
- Misch, Peter, 1966, Tectonic evolution of the northern Cascades of Washington State—A west-cordilleran case history. In Gunning, H. C., editor, 1966, *Symposium on tectonic history and mineral deposits of the western Cordillera in British Columbia and neighboring parts of the United States*, Vancouver, 1964: Canadian Institute of Mining and Metallurgy Special Volume 8, p. 101-148.
- Misch, Peter, 1977, Dextral displacements at some major strike faults in the north Cascades [abstracts]: *Geological Association of Canada Program with Abstracts* 1977, v. 2, p. 37.
- Monger, J. W. H.; Price, R. A.; Tempelman-Kluit, D. J., 1982, Tectonic accretion and the origin of the two major metamorphic and plutonic belts in the Canadian Cordillera: *Geology*, v. 10, no. 2, p. 70-75.
- Palmer, A. R., compiler, 1983, The decade of North American geology—1983 geologic time scale: *Geology*, v. 11, no. 9, p. 503-504.
- Pessagno, E. A., Jr., 1977a, Upper Jurassic Radiolaria and radiolarian biostratigraphy of the California Coast Ranges: *Micropaleontology*, v. 23, no. 1, p. 56-113.
- Pessagno, E. A., Jr., 1977b, Lower Cretaceous radiolarian biostratigraphy of the Great Valley Sequence and Franciscan complex, California Coast Ranges: *Cushman Foundation for Foraminiferal Research Special Publication* 15, 86 p.
- Pessagno, E. A., Jr.; Blome, C. D.; 1980, Upper Triassic and Jurassic Pantanellinae from California, Oregon and British Columbia: *Micropaleontology*, v. 26, no. 3, p. 225-273.
- Pessagno, E. A., Jr.; Blome, C. D.; Longoria, J. F., 1984, A revised radiolarian zonation from Upper Jurassic of western North America: *Bulletins of American Paleontology*, v. 87, no. 320, 51 p.
- Pessagno, E. A., Jr.; Finch, W.; Abbott, P. L., 1979, Upper Triassic Radiolaria from the San Hipolito Formation, Baja California: *Micropaleontology*, v. 25, no. 2, p. 160-197, 9 pl.
- Pessagno, E. A., Jr.; Whalen, P. A., 1982, Lower and Middle Jurassic Radiolaria (multicyrtid Nassellariina) from California, east-central Oregon and the Queen Charlotte Islands, B. C.: *Micropaleontology*, v. 28, no. 2, p. 111-169.
- Plummer, C. C., 1964, The geology of the Mount Index area of Washington State: University of Washington Master of Science thesis, 62 p., 1 pl.
- Russell, R. H., editor, 1975, *Geology and water resources of the San Juan Islands, San Juan County, Washington*: Washington Department of Ecology Water-Supply Bulletin 46, 171 p., 3 pl.
- Silberling, N. J.; Jones, D. L.; Blake, M. C., Jr.; Howell, D. G.; and others, 1984, Lithotectonic terrane map of the western conterminous United States. In Silberling, N. J.; Jones, D. L., editors, 1984, *Lithotectonic terrane maps of the northern American Cordillera*: U.S. Geological Survey Open-File Report 84-523, p. CI-C43, 1 pl.
- Silver, E. A.; Buetner, E. C., conveners, 1980, Penrose Conference report; *Melanges: Geology*, v. 8, no. 1, p. 32-34.
- Smith, W. S., 1916, Stratigraphy of the Skykomish basin, Washington: *Journal of Geology*, v. 24, no. 6, p. 559-582.
- Sondergaard, J. N., 1979, Stratigraphy and petrology of the Nooksack Group in the Glacier Creek-Skyline Divide area, north Cascades, Washington: Western Washington University Master of Science thesis, 103 p., 2 pl.
- Tabor, R. W.; Frizzell, V. A., Jr., 1979, Tertiary movement along the southern segment of the Straight Creek Fault and its relation to the Olympic-Wallowa lineament in the central Cascades, Washington [abstract]: *Geological Society of America Abstracts with Programs*, v. 11, no. 3, p. 131.
- Tabor, R. W.; Frizzell, V. A., Jr.; Booth, D. B.; Whetten, J. T.; Waitt, R. B., Jr.; Zartman, R. E., 1982, Preliminary geologic map of the Skykomish River 1:100,000 quadrangle, Washington: U.S. Geological Survey Open-File Report 82-747, 31 p., 1 pl.
- Tabor, R. W.; Mark, R. K.; Wilson, F. H., 1985, Reproducibility of the K-Ar ages of rocks and minerals—An empirical approach: U.S. Geological Survey Bulletin 1654, 5 p.
- Tabor, R. W.; Zartman, R. E.; Frizzell, V. A., Jr., 1987, Possible tectonostratigraphic terranes in the North Cascades crystalline core, Washington. In Schuster, J. E., editor, 1987, *Selected papers on the geology of Washington*: Washington Division of Geology and Earth Resources Bulletin 77, (this volume).
- Thompson, M. L.; Wheeler, H. E.; Danner, W. R., 1950, Middle and Upper Permian fusulinids of Washington and British Columbia: *Cushman Foundation for Foraminiferal Research Contributions*, v. 1, pts. 3-4, no. 8, p. 46-63.
- Turner, D. L.; Frizzell, V. A. Jr.; Triplehorn, D. M.; Naeser, C. W., 1983, Radiometric dating of ash partings in coals of the Eocene Puget Group, Washington—Implications for paleobotanical stages: *Geology*, v. 11, no. 9, p. 527-531.
- Vance, J. A., 1957, The geology of the Sauk River area in the northern Cascades of Washington: University of Washington Doctor of Philosophy thesis, 312 p.
- Vance, J. A., 1975, Bedrock geology of San Juan County. In Russell, R. H., editor, 1975, *Geology and water resources of the San Juan Islands, San Juan County, Washington*: Washington Department of Ecology Water-Supply Bulletin 46, p. 3-19.
- Vance, J. A., 1977, The stratigraphy and structure of Orcas Island, San Juan Islands. In Brown, E. H.; Ellis, R. C., editors, 1977, *Geological excursions in the Pacific Northwest: Geological Society of America field guide, annual meeting, Cordilleran section* [Seattle], p. 170-203.
- Vance, J. A., 1982, Cenozoic stratigraphy and tectonics of the Washington Cascades [abstract]: *Geological Society of America Abstracts with Programs*, v. 14, no. 4, p. 241.
- Vance, J. A.; Dungan, M. A., 1977, Formation of peridotites by deserpentinization in the Darrington and Sultan areas, Cascade Mountains, Washington: *Geological Society of America Bulletin*, v. 88, no. 10, p. 1497-1508.
- Vance, J. A.; Dungan, M. A.; Blanchard, D. P.; Rhodes, J. M., 1980, Tectonic setting and trace element geochemistry of Mesozoic ophiolitic rocks in western Washington: *American Journal of Science*, v. 280-A, part 1, p. 359-388.
- Vance, J. A.; Miller, R. B., 1981, The movement history of the Straight Creek fault in Washington State [abstract]: *Geological Association of Canada Cordilleran Section, Program and Abstracts for 1981 Meeting*, p. 39-41.
- Vine, J. D., 1969, Geology and coal resources of the Cumberland, Hobart, and Maple Valley quadrangles, King County, Washington: U.S. Geological Survey Professional Paper 624, 67 p.

- Weaver, C. E., 1912, Geology and ore deposits of the Index mining district: Washington Geological Survey Bulletin 7, 96 p.
- Whetten, J. T.; Jones, D. L., 1981, Age, character, and time of accretion of the Trafton unit, northwest Washington [abstract]: Geological Society of America Abstracts with Programs, v. 13, no. 2, p. 114.
- Whetten, J. T.; Zartman, R. E.; Blakely, R. J.; Jones, D. L., 1980, Allochthonous Jurassic ophiolite in northwest Washington: Geological Society of America Bulletin, Part 1, v. 91, no. 6, p. 359-368.
- Yeats, R. S., 1958a, Geology of the Skykomish area in the Cascade Mountains of Washington: University of Washington Doctor of Philosophy thesis, 243 p.
- Yeats, R. S., 1958b, Geology of the Skykomish area in the Cascade Mountains of Washington [abstract]: Dissertation Abstracts, v. 19, no. 4, p. 775.
- Yeats, R. S., 1964, Crystalline klippen in the Index district, Cascade Range, Washington: Geological Society of America Bulletin, v. 75, no. 6, p. 549-562.
- Yeats, R. S.; Engels, J. C., 1971, Potassium-argon ages of plutons in the Skykomish-Stillaguamish areas, north Cascades, Washington: U.S. Geological Survey Professional Paper 750-D, p. D34-D38.

Table A1.—New U-Th-Pb ages of zircons from rocks in the eastern and western melange belts, 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 $^{204}\text{Pb}:^{206}\text{Pb}:^{207}\text{Pb}:^{208}\text{Pb} = 1:18.40:15.60:38.20$

| Sample number | Rock type | Size fraction (mesh sizes) | Concentration (ppm) | | | Isotopic composition of lead (atom %) | | | | Age (Ma) | | | |
|---------------|---------------------------------|----------------------------|---------------------|-------|--------|---------------------------------------|-------------------|-------------------|-------------------|-------------------|-------------------|-------------------|-------------------|
| | | | U | Th | Pb | ^{204}Pb | ^{206}Pb | ^{207}Pb | ^{208}Pb | ^{206}Pb | ^{207}Pb | ^{207}Pb | ^{208}Pb |
| | | | | | | | | | | ^{238}U | ^{235}U | ^{206}Pb | ^{232}Th |
| RWT-513-80 | Metatonalite in eastern melange | -150,+200 | 283.2 | 104.1 | 2.003 | 0.1786 | 77.94 | 6.313 | 15.57 | 39.4 | 39.8 | 62.6 | 37.9 |
| | | -250,+325 | 290.9 | 104.2 | 2.477 | 0.1874 | 77.51 | 6.436 | 15.87 | 47.0 | 47.0 | 47.5 | 46.6 |
| JTW-79-215 | Metatonalite in eastern melange | -200,+250 | 877.5 | 254.2 | 4.441 | 0.1995 | 78.14 | 6.657 | 15.00 | 28.1 | 28.7 | 77.2 | 29.1 |
| | | -250,+325 | 684.2 | 205.7 | 4.276 | 0.2163 | 77.60 | 6.873 | 15.31 | 34.4 | 34.9 | 73.3 | 33.0 |
| JTW-79-216 | Metagabbro in eastern melange | -150,+200 | 337.6 | 117.8 | 9.036 | 0.0369 | 83.86 | 4.714 | 11.39 | 164.7 | 165.9 | 183.4 | 173.4 |
| | | -250,+325 | 385.8 | 156.4 | 9.860 | 0.0214 | 84.02 | 4.481 | 11.48 | 158.2 | 159.3 | 175.9 | 152.3 |
| RWT-278-80 | Metatonalite in western melange | -150,+200 | 685.1 | 375.9 | 17.257 | 0.0354 | 80.04 | 4.468 | 15.46 | 148.1 | 149.0 | 163.3 | 146.7 |
| | | -250,+325 | 629.9 | 324.4 | 15.714 | 0.0173 | 81.41 | 4.266 | 14.31 | 149.8 | 150.5 | 161.1 | 149.8 |
| RWT-316-80 | Metatonalite in western melange | -150,+200 | 310.2 | 139.7 | 7.444 | 0.0090 | 83.56 | 4.252 | 12.18 | 148.2 | 149.1 | 162.3 | 142.0 |
| | | -250,+325 | 337.5 | 144.6 | 8.098 | 0.0143 | 83.60 | 4.349 | 12.04 | 148.1 | 149.5 | 172.0 | 145.9 |

Table A2.—New K-Ar ages from rocks in the eastern and western melange belts, Washington

(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 derived from an empirically derived curve relating coefficient of variation in the age to percent radiogenic argon)

| Sample number | Mineral | Rock | K ₂ O percent | ⁴⁰ ArRad (moles/gm x 10 ⁻¹⁰) | ArRad (percent) | Age (m.y.) |
|---------------|------------|--------------|--------------------------|---|--------------------|---------------|
| RWT | Hornblende | Metagabbro | 0.153; .132; .136; .132 | 0.200 | 33.8 | 96.3 ± 6.1 |
| JTW 76-199 | Hornblende | Metagabbro | 0.193; .191; .195; .189 | .337 | 20.5 | 118.0 ± 10.6 |
| RWT 372-80 | Sericite | Phyllite | 8.9; 8.9; | 6.21 | 79.5 | 47.8 ± 1.1 |
| DB 81-568 | Hornblende | Granodiorite | 0.535; .537; .529; .543 | .364 | 31.2 | 46.6 ± 2.8 |

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MAJOR CENOZOIC FAULTS IN THE NORTHERN PUGET LOWLAND OF WASHINGTON

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ABSTRACT

Regional geological studies and geophysical surveys (especially seismic reflection profiling) have discovered several Eocene and younger faults in the northern Puget Lowland that were unknown a dozen years ago. The southeast-striking Devils Mountain fault near the North Fork of the Stillaguamish River definitely cuts the mid-Eocene Chuckanut Formation. Geophysical surveys failed to find a western continuation of the Devils Mountain fault but found instead the Northern Whidbey Island fault. The 500-meter thickness of Quaternary sediments south of this fault could indicate very young movement.

The previously undescribed northwest-trending Mount Vernon fault is the boundary between the northern Puget Lowland and the Cascade Range. It dextrally offsets by 47 kilometers the Devils Mountain fault and the pre-Tertiary terranes of the San Juan Islands. Seismic profiling near the International Border shows that the fault vertically offsets terrestrial Miocene strata by 2 kilometers.

The Haro fault in the northern San Juan Islands is inferred to be the offset part of the Devils Mountain fault; together they define a huge, post-Chuckanut, spoon-shaped thrust that extends from the San Juan Islands to at least the latitude of Snoqualmie Pass. The San Juan and Survey Mountain faults of Vancouver Island and the northwest-trending Southern Whidbey Island fault are inferred to be segments of a fault that seems to dextrally offset the thrust plate more than 100 kilometers. However, the major dextral fault could be the Leech River fault, which separates the Olympic terrane from the terranes previously accreted to North America. Oroclinal bending of the northwestern parts of the Mount Vernon, San Juan/Survey Mountain/Southern Whidbey Island, and Leech River faults has disguised their strike-slip nature. Miocene folding caused the Whatcom basin, Clallam syncline, and the antiform of the Olympic Mountains and must have affected the dips of the Eocene faults.

High-resolution seismic reflection profiling indicates that some of the buried major faults can be identified by offsets in Pleistocene and Holocene sediments. The offsets may have been generated during 140 to 350 meters of postglacial isostatic rebound. Earthquakes less than 25 kilometers deep do not define tight linear arrays that can be identified as active faults. Numerous faults probably are being reactivated during the present north-south compression.

INTRODUCTION

Major northwest- to north-trending faults, which cut Eocene or older rocks, are well known in northwestern Washington and southwestern British Columbia (Fig. 1). Most of these faults probably formed in response to the northward transport of the northeastern Pacific basin relative to the North American plate. The absence of major faults west of the Straight Creek fault is conspicuous. The purpose of this paper is to assemble the evidence for previously poorly recognized major faults in this area.

A major obstacle to recognizing such faults in the northern Puget Lowland is that the bedrock is largely obscured by luxuriant vegetation, widespread unconsolidated Quaternary sediments, and numerous bodies

of water. Considerable progress in fault recognition has been made since 1975, due, in part, to (1) the need to understand the seismic hazards to cities and proposed major engineering structures (Yount and Crosson, 1983) and (2) the growing realization that dextral transcurrent faults have segmented the western edge of the North American plate during the past 80 m.y. (Cowan, 1982; Johnson, 1984b).

Much of this paper is devoted to elucidating a northwest-striking, dextral strike-slip fault about 140 km long in the northern Puget Lowland. The fault is herein named the Mount Vernon fault. It marks the boundary between the Cascade Range and the northern Puget Lowland. This fault has not been described before, and its recognition is the key to understanding the structure of the northern part of the Puget Low-

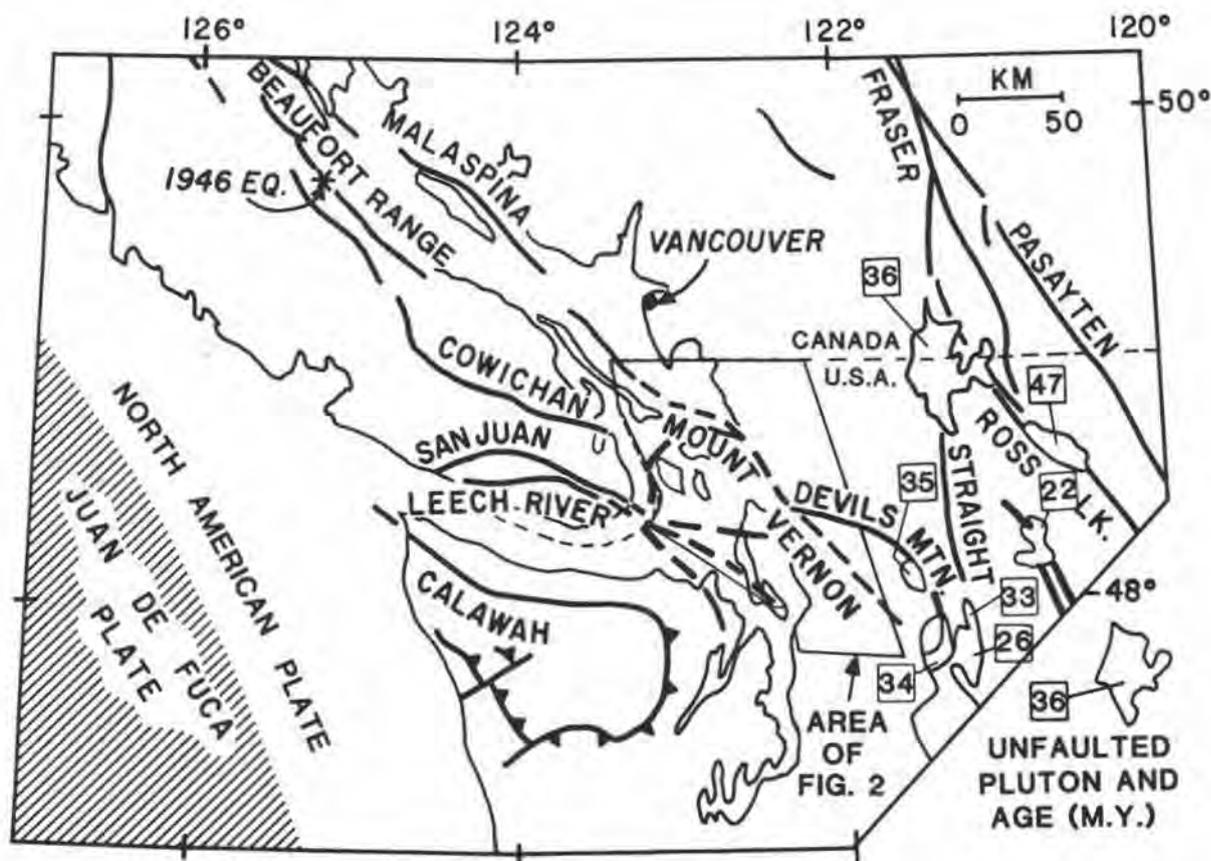


Figure 1.—Tectonic setting of the northern Puget Sound Lowland. Faults described in this paper are dashed. Sources of data: Figure 2 of this paper; Jackson (1976), Tabor and others (1978), Roddick and others (1979), Tipper and others (1981), Price and others (1981), and Fairchild and Cowan (1982).

land. The resultant pre-late Eocene structural model presented here has important implications for the geology of northwestern Washington and southern Vancouver Island.

Johnson (1984b) has summarized the regional geology of the northern Puget Lowland. The geology of the northwestern part of the area, the San Juan Islands, is described by Brandon and others (1983). Much of the regional geology will be reviewed below as each of the major faults is discussed.

MOUNT VERNON FAULT

Introduction

The Mount Vernon fault is not exposed. Its presence is inferred from regional geologic patterns, aeromagnetic and seismic reflection surveys, and the presence of nearby, presumably satellitic, northwest-trending faults. Most of these data are compiled on

Figure 2, which also emphasizes the lack of outcrop over large parts of the study area.

The possibility of a major fault coincident with the Mount Vernon fault was raised first by Rogers (1970, p. 54-55 and plate IX); he suggested that northwest-trending linear topographic features in the northern Puget Lowland are an "inferred fault or other major structural feature", but he provided very little geological evidence for this conclusion. W. A. Brewer (fig. 9b in Cheney, 1979) and Lawrence and Rosenfeld (1978) also identified lineaments coincident with the Mount Vernon fault.

The Mount Vernon fault originally was named the Bellingham Bay–Chaplain fault zone (Cheney, 1976). This name is herein abandoned because the faults in Bellingham Bay and at Lake Chaplain may not be temporally and spatially equivalent. The fault is now named for the city of Mount Vernon, where the Skagit River crosses the inferred trace of the fault.

Because the evidence for the Mount Vernon fault is most convincing near the International Border, the northern part of the study area will be discussed first. Figure 2 shows that several northwest-trending faults have been mapped within a 15-km-wide zone generally northeast of the Mount Vernon fault. Because references providing detailed descriptions of these mapped faults are listed in Cheney (1979), only a few will be described here. The interpretation of lineaments is so debatable that the aforementioned topographic lineaments will not be considered. The patterns of seismicity in the northern Puget Lowland will be reviewed, but they fail to unambiguously define major faults.

Evidence for the Mount Vernon Fault

Strait of Georgia

Few bedrock exposures are present in the Whatcom basin, which straddles the International Border between Bellingham and the Fraser River south of Vancouver, British Columbia. However, seismic reflection surveys by Mobil Oil Corporation in 1970 have provided a great deal of information about the offshore part of the basin. Figure 2 shows the locations of the survey lines. The seismic reflection profiles are still proprietary, but many of the interpretations are not.

The seismic profiles can be matched with the stratigraphic log of a well north of Point Roberts (Fig. 2). Table 1 includes a summary of this log by Hopkins (1966, 1968) for the dry well. Because the strata depicted on the nearby seismic lines are gently dipping, Dobrin (1975-1977) was able to correlate the seismic stratigraphy with the stratigraphic log (Table 1). The most prominent reflector on the seismic lines is the top of the upper Eocene unit (Dobrin, 1975-1977, p. 10). This unit probably is the continental, arkosic Huntingdon Formation described by Miller and Misch (1963) northeast of Bellingham. In the seismic reflection lines closest to the well, the terrestrial Miocene strata have a two-way travel time of 1.1 to 1.2 seconds; the maximum travel time through them, about 1.7 seconds, on the southern end of line W 70-2 just north of line W 70-7, indicates a maximum thickness of ≥ 1.3 km. Because these strata do not crop out on the mainland, their existence is not widely known.

On lines W 70-4, W 70-5, and W 70-7 the upper Eocene and Miocene strata are faulted against non-reflective rocks. The faults dip steeply northeast. Because the American and Canadian islands to the southwest (Fig. 2) are underlain by folded strata of the Cretaceous Nanaimo Group and Eocene Chuckanut Formation, the nonreflective units probably are these

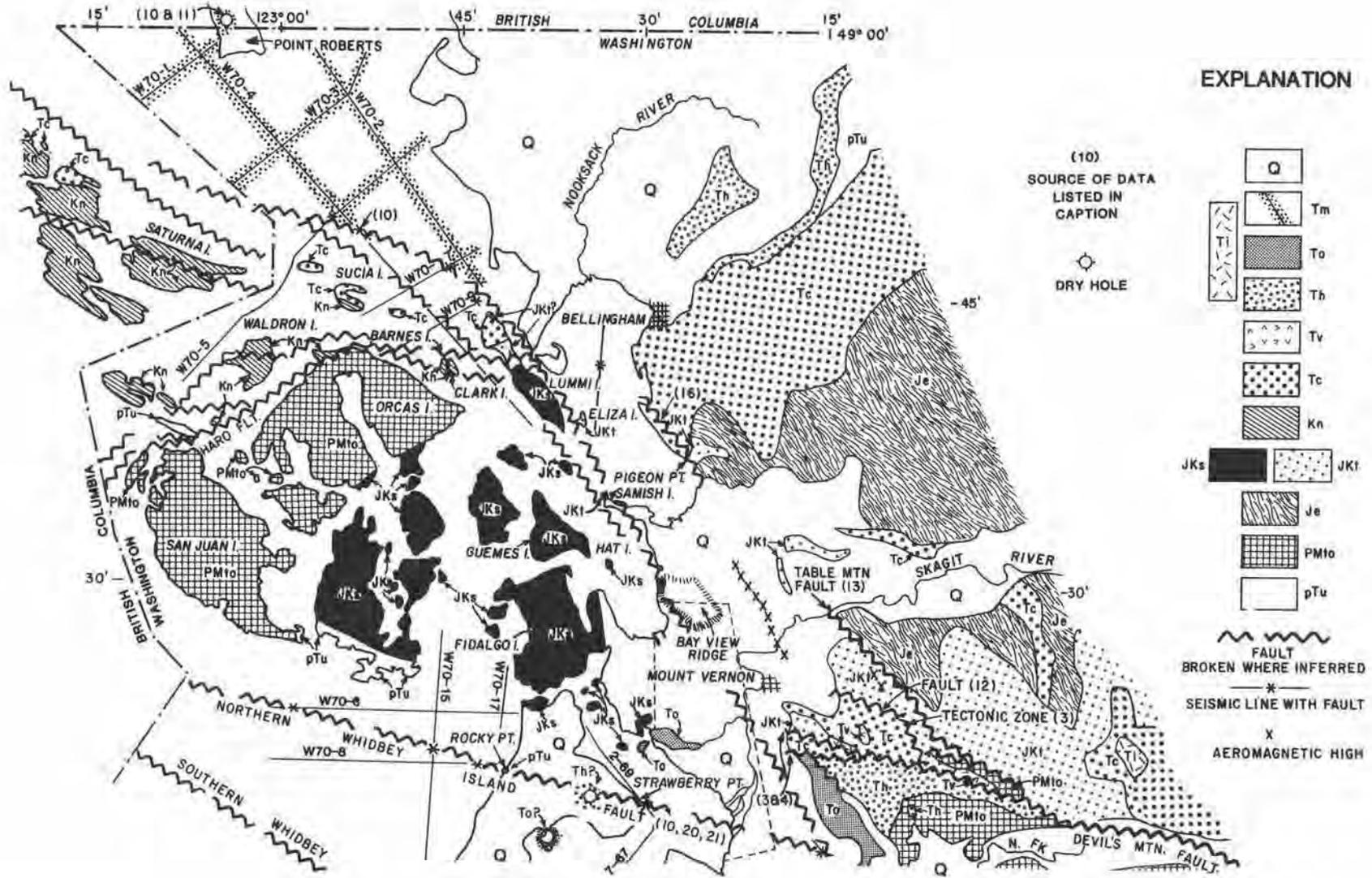
or older rocks. Thus, the Miocene strata are down-faulted to the northeast. Given the thickness of the Miocene strata on the south end of line W 70-2 and of the upper Eocene strata in the well at Point Roberts, the vertical throw on the fault is about 2 km. Because the Miocene and underlying strata are not broken by major faults elsewhere on the seismic lines, I have drawn a single northwest-trending fault through the faults on the three lines beyond the southern ends of lines W 70-2 and W 70-3 (Fig. 2).

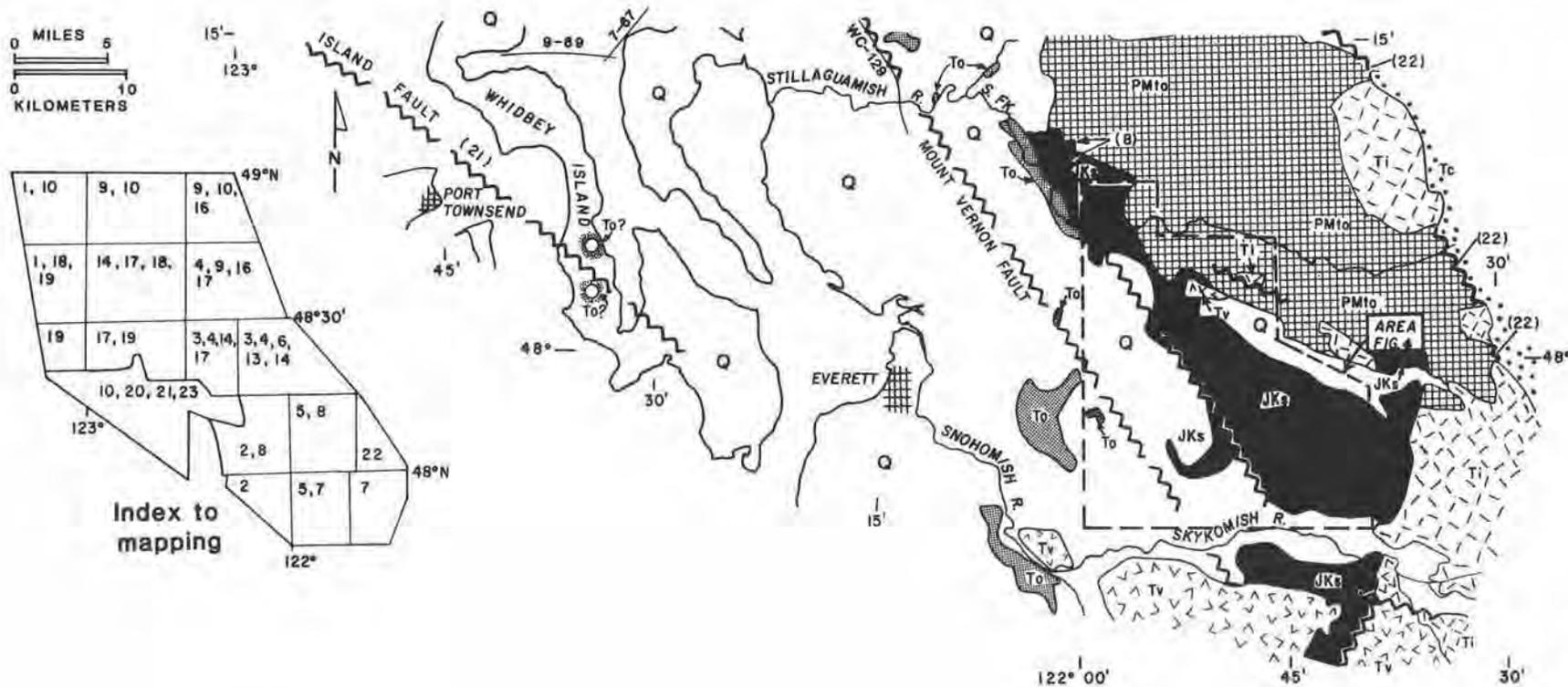
The southeast and northwest extent of this fault can best be defined by future seismic reflection surveys. Several authors (including Seraphim and others, 1976; Roddick and others, 1979; Price and others, 1981, fig. 50; Tipper and others, 1981; Brandon and others, 1983, fig. 2) already have shown one or more major northwest-trending faults on Vancouver Island or in the Strait of Georgia northwest of the International Border. Some of these faults are shown on Figure 1. Figure 2 shows the southern end of the one that Roddick and others (1979) extended from north of Staurna Island to the edge of the Quaternary sediments off the mouth of the Fraser River about 50 km to the north. Whether the Mount Vernon fault is continuous with a particular major fault in British Columbia is still unknown.

Offset of pre-Tertiary Units in the San Juan Islands and the Mainland

A number of tectonostratigraphic units shown by Brandon and others (1983, fig. 5A) in the San Juan Islands are combined into one (PMto) in Figure 2 and will be informally called the western terranes. These are mixtures of variously metamorphosed Paleozoic rocks (some with Tethyan fossils) and mixtures of Mesozoic rocks. The western terranes are structurally overlain by the Decatur terrane (JKs in Fig. 2), which consists of Jurassic ophiolitic rocks and Jura-Cretaceous turbidites (Whetten and others, 1980b; Brandon and others, 1983). The Nanaimo Group, consisting of unmetamorphosed predominantly Upper Cretaceous marine strata, is in fault contact with the western terranes in the northern San Juan Islands.

Danner (1957, 1966, 1977), Whetten (1978), Vance and others (1980), Dethier and others (1980), and Dethier and Whetten (1980) correlated rocks in the western terranes of the San Juan Islands with Danner's Olo Mountain and Trafton units on the mainland, and correlated the Decatur terrane with Danner's Sultan unit on the mainland, as shown in Figure 2. Whetten (1978) and Dethier and others (1980) suggested that the terranes in the San Juan





Sources of data: (1) Roddick and others, 1979; (2) Minard, 1980a, 1980b, 1980c, 1980d, 1981a, 1981b, 1981c; (3) Lovseth, 1975, (4) Bechtel, Inc., 1979; (5) Figure 4; (6) Jones, 1959; (7) Tabor and others, 1982; (8) Danner, 1957; (9) Johnson, 1984a; (10) Dobrin, 1975 to 1977; (11) Hopkins, 1966; (12) Jenkins, 1924; (13) Whetten and others, 1980a; (14) Whetten and others, 1979; Dethier and others, 1980; Dethier and Whetten, 1980; (15) Carroll, 1980; (16) Miller and Misch, 1963; (17) Brown and others, 1981; (18) Whetten and others, 1978; Johnson, 1978; Pacht, 1980; Vance (1975); (19) Brandon and others, 1983; (20) MacLeod and others, 1977; (21) Gower, 1980; Wagner and Wiley, 1983; (22) Vance and others, 1980; (23) Walsh, 1984. See Index to mapping.

Figure 2.—Tectonic map of the northern Puget Lowland. Blank areas have no bedrock outcrops. Thin dashed lines west of Mount Vernon enclose the area of Schreiner's gravity survey (1976). For clarity, all pre-Eocene faults have been omitted. Explanation of units: blank area, water; Q, Quaternary sediments, Tm, continental Miocene strata along seismic reflection lines; To, upper part of the rocks of Bulson Creek plus other upper Eocene to lower Oligocene marine strata; Th, Huntingdon Formation plus lower part of the rocks of Bulson Creek; Tv, rhyolitic intrusions plus volcanic rocks of Mount Persis; Tc, Chuckanut and Gabriola Formations; Kn, Nanaimo Group not including the Gabriola Formation; JKs, Sultan Group and Decatur terrane; JKt, rocks of Table Mountain; Je, Darrington Phyllite and Shuksan Greenschist; PMto, Trafton and Olo Mountain units plus the western terranes of the San Juan Islands; pTu, undifferentiated pre-Tertiary rocks. The numbers in parentheses indicate which of the references listed below discuss the indicated fault.

Table 1.—Dobrin's correlation (1975-1977) of seismic reflections with the stratigraphic log made by Hopkins (1966, 1968) of the dry well at Point Roberts

| Richfield-Pure Pt. Roberts well (depth in feet) | Northeastern end line W70-1 (depth in seconds) | Southwestern end line W70-1 (depth in seconds) | Line W70-5 at cross- ing with line W70-4 (depth in seconds) |
|---|--|--|---|
| | 0.0 | 0.0 | 0.0 |
| | Sea water | Sea water | Sea water |
| 0 | 0.1 | 0.13 | 0.26 |
| Pleistocene | Pleistocene | Pleistocene | Pleistocene |
| 847 | 0.25 | 0.30 | 0.51 |
| Miocene | Miocene | Miocene | Miocene |
| 4,300 | 1.45 | 1.20 | 2.0 |
| Upper Eocene | Upper Eocene | Upper Eocene | Upper Eocene |
| 6,200 | | | |
| Lower Eocene | | | |
| 7,290 | | | |
| Cretaceous | | | |
| 10,050 | Same pattern to >2.4 seconds | Same pattern to >2.0 seconds | Same pattern to >2.6 seconds |
| Upper Cretaceous-- Santonian | | | |
| 11,050 | | | |
| Lower Cretaceous | | | |
| 14,792 | | | |

Islands were displaced onto the mainland by the east-striking Devils Mountain fault. Because the Olo Mountain and Trafton sequences lie east of the Sultan unit on the mainland (Fig. 2), they cannot be the former southward continuation of the belt of rocks on the San Juan Islands. The interpretation offered in Figure 2 is that the terranes on the San Juan Islands are the western limb of a southeast-plunging synform, and the rocks on the mainland are the eastern limb of that synform that has been dextrally displaced by the Mount Vernon fault.

Truncation of the Devils Mountain Fault

The dextral displacement suggested above can be tested by examining the Devils Mountain fault. The west end of that fault was first described by Hobbs and Pecora (1941). North of the North Fork of the Stillaguamish River, the fault is straight and presumably vertical, or nearly so. Here the fault zone is as much as 2 km wide (Lovseth, 1975; Bechtel, Inc., 1979; Dethier and others, 1980; Dethier and Whetten, 1980). It cuts the Eocene Chuckanut Formation and

the slightly younger lower part of the rocks of Bulson Creek of Marcus (1981). South of the North Fork of the Stillaguamish River, the fault curves southeastward (Whetten and others, 1980b, fig. 1) and is intruded by the Squire Creek batholith (35 m.y.) (Vance and others, 1980) as shown on Figure 2. Thus, the length of the arcuate Devils Mountain fault is at least 60 km, and the last movement of the fault is definitely younger than the Chuckanut Formation.

Lovseth (1975), Whetten (1978), and Whetten and others (1980b) recognized that the fault is a major one and inferred that it extends in the subsurface to at least 123° W., south of San Juan Island. Whetten and his colleagues suggested that this is the fault that sinistrally displaced the terranes of the San Juan Islands onto the mainland.

However, Dobrin (1975 to 1977) noted that deep seismic exploration profiles obtained by Mobil Oil Corporation east and west of Whidbey Island (Fig. 2) failed to find a continuation of the Devils Mountain fault. According to Dobrin, sparker, Acoustipulse, and modified Acoustipulse with concurrent magnetic and gravimetric surveys in the same area also failed to find the fault. Schreiner (1976) conducted a gravity survey of the Skagit delta (Fig. 2) and found north-south residual anomalies, instead of the east-west ones that might be expected if the Devils Mountain fault continues westward.

The only evidence for a westward continuation of the Devils Mountain fault is the west-striking gradient on the aeromagnetic map of Whetten and others (1980b), reproduced here in Figure 3. Inspection of the large magnetic high that straddles lat. $48^{\circ}30'$ N. suggests that the aeromagnetic map could indicate that the northwest-trending Mount Vernon fault truncates the west-trending Devils Mountain fault. A prominent southeast-trending low (less than 200 nT) is present in the northern part of the high, and the areally smaller but more intensely magnetic knobs (which are characteristic of the rest of the high) do not interrupt the strike of this trend to the southeast. East of this southeast trend, the aeromagnetic grain is west striking (similar to the structural grain of the Devils Mountain fault), but to the west the magnetic grain trends more to the north (over the ophiolitic rocks of Fidalgo Island). In addition, the areally smaller highs greater than 300 nT tend to be farther northward west of the southeast-trending magnetic low than east of it.

Dobrin (1975) reported that seismic reflection lines east and west of Whidbey Island indicate a large fault (south side down) about 6 km south of the westward projection of the Devils Mountain fault. Although this fault is not readily apparent on the aeromagnetic map (Fig. 3), it does have a slight

expression on the magnetic and gravimetric maps of MacLeod and others (1977, figs. 3 and 4, respectively), which were derived from ship-borne surveys. Gower (1980) called this the Northern Whidbey Island fault.

Pre-Tertiary foliated sandstone and argillite crop out north of the Northern Whidbey Island fault at Rocky Point (Gower, 1980). A well on, or just south of, the fault penetrated 190 m of Quaternary sediments (Walsh, 1984), but a well 4 km south of the fault penetrated 500 m of Quaternary sediments (Gower, 1980). Thus, some displacement on the Northern Whidbey Island fault may be quite young. Cheney (1979) and Gower (1980) noted that Quaternary sediments older than 18,000 yr B.P. are tilted and faulted over the trace of the fault at Strawberry Point. Pleistocene sediments along the fault between Whidbey and Vancouver Islands also are displaced (Wagner and Wiley, 1983, fig. 11).

No Tertiary fault that could be an eastern extension of the Northern Whidbey Island fault is known 5 to 10 km south of the Devils Mountain fault in the foothills on the mainland, but an intercept on line WC-129 could be the Northern Whidbey Island fault (Dobrin, 1975-1977). The Northern Whidbey Island fault could hardly be the Devils Mountain fault, unless the Devils Mountain fault is offset by some other fault. Cheney (1979) made exactly this suggestion, illustrating sinistral offset along what is here called the Mount Vernon fault. However, because dextral faults have been so common on the western margin of the continent during the last 80 m.y. (Cowan, 1982; Johnson, 1984b), sinistral offset seems improbable. Cheney (1977), MacLeod and others (1977, fig. 2), and Wagner and Wiley (1983, fig. 11) showed the Northern Whidbey Island fault striking westward toward the San Juan-Survey Mountain fault of Vancouver Island.

Offset of the Devils Mountain Fault

Because the Devils Mountain fault is north of the Sultan unit and truncates the Trafton and Olo Mountains units, the place to look for the offset part of the Devils Mountain fault is north of the equivalent terranes in the San Juan Islands. Carroll (1980) described such a west-trending fault in the center of Lummi Island, 47 km to the northwest (Fig. 2). About 0.1 to 0.6 km north of this fault are foliated mafic and trondjemitic rocks; rocks of similar composition (Dethier and others, 1980) and age (Whetten and others, 1980b, table 2) are present in the Devils Mountain fault zone. If these rocks are included in the fault zone on Lummi Island, as in Figure 2, the zone is 1 km or more wide. Lithologies similar to the exotic rocks in

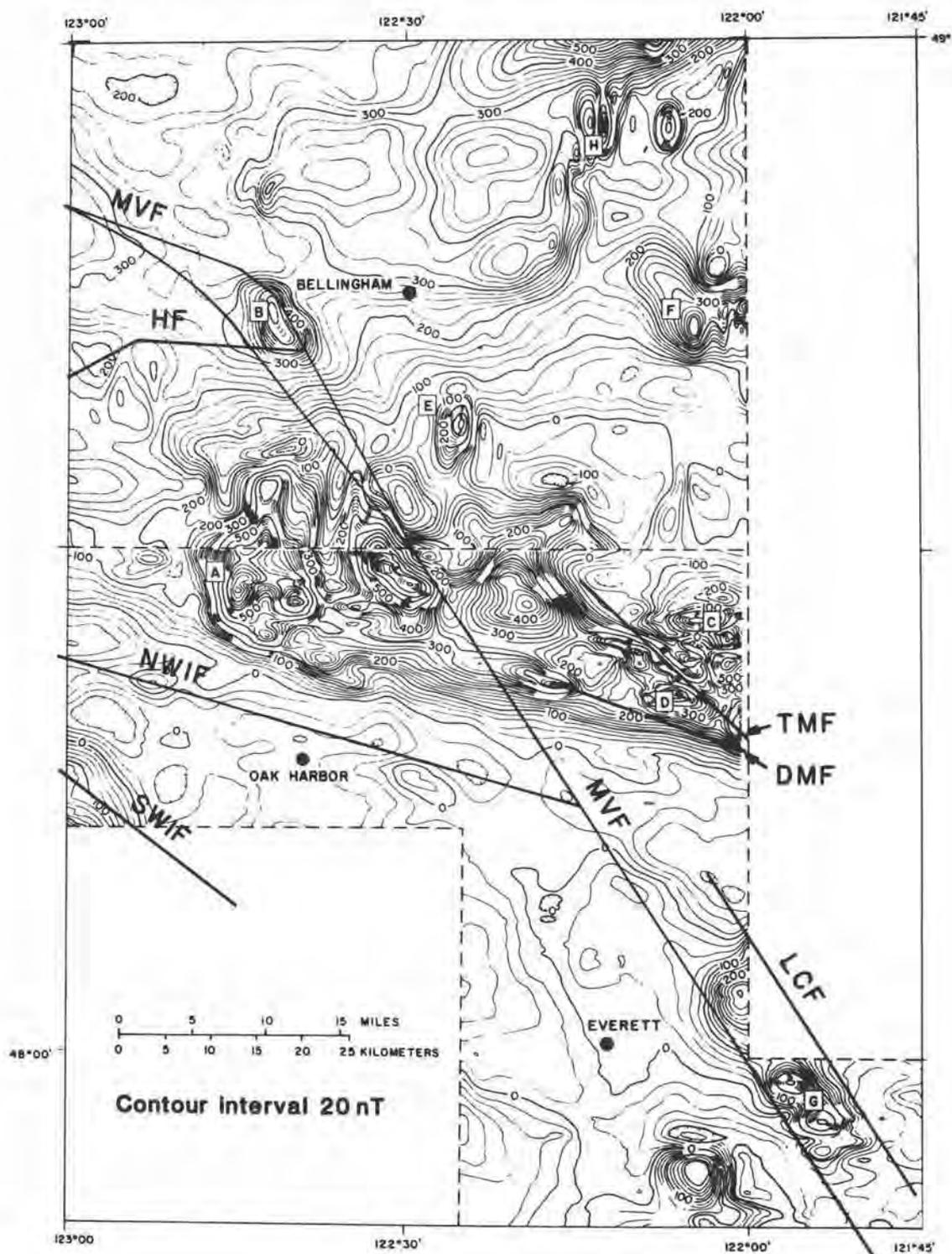


Figure 3.—Aeromagnetic map of part of the northern Puget Lowland. The map (without the faults) was published as figure 2 of Whetten and others (1980b). These authors reported that the data consist of two surveys by the U.S. Geological Survey—north of 48°30'N. flown east-west, and south of 48°30' N. flown north-south. Both surveys were flown at 3.2-km spacing and 0.91-km altitude. Letters in boxes are localities discussed by Whetten and others. Faults: DMF, Devils Mountain fault; HF, Haro fault; LCF, Lake Chaplain fault; MVF, Mount Vernon fault; NWIF, Northern Whidbey Island fault; SWIF, Southern Whidbey Island fault; TMF, Table Mountain fault.

the Devils Mountain fault on Lummi Island occur in the rocks of Table Mountain (Whetten and others, 1979; Dethier and others, 1980) and in the Sultan unit (R. W. Tabor, USGS, written commun., May 1983); one, or both, of these units is cut by each of the faults. In summary, the fault on Lummi Island has the same strike, width, and types of exotic rocks and, as shown below, probably truncates the same rock units on the south as the Devils Mountain fault. In addition, the Chuckanut Formation is restricted to the north side of both faults.

To the west of Lummi Island a major fault is present along the northern edge of Orcas and San Juan Islands. Vance (1977, p. 188) named this the Haro fault. Figure 1 of Cowan and others (1977) and figure 5A of Brandon and others (1983) show that this fault places the unmetamorphosed Nanaimo Group, the Upper Jurassic and Lower Cretaceous Spieden Group on Spieden Island (Johnson, 1981), and the Triassic Haro Formation on the northern tip of San Juan Island against the western terranes of the San Juan Islands on the south.

The Haro fault zone is not on strike with the east-west fault zone on Lummi Island. The explanation preferred here is that because both zones bound the terranes of the San Juan Islands, they are parts of the same arcuate fault. Figure 2 shows that by this interpretation the Haro fault truncates virtually the entire belt of southeast-striking strata of the Nanaimo Group. The Spieden Group, the Haro Formation, and the east to northeast strikes of Nanaimo strata on Cactus Island and Flattop Island north of Spieden Island (Johnson, 1978), on the southern part of Waldron Island (Ward, 1978; Pacht, 1980), and on Clark and Barnes Islands (Vance, 1975) are interpreted here as blocks within the same fault zone. The maximum width of this zone is almost 4 km. Although the contacts are covered by seawater, the Haro fault zone is considered to consist of south-dipping thrusts or reverse faults (Cowan and others, 1977; Vance, 1977; Whetten and others, 1978; Johnson, 1978; Brandon and others, 1983).

Because this inferred arcuate fault contains the same types of meta-igneous rocks and bounds the same terranes on the south as the Devils Mountain fault, the Haro fault and Devils Mountain fault are inferred to be offset segments of the same arcuate fault (Devils Mountain-Haro fault). If so, blocks of Nanaimo strata might yet be found in the Devils Mountain fault interspersed with the more abundant blocks of the lithologically similar Chuckanut Formation. Figure 2 illustrates that the Chuckanut Formation and its probable correlative, the Gabriola Formation of the Nanaimo

Group, are restricted to the northern sides of the Devils Mountain and Haro faults.

Other Regional Patterns

Other units of regional extent in Figure 2 have distributions that seem to be explained by the Mount Vernon fault. Lovseth (1975) described a melange of phyllites, greenstones, and serpentinites on the northeast side of the Devils Mountain fault as the rocks of Table Mountain. Whetten and others (1980b) showed that these rocks are widespread south of the Skagit River and are coeval with the Decatur terrane. However, the greenstones in the Decatur terrane on Fidalgo Island were not metamorphosed in the same high-pressure, low-temperature environment as the rocks of Table Mountain (Brown and others, 1981; Whetten and others, 1980b), and the turbidites in the Decatur terrane are not as phyllitic as those in the rocks of Table Mountain (Whetten and others, 1980b). Brown and others (1981) suggested that a major fault is present between these units. That fault is herein considered to be the Mount Vernon fault.

The distinction between the rocks of Table Mountain and the Decatur terrane defines the Mount Vernon fault in the easternmost San Juan Islands (Fig. 2). Samish Island is part of the rocks of Table Mountain (Brown and others, 1981). Because the pelitic rocks of Eliza Island are more phyllitic and better cleaved than those on Lummi Island (Carroll, 1980), they probably are in or east of the fault zone. Aragonite-bearing greenstones are typical of the rocks of Table Mountain (Whetten and others, 1980b; Brown and others, 1981); therefore, the presence of such fault-bounded rocks on the southeastern part of Lummi Island (Carroll, 1980, p. 29, 60), which is otherwise composed of rocks of the Decatur terrane, implies that the main strand of the Mount Vernon fault is adjacent to the northeastern side of Lummi Island. The pelitic rocks of southern Lummi, Eliza (Carroll, 1980), and Samish Islands also have prominent northwest-trending kink bands, which may indicate proximity to a major northwest-trending fault.

Other regionally extensive units are restricted to either the foothills of the Cascade Range on the northeast or to the Puget Lowland to the southwest (Fig. 2). The Jurassic Darrington Phyllite and Shuksan Greenschist and the Eocene continental, arkosic, Chuckanut and Huntingdon Formations are not present in the Lowland. In contrast, the Tertiary marine strata are uncommon in the foothills. The Cretaceous Nanaimo Group is restricted to the islands near the International Border. Each of these restricted distributions

might be explained by individual depositional or tectonic patterns. However, the southwest or northeast limit of each of these regional units defines a northwest-trending zone amenable to a single (common) explanation; each may be caused by one or more northwest-striking faults. Accordingly, the regional distributions of these formations are regarded as additional evidence for the Mount Vernon fault.

Another reason for concluding that the Mount Vernon fault displaces the Chuckanut is that the Mount Vernon fault is inferred to dextrally displace the Devils Mountain fault, which definitely cuts the Chuckanut. Nonetheless, west of Bellingham the Chuckanut strata have been interpreted as extending essentially unbroken from Sucia Island and the northern end of Lummi Island to the mainland (Johnson, 1984a). Yet, some interesting differences do occur on the mainland and Lummi Island: Johnson (1984a) showed that on the mainland the basal Bellingham Bay Member of the Chuckanut Formation overlies the Darrington Phyllite and, at Pigeon Point, the rocks of Table Mountain. On Lummi Island, he mapped the stratigraphically higher Padden Member unconformably over rocks that Whetten and others (1980b) concluded were the rocks of Table Mountain or the Decatur terrane. However, on the mainland, Johnson (1982, 1984a) did not recognize any unconformities within strata of the Chuckanut Formation and did not find the Padden Member overlying pre-Tertiary rocks.

These differences are permissive evidence for a fault between Lummi Island and the mainland. The rocks on Lummi Island may be basal Chuckanut (originally a more distant facies of the Bellingham Bay Member) that texturally and compositionally resembles the part of the Padden Member on the mainland. The seemingly obvious place to test this hypothesis is south of the Skagit River, but in that area, the Chuckanut strata are in fault contact, not depositional contact, with the rocks of Table Mountain (Whetten and others, 1980a; Dethier and Whetten, 1980).

Similarly, because Johnson (1984a) shows the west-trending fault in the middle of Lummi Island extending northeastward to the mainland, one could question whether a major northwest-striking fault occurs between Lummi Island and the mainland. Johnson postulated a northeast extension of the east-west fault because he inferred that the conglomeratic Governors Point Member of the Chuckanut Formation, which underlies the Padden Member on the mainland, was derived from a nearby and active fault block. He did not provide any other geological or geophysical evidence for such a northeast-trending fault.

Area South of the Devils Mountain Fault

Bedrock along the inferred trace of the Mount Vernon fault south of the Devils Mountain fault is poorly exposed, and only seismic line WC-129 crosses the area. Dobrin (1975-1977, fig. 2) showed a fault on this line and chose to connect it with the Northern Whidbey Island fault; however, the fault could equally well be the Mount Vernon fault. Seismic reflection profiles would be a good method of determining the location of the Mount Vernon fault and the Northern Whidbey Island fault south of the Devils Mountain fault.

Due to limited outcrops, the only place to study the geology of the southern part of the Mount Vernon fault or related faults is the Lake Chaplain-Granite Falls area shown on Figure 4. The mafic rocks are part of a widespread Jurassic ophiolitic suite (Vance and others, 1980; Whetten and others, 1980b). The meta-volcanic rocks are assumed to be part of the ophiolitic suite. The sedimentary rocks are part of the Jura-Cretaceous Sultan unit of Danner (1957, 1966) and the pre-Tertiary argillites and slaty argillites and phyllites of Tabor and others (1982). In the equivalent Decatur terrane of the San Juan Islands, the ophiolitic and volcanic rocks are overlain by a thick sequence of well-bedded lithic sandstone and mudstone (Brandon and others, 1983).

Tabor and others (1982) described the rocks around Lake Chaplain as a melange. Nonetheless, in the hydroelectric tunnel of the Snohomish County Public Utility District, steeply dipping faults, which locally are hundreds of meters wide, separate wider intervals of weakly sheared to unshaped rocks. Likewise, at the Lockwood prospect, a folded, pyritic felsic phyllite can be mapped for at least 570 m within a fault-bounded block approximately 425 m wide. Most of the other contacts in Figure 4 may also be tectonic, but as they have not been (and possibly cannot be) mapped in detail, they are not shown as faults.

A fault (which was not recognized by Tabor and others, 1982) passes northwestward through Lake Chaplain (Fig. 4). The width of brecciated and sheared rock at the dam on the south end of the lake is at least 400 m. Most slickensides rake less than 25° from the horizontal. In the Menzel Lake area 10 km to the north, the argillites are sheared, foliated, and interleaved with bodies of serpentinite and silica-carbonate rocks (altered serpentinites), which range in size from hand specimens to mappable units. The fault zone is greater than 1 km wide.

The faults in the tunnel and at Lake Chaplain could be pre-Tertiary and thus unrelated to the Mount Vernon fault. However, Wiebe (1963) mapped northwest-striking faults cutting the 48.8 ± 1.4 -m.y.-old Pilchuck Mountain pluton, in the northern part of Figure 4. The Granite Falls stock (44.4 ± 1.4 -m.y. K-Ar date tabulated by Engels and others, 1976, corrected by table 2 of Dalrymple, 1979) and hornfels are cut by a northwest-trending fault and numerous small shear zones that could be on strike with the Lake Chaplain fault. On the northern bank of the South Fork of the Stillaguamish River northwest of Granite Falls, a covered interval of 75 m separates the hornfels from arkosic sandstone that contains locally abundant fossil pelecypods. The arkosic rocks are presumed to be correlative with the compositionally similar strata that Danner (1957) described 3 km to the northwest as the Riverside formation. Both Danner (1957) and Addicott (*in* Minard, 1981a, 1981b, 1981c) infer that these strata are Oligocene.

The northeast side of the 23-km-long, northwest-trending, aeromagnetic high (anomaly G of Fig. 3) is coincident with the Lake Chaplain fault of Figure 4. The abrupt southwest margin of the high suggests that the dioritic and andesitic rocks are truncated on the southwest. This linear is chosen as the trace of the Mount Vernon fault in Figure 2.

SOUTHERN WHIDBEY ISLAND FAULT

The previous discussion of the Mount Vernon fault led to descriptions of the Devils Mountain fault, the Haro fault, and the Northern Whidbey Island fault. Gower (1980) inferred one other major fault, a northwest-striking one under the central part of Whidbey Island (Fig. 2). This he named the Southern Whidbey Island fault. He postulated this fault on the basis of a linear aeromagnetic high (part of which is shown in Fig. 3) and the southern edge of a gravity low. He noted the possible evidence for Quaternary displacement on this fault is that a dry hole 2 km northeast of the proposed fault penetrated 625 m of Quaternary sediments, but a well 0.5 km southwest of the trace of the proposed fault penetrated only 204 m. He extended the fault northwestward to the latitude of Port Townsend because high resolution seismic profiling by Wagner and Wiley (1983) showed evidence for Holocene faulting. On the basis of additional seismic profiling that showed offsets in Holocene and Pleistocene sediments, Wagner and Wiley (1983, p. 188, fig. 12) extended this fault to the International Border as two or three strands as much as 7 km apart (Fig. 2).

PATTERNS OF SEISMICITY

The only fairly accurately known earthquakes in northwestern Washington are those recorded by the telemetric seismograph network installed since 1973 (Noson and Crosson, 1980). According to R. S. Crosson (Univ. of Washington, personal commun., Oct. 1976), the published epicenters may have errors of 1 to 2 km relative to each other, and the error in the absolute location of the epicenters may be as great as 10 km in a few rare instances. In order to test whether the Mount Vernon fault or other faults might be active, I plotted Crosson's data for all earthquakes having magnitudes more than 2 and depths less than 25 km that had been recorded from 1973 through 1981 and all earthquakes of magnitudes more than 2.5 and less than 25 km deep from 1971 to December 1985 ($N = 113$). The epicenters fall in a 100-km-wide northwest-trending belt that is roughly centered along the Mount Vernon fault. Equally diffuse patterns are shown by Crosson (1983, fig. 1) for all earthquakes from 1970 to 1978 and by Rogers (1983, fig. 3) for all earthquakes from 1951 to 1969. No well-defined linear arrays of epicenters have emerged yet within the belt; therefore, the presently known epicenters cannot be used to independently define any of the major faults shown on Figure 2.

STRUCTURAL SYNTHESIS

This section integrates the previous descriptions of the major faults into a structural model for the northern Puget Lowland (Fig. 5). Obviously, a certain amount of speculation is involved, even though the text is written in a definitive style. The first step in constructing the structural model is to describe the history of the major faults.

History of the Major Faults

Devils Mountain Fault

Major movement on the Devils Mountain fault postdates the deposition of the Chuckanut Formation but predates the 47 km of displacement inferred on the Mount Vernon fault. The Devils Mountain fault also cuts rhyolitic rocks that intrude the Chuckanut Formation (Lovseth, 1975; Bechtel, Inc., 1979; Dethier and others, 1980; Dethier and Whetten, 1980) as shown on Figure 2. These rhyolitic rocks have yielded fission-track ages of 41.5 ± 3.4 m.y. (Lovseth) and 52.7 ± 2.5 m.y. (C. W. Naeser, *in* Dethier and Whetten, 1980). The Devils Mountain fault also cuts (Fig. 2) Narizian to Zemorrian marine rocks that contain

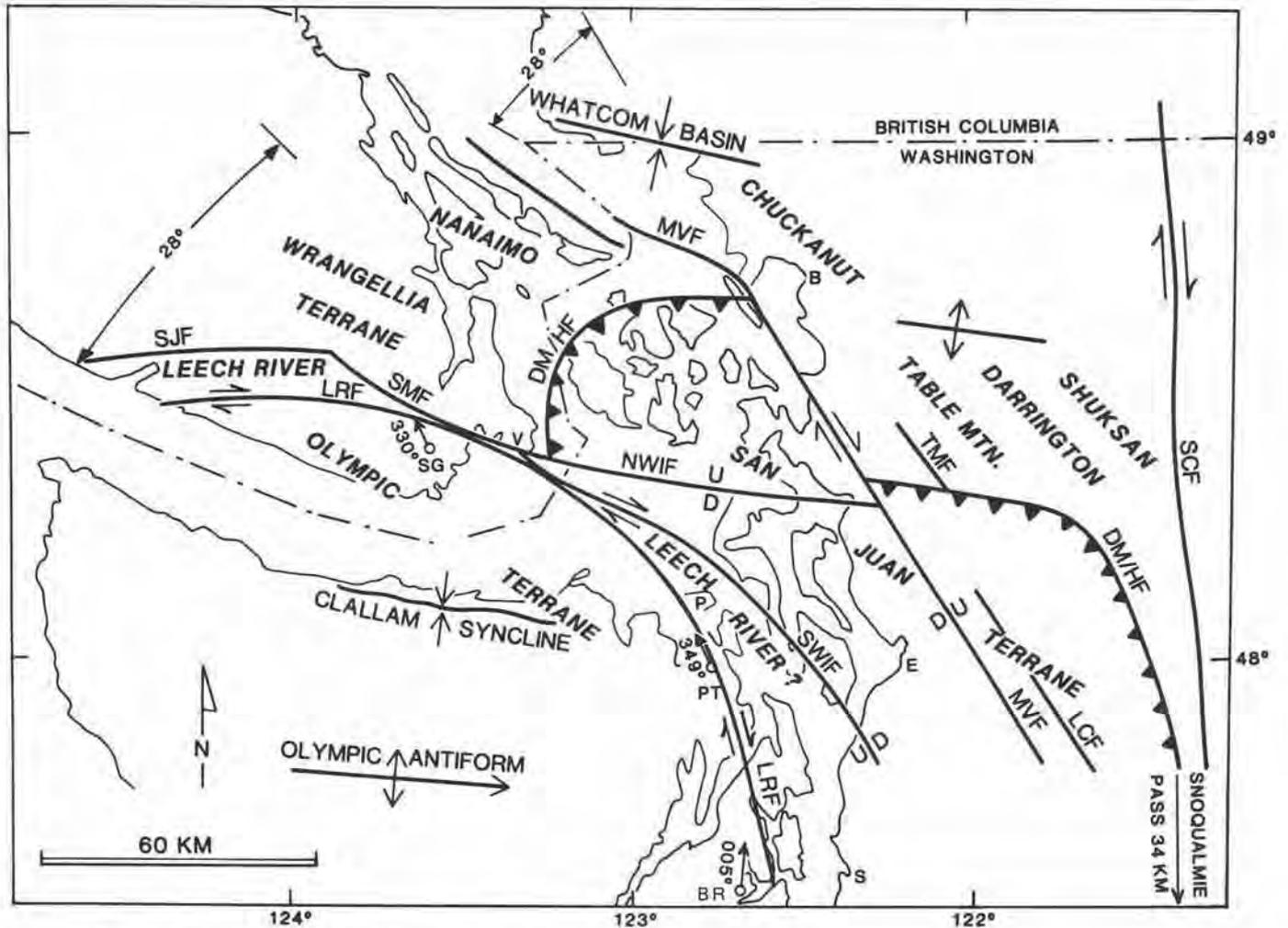


Figure 5.—Tectonic model for northwestern Washington and southern British Columbia. Pre-Eocene structures have been largely eliminated. Abbreviations for the faults are the same as those used in Figures 2 and 3, with the addition of LRF (Leech River fault), SJF (San Juan fault), and SMF (Survey Mountain fault). Rocks in the upper plate of the Devils Mountain/Haro fault probably occur at least as far south as Snoqualmie Pass, which is 34 km south of the illustrated end of Devils Mountain/Haro fault. Note the westward bending of the northwest-trending faults. SG and PT plus BR indicate the paleomagnetic results tabulated by Beck (1984) for the Sooke gabbro and by Beck and Engebretson (1982) for middle Eocene basalts, respectively. The positions of major Miocene folds are indicated. Other abbreviations are B, Bellingham; E, Everett; P, Port Townsend; S, Seattle; and V, Victoria.

clasts of the rhyolitic rocks (Lovseth, 1975; Marcus, 1981), but the Mount Vernon fault does not seem to dextrally displace these strata. Evidently, the Devils Mountain fault was reactivated in post-Narizian to Zemorrian time.

Mount Vernon Fault

The Mount Vernon fault dextrally offsets pre-Tertiary rocks and the Devils Mountain fault by 47 km and is inferred to offset the Chuckanut Formation. The oldest rocks that are not significantly offset are the predominantly andesitic rocks south of the

Skykomish River (Fig. 2) that Tabor and others (1982) referred to as the rocks of Mount Persis. Tabor and others (1982) reported dates of 41.7 ± 4.3 (K-Ar) and 47 ± 4 (fission track on apatite) m.y. from these rocks, and 35.7 ± 4.3 (K-Ar) m.y. for a phase of the Index batholith intrusive into them. The Oligocene marine rocks that are believed to overlie the volcanic rocks of Mount Persis (Tabor and others, 1982) also do not appear to be dextrally offset by the Mount Vernon fault.

These relationships indicate that strike-slip movement on the Mount Vernon fault is constrained by the age of the Chuckanut Formation and by the unfaulted

rocks of Mount Persis. Johnson (1984a) obtained a fission-track age of 49.9 ± 1.2 m.y. from a zircon in tuff near the top of the basal member of the Chuckanut Formation. Based on numerous fission-track ages from detrital zircons in the Chuckanut Formation, Johnson (1984a) concluded that the Chuckanut cannot be older than 55 m.y. The oldest dates yet reported for rocks intrusive into the Chuckanut Formation are 49.9 ± 1.1 m.y. (Bechtel, Inc., 1979) and 52.7 ± 2.5 m.y. (C. W. Naeser in Dethier and Whetten, 1980). Thus, at present, the minimum age of the dextrally displaced Chuckanut is 49.9 ± 1.2 m.y. and the oldest unfaulted rocks of Mount Persis are 47 ± 4 m.y. Whether the displacement of the hornfels or the 44 ± 1.4 -m.y.-old Granite Falls pluton is strike slip or dip slip is not yet known.

The youngest rocks known to be cut by the Mount Vernon fault are the Miocene strata of the Whatcom basin. Because the rocks of Mount Persis are unfaulted, displacement of these Miocene strata is inferred to be dip slip (>2 km). A review of Hopkins' original floral lists (1966) by J. C. Barnett (personal commun., Feb. 1982) indicates that the pollen from the Miocene strata penetrated by the dry hole on Point Roberts are not diagnostic of any particular part of the Miocene, but could be as young as late Miocene.

The paucity of evidence for Quaternary displacement on the Mount Vernon fault may, in part, be due to lack of detailed geologic mapping of the Quaternary sediments in the Puget Lowland, including the lack of appropriate seismic reflection surveys across the trace of the Mount Vernon fault. However, two areas may prove to record movement on the Mount Vernon fault. Figure 2 shows that the northeast and southwest sides of Bay View Ridge in the Skagit delta have pronounced northwest trends. Figure 3.41-3 and Appendix H of Bechtel, Inc. (1979) indicate that Bay View Ridge is underlain by Everson glaciomarine drift, deposited 13,500 to 11,000 yr B.P. (Hansen and Easterbrook, 1974). Secondly, according to Palmer (1977), post-Everson marine terraces are absent on Lummi Island but are present on the adjacent mainland, and strand lines occur at both lower and higher altitudes on Lummi Island than on the mainland. These anomalies suggest to Palmer that either the coastal environment was different on Lummi than on the mainland, or that tectonic or isostatic differences existed between Lummi Island and the mainland.

Precursors of the Mount Vernon fault may have been active prior to the Devils Mountain fault. The northwest-trending Table Mountain fault north of the Devils Mountain fault (Fig. 2) cuts pre-Tertiary rocks but not the Devils Mountain fault (Whetten and others, 1980a; Dethier and Whetten, 1980). The north-

west-trending tectonic zone mapped by Lovseth (1975) to the west of the Table Mountain fault does cut the Chuckanut and younger rhyolitic rocks but not the Devils Mountain fault; perhaps this tectonic zone is an older fault that was reactivated by movement on the Devils Mountain fault or the Mount Vernon fault.

Northern Whidbey Island Fault

The pre-Pleistocene history of the Northern Whidbey Island fault is uncertain because its relation to the Mount Vernon fault is obscure. If the Mount Vernon fault dextrally offsets it by about 50 km, it would be at the latitude of the Skykomish River (Fig. 2). Tabor and others (1982) did not recognize a major fault in this area, nor can their map be reinterpreted as indicating a major block down-dropped to the south. Thus, it seems unlikely that the Northern Whidbey Island fault is dextrally offset by the Mount Vernon fault.

Because the Northern Whidbey Island fault strikes westward toward the San Juan/Survey Mountain fault of Vancouver Island, the two faults might be one and the same. The San Juan fault cuts Leech River complex that has a metamorphic age of 40 to 44 m.y., and it is overlain by the upper Eocene-lower Oligocene (Refugian-Zemorrian) Carmannah Formation (Fairchild and Cowan, 1982). The Cenozoic time scale of Berggren and others (1985) brackets the late Eocene between 40.0 and 37 to 36 m.y.; thus San Juan fault probably is older than 36 m.y.

Southern Whidbey Island Fault

The history of the Southern Whidbey Island fault is obscure except that it cuts Quaternary sediments. In the model discussed below, it is inferred to be an extension of the San Juan/Survey Mountain fault.

Structural Model

Devils Mountain/Haro Fault

Because the Haro fault does not occur on Vancouver Island, it probably curves sharply southward along the International Border; thus, the Haro fault may mirror the Devils Mountain fault, which curves sharply southward south of the North Fork of the Stillaguamish River (Figs. 2 and 5). Presumably, the Haro fault continues southward until truncated by the Northern Whidbey Island fault or the San Juan/Survey Mountain fault. Alternatively, the Haro fault is truncated between San Juan Island and Vancouver Island by a north-trending fault (for which no other evidence exists).

The Devils Mountain fault/Haro fault is inferred, therefore, to be a regional, spoon-shaped fault. The upper plate is composed of the synformal Mesozoic terranes of the San Juan Islands and the mainland that were later cut by the Mount Vernon fault. The youngest rocks in the lower plate are the unmetamorphosed Nanaimo and Chuckanut strata. Exotic rocks within the fault are the meta-igneous rocks and the Haro and Spieden strata. Because the lowest upper plate rocks are metamorphosed, the unmetamorphosed Haro and Spieden rocks probably were derived from the lower plate (and could be many kilometers out of place).

The areal extent of the upper plate of the Devils Mountain fault/Haro fault may be large. Vance and others (1980) mapped the probable continuation of the Devils Mountain fault southward to the 34-m.y.-old Index batholith (Fig. 2). Vance and others (1980, Fig. 3) show that in this area the contact is the west-dipping Big Four thrust, with rocks of the Olo Mountain unit (part of PMto of Fig. 2) over the Chuckanut-equivalent Swauk Formation. Danner (1977) recognized rocks of the Olo Mountain unit near Snoqualmie Pass, another 34 km to the south of the Devils Mountain/Haro fault shown in Figure 5. Danner's observation appears to be confirmed by aligning the maps of Vance and others (1980), Tabor and others (1982), and Frizzell and others (1984) from north to south: Rocks apparently equivalent to JKs and PMto on Figure 2 can be traced as far south as about lat. 47°27' N., where they pass beneath Tertiary volcanic rocks. In the southern Cascade Range of Washington, the presence of Jurassic mafic intrusive rocks (references cited by Whetten and others, 1980b) and other rocks that may be similar to those labeled JKs and PMto in Figure 2 (Miller, 1982) suggests that the upper plate might continue far to the south beneath the Tertiary volcanic rocks.

This concept of a major thrust plate is a scaled-up version of Vance's earlier suggestion (1977, p. 188) that the Haro fault is a major southerly dipping thrust that carried the relatively high pressure metamorphic rocks (containing aragonite and lawsonite) of Orcas Island over the unmetamorphosed rocks north of the fault. Vance and others (1980) and Whetten and others (1980b) regarded the ophiolitic rocks of the San Juans as being part of an even larger thrust plate containing many of the ophiolitic rocks of the Cascade Range, including the Ingalls peridotite east of the Straight Creek fault. The essential difference between their hypothesis and the present one is that they regarded the thrusting as mid-Cretaceous (and in the San Juan Islands to be at the base of the Decatur terrane marked JKs in Fig. 2). Brandon and others (1983, p. 25) also recognized that the Haro fault is a

young, post-metamorphic fault that is at least partially responsible for the large uplift necessary to bring the terranes of the San Juan Islands to the surface, but these authors did not speculate how this uplift was accomplished.

Southern Whidbey Island Fault

The identity of the Southern Whidbey Island fault can be inferred from regional patterns. On the southern end of Vancouver Island the middle Eocene Metchosin basalt of the Olympic terrane occurs south of the Leech River fault, whereas the Leech River metamorphic complex is south of the San Juan/Survey Mountain fault (Figure 5). A pronounced magnetic high (recorded on ship-borne equipment) curves southeastward beneath the Strait of Juan de Fuca from the eastern end of the Metchosin basalts on Vancouver Island. MacLeod and others (1977) and Gower (1980) considered the abrupt east boundary of this magnetic high to be the southeastern continuation of the Leech River fault (and it is shown as such on Fig. 5). Because this magnetic high occurs as much as 33 km southwest of the Southern Whidbey Island fault, the Southern Whidbey Island fault is more likely to be the San Juan/Survey Mountain fault than the Leech River fault. MacLeod and others (1977) also reasoned that basement in the Strait of Juan de Fuca northeast of the magnetic high probably is Leech River metamorphic complex; so the area between the magnetic high and the Southern Whidbey Island fault most likely is this complex or related rocks.

Leech River and San Juan Faults

By assuming that the Leech River metamorphic complex and the Metchosin basalt were emplaced from the west, Fairchild and Cowan (1982) inferred more than 40 km of sinistral displacement on both the San Juan fault and the Leech River fault, respectively. However, Fairchild and Cowan (1982) and Brandon and others (1983) did note that the Leech River metamorphic complex resembles the Constitution Formation in the western terranes of the San Juan Islands. Thus, Johnson (1984b) suggested that the San Juan/Survey Mountain fault is a major dextral fault that plucked the Leech River complex from the margin of the continent to the southeast. If the Southern Whidbey Island fault is part of the San Juan/Survey Mountain fault, the probability of such large-scale offset is enhanced, because the Southern Whidbey Island fault probably does truncate the upper plate of the Devils Mountain/Haro fault, from which the Leech River complex could have been derived.

Figure 5 provides a hint of the amount of any such dextral movement on the San Juan/Survey Mountain/Southern Whidbey Island fault. The westernmost extent of the San Juan fault on Vancouver Island is about 100 km west of the Devils Mountain/Haro fault. Because no tectonic discontinuity equivalent to the Devils Mountain/Haro fault has been recognized on Vancouver Island south of the San Juan/Survey Mountain fault, dextral movement on the San Juan/Survey Mountain fault probably is more than 100 km. However, two other possibilities do exist: (1) Perhaps the Leech River complex is equivalent not to the Constitution Formation but to coeval rocks of Table Mountain in the footwall of the Devils Mountain/Haro fault; then dextral displacement of less than 20 km on the San Juan/Survey Mountain/Southern Whidbey Island fault (with the southwest side up) would fail to expose the Devils Mountain/Haro fault and its hanging wall rocks on Vancouver Island. (2) If the southwestern side of the San Juan/Survey Mountain/Southern Whidbey Island fault is greatly uplifted, the entire upper plate of the Devils Mountain/Haro fault could have been completely eroded from Vancouver Island, and the amount of offset would be indeterminable.

Johnson (1984a) and Beck (1984) have speculated that a major strike-slip fault exists inboard of the Olympic terrane. Johnson regarded this as the San Juan/Survey Mountain fault (the San Juan/Survey Mountain/Southern Whidbey Island fault of this paper), but it probably is the Leech River fault. As Johnson noted, the San Juan/Survey Mountain fault juxtaposes continental rocks (that is, previously assembled rocks) against continental rocks; the Leech River fault juxtaposes rocks of the Olympic terrane against the continental rocks. Alternatively, perhaps the San Juan/Survey Mountain/Southern Whidbey Island fault and the Leech River fault are opposite sides of the same major fault zone, in which the Leech River fault may have been active slightly longer.

Mount Vernon Fault

The Mount Vernon fault is but one of several major, high-angle, northwest- to north-trending faults in western Washington and adjacent British Columbia (Fig. 1). The Fraser-Straight Creek fault has a similar history of early dextral strike-slip and later dip-slip reactivation. Dextral displacement, perhaps as much as 200 km, was first recognized by Misch (1977) on the basis of offsets of Mesozoic rocks. Several lines of evidence reviewed by Tabor and others (1984, p. 41) suggest some dextral displacement of Chuckanut-equivalent rocks. The middle Eocene Naches Forma-

tion is disturbed, but not dextrally displaced, along the fault (Vance and Miller, 1981; Tabor and others, 1984). The Naches Formation is equivalent in age to the volcanic rocks of Mount Persis (Tabor and others, 1982), which appear to overlie the Mount Vernon fault. Dip slip on the Straight Creek fault does not cut plutons as old as 33 to 35 m.y. (Vance and Miller, 1981).

The relation of the Mount Vernon fault to either the Leech River fault or the San Juan/Survey Mountain/Southern Whidbey Island fault is uncertain. The Mount Vernon fault with its 47 km of dextral displacement might be satellitic to one of these larger faults. Even if these three faults are not coeval, they probably formed in response to similar stresses.

Late Eocene to Oligocene Cover Rocks

The broad contemporaneity of the major faults is shown by the fact that they are overlain by rocks of roughly the same age. The San Juan fault is overlain by upper Eocene-lower Oligocene strata (Fairchild and Cowan, 1982). The Devils Mountain fault does cut rocks of the same age (Th in Fig. 2); because rocks of the same age occur in the Whatcom basin, these strata originally may have been deposited across the Devils Mountain fault and then cut by reactivation of the fault. The slightly older rocks of Mount Persis overlie the Mount Vernon fault. The abrupt change in thickness of Quaternary sediments south of the Northern Whidbey Island fault and northwest of the Southern Whidbey Island fault may indicate that some major movement on these faults was much later.

Post-Eocene Deformation

Beck and Engebretson (1982) noted that most paleomagnetic directions obtained from rocks in southwestern Washington and western Oregon are northeasterly but that at the three sites in middle Eocene rocks plotted on Figure 5 the directions are northerly to northwesterly. This they attributed to 65° of westward oroclinal bending. They and Beck (1984) preferred to lump the results from the Port Townsend and Bremerton sites, the mean declination of which is 355.5°; thus the bending from the mean of the Bremerton-Port Townsend area to the 330° mean of the Sooke Formation (gabbro) is 25.5°.

The northwest parts of the Mount Vernon fault, San Juan/Survey Mountain/Southern Whidbey Island fault, and Leech River fault have westward bends (Fig. 5). The Mount Vernon fault and the San Juan/Survey Mountain/Southern Whidbey Island fault each bend about 28°. The bend on the Leech River fault

varies from 48° if measured near Port Townsend (where its trace is least well known) to 12° if measured near Victoria. I suggest that the westward bending of the faults is independent evidence for oroclinal bending, which, evidently, was not restricted to the Olympic terrane. Bending of Mount Vernon fault where it cuts Miocene rocks shows that the bending is not Eocene as implied by Beck and Engebretson (1982).

Three, and possibly four, large folds have affected the Tertiary rocks of northwestern Washington and adjacent British Columbia. The size, structural relief, and trend of the synformal Whatcom basin (as shown by Hopkins, 1968, fig. 1) are about the same as those of the antiform that underlies the Olympic Mountains; both of these folds involve Miocene strata. Belts of Chuckanut Formation may define an antiformal axis north of the Skagit River. The Clallam syncline (MacLeod and others, 1977, fig. 2) borders the antiform of the Olympic Mountains on the north. Brecciated rocks within the core of the Olympic Mountains and the unconformity below the upper Miocene Montesano Formation on the southern flank of the Olympic Mountains suggest that most of the Olympic antiform formed from 17 to 12 m.y. ago (Tabor, 1972). These folds probably have locally modified the dip (and possibly the strike) of the older Mount Vernon fault, Leech River fault, and San Juan/Survey Mountain fault.

If the Southern Whidbey Island fault is the southeastern continuation of the San Juan/Survey Mountain fault, the nature of the Northern Whidbey Island fault, which strikes into either the Leech River fault or the San Juan/Survey Mountain fault, becomes even more enigmatic. The Northern Whidbey Island fault is nearly tangent to the greatest curvature of the Leech River fault and the San Juan/Survey Mountain/Southern Whidbey Island fault. Perhaps tectonic stresses caused reactivation of one of the faults to be propagated eastward as the Northern Whidbey Island fault, forming a graben bounded on three sides by the Southern Whidbey Island fault, Northern Whidbey Island fault, and Mount Vernon fault.

Holocene Deformation and Seismicity

According to Thorson (1981), (1) the amount of isostatic rebound during the last 13,000 yr varied from 140 m in the southern part of the area of Figure 2 to more than 350 m in the Fraser Lowland of British Columbia, and (2) the belt of shallow earthquakes in Washington is largely coincident with the former area of the Puget ice lobe. However, few shallow earthquakes occur north of 49° N. (Rogers, 1983) where the ice was thicker and isostatic rebound greater than

to the south. Similarly, the western side of the Puget Lowland has far fewer shallow earthquakes than the eastern side. Additionally, Rogers (1983) has reaffirmed an earlier observation of Crosson (1972) and Riddihough and Hyndman (1976) that the orientation of pressure axes in the focal mechanism solutions of the shallow earthquakes is compatible with north-south compression or dextral shear in the Puget Lowland.

Much of the faulting of Pleistocene and Holocene sediments may have been caused by reactivation of old faults during rapid isostatic rebound, as Thorson suggested, but perhaps the present seismicity is truly tectonic. The diffuse pattern of the epicenters may be caused by various faults responding to the present north-south compression. The shallow 1946 earthquake near Campbell River on Vancouver Island (Fig. 1) was large ($M = 7.1/4 \pm 1/4$) and is best explained by dextral movement on a northwest-striking strike slip fault (Rogers and Hasegawa, 1978; Slawson and Savage, 1979), as would be expected during the present north-south compression. If a similar shallow earthquake were to occur in the northern Puget Lowland, it would likely be much more destructive than the better known but deeper (30-70 km) earthquakes of lesser magnitude that are associated with the subducted oceanic crust beneath the western part of the Puget Lowland.

CONCLUSIONS

The primary conclusion of this paper is that major structural features in the northern Puget Lowland are much younger than is commonly supposed. Dobrin (1975) was the first to describe the Northern Whidbey Island fault, which may have significant Quaternary displacement. Likewise, he was the first to demonstrate 2 km of vertical displacement of Miocene strata in the Whatcom basin along what is here considered to be the Mount Vernon fault. In fact, the presence of 1 km or more of terrestrial Miocene strata in the Whatcom basin, although described by Hopkins (1966, 1968), is not widely known today. Lovseth (1975) rediscovered the Devils Mountain fault. Fairchild and Cowan (1982) and Johnson (1984b) showed that pre-late Eocene transcurrent faulting is important, and this paper describes 47 km of pre-late Eocene dextral displacement on the Mount Vernon fault. Gower (1980) described the Southern Whidbey Island fault, which may be the southeastern extension of the pre-late Eocene San Juan/Survey Mountain fault of Vancouver Island. Beck and Engebretson (1982) inferred the oroclinal bending of the Olympic terrane.

The major pre-late Eocene structure is the huge thrust bounded by the Devils Mountain/Haro fault, which replaced the terranes of the San Juan Islands

and foothills of the Cascades over unmetamorphosed Chuckanut and Nanaimo Formations (and associated metamorphic rocks). This plate extends from the San Juan Islands (lat. 48.7°N.) southward beyond Snoqualmie Pass at lat. 47.3°N.

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REFERENCES CITED

- Bechtel, Inc., 1979, Report of geologic investigations in 1978-79 Skagit Nuclear Power Project: 3 v. [Available from Puget Sound Power and Light Company, Bellevue, Washington].
- Beck, M. E., Jr., 1984, Has the Washington-Oregon Coast Range moved northward?: *Geology*, v. 12, no. 12, p. 737-740.
- Beck, M. E., Jr.; Engebretson, D. C., 1982, Paleomagnetism of small basalt exposures in the west Puget Sound area, Washington, and speculations on the accretionary origin of the Olympic Mountains: *Journal of Geophysical Research*, v. 87, no. B5, p. 3755-3760.
- Berggren, W. A.; Kent, D. V.; Flynn, J. J.; Van Couvering, J. A., 1985, Cenozoic geochronology: *Geological Society of America Bulletin*, v. 96, no. 11, p. 1407-1418.
- Brandon, M. T.; Cowan, D. S., 1985, The Late Cretaceous San Juan Islands—Northwestern Cascades thrust system [abstract]: *Geological Society of America Abstracts with Programs*, v. 17, no. 6, p. 343.
- Brandon, M. T.; Cowan, D. S.; Vance, J. A., 1983, The geology of the San Juan Islands, Washington: *Geological Association Canada Field Trip Guidebook of Joint Annual Meeting in Victoria, Field Trip No. 5*, p. 4-29.
- Brown, E. H.; Bernardi, M. L.; Christenson, B. W.; Cruver, J. R.; Haugerud, R. A.; Rady, P. M.; Sondergaard, J. N., 1981, Metamorphic facies and tectonics in part of the Cascade Range and Puget Lowland of northwestern Washington: *Geological Society of America Bulletin*, Part 1, v. 92, no. 4, p. 170-178.
- Carroll, P. E., 1980, Petrology and structure of the pre-Tertiary rocks of Lummi and Eliza Islands, Washington: University of Washington Master of Science thesis, 78 p.
- Cheney, E. S., 1976, Newly discovered major Cenozoic faults in northwestern Washington [abstract]: *Geological Society of America Abstracts with Programs*, v. 8, no. 6, p. 810-811.
- Cheney, E. S., 1977, Preliminary Cenozoic tectonic map of northwestern Washington and southwestern British Columbia [abstract]: *Geological Society of America Abstracts with Programs*, v. 9, no. 7, p. 926-927.
- Cheney, E. S., 1979, Interim report of the seismic and geologic hazards to the proposed Skagit nuclear power site, Sedro Woolley, Washington (7th revision); prefilled testimony submitted October 1979 to the Atomic Safety and Licensing Board: U.S. Nuclear Regulatory Commission Docket Nos. STN 50-522 and STN 50-523, 134 p.
- Cowan, D. S., 1982, Geological evidence for post-40 m.y.B.P. large-scale northwestward displacement of part of southeastern Alaska: *Geology*, v. 10, p. 310-313.
- Cowan, D. S.; Whetten, J. T.; Brown, E. H., 1977, Geology of the southern San Juan Islands. In Brown, E. H.; Ellis, R. C., editors, 1977, *Geological Excursions in the Pacific Northwest*; Geological Society of America (Guidebook), 1977: Western Washington University, p. 309-338.
- Crosson, R. W., 1972, Small earthquakes, structure, and tectonics of the Puget Sound region: *Seismological Society of America Bulletin*, v. 62, no. 5, p. 1133-1171.
- Crosson, R. S., 1983, Review of seismicity in the Puget Sound region from 1970 through 1978. In Yount, J. C.; Crosson, R. S., editors, 1983, *Proceedings of Workshop XIV, Earthquake hazards of the Puget Sound region*, Washington: U.S. Geological Survey Open-File Report 83-19, p. 6-18.
- Dalrymple, G. B., 1979, Critical tables for conversion of K-Ar ages from old to new constants: *Geology*, v. 7, no. 12, p. 558-560.
- Danner, W. R., 1957, A stratigraphic reconnaissance in the northwestern Cascade Mountains and San Juan Islands of Washington State: University of Washington Doctor of Philosophy thesis, 562 p.
- Danner, W. R., 1966, Limestone resources of western Washington: Washington Division of Mines and Geology Bulletin 52, 474 p.
- Danner, W. R., 1977, Paleozoic rocks of northwest Washington and adjacent parts of British Columbia. In Stewart, J. H.; Stevens, C. N.; Fritsche, A. E., editors, 1977, *Paleozoic paleogeography of the western United States*; Pacific Coast Paleogeography Symposium 1: Society of Economic Paleontologists and Mineralogists, Pacific Section, p. 481-502.
- Dethier, D. P.; Whetten, J. T., 1980, Preliminary geologic map of the Clear Lake SW quadrangle, Skagit and Snohomish Counties, Washington: U.S. Geological Survey Open-File Report 80-825, 10 p., 2 pl., scale 1:24,000.
- Dethier, D. P.; Whetten, J. T.; Carroll, P. R., 1980, Preliminary geologic map of the Clear Lake SE quadrangle, Skagit County, Washington: U.S. Geological Survey Open-File Report 80-393, 11 p., 2 pl., scale 1:24,000.
- Dobrin, M. B., 1975 to 1977, In Consultants' letters, Appendix 2G, Preliminary Safety Analysis Report for the Skagit Nuclear Power Project, Units 1 and 2, Atomic Safety and Licensing Board, U.S. Nuclear Regulatory Commission Docket Nos. STN 50-522 and 50-523, issued by Puget Sound Power and Light Company, Bellevue, Washington.
- Engels, J. C.; Tabor, R. W.; Miller, F. K.; Obradovich, J. D., 1976, Summary of K-Ar, Rb-Sr, U-Pb, Pb, and fission-track ages of rocks from Washington State prior to 1975 (exclusive of Columbia Plateau basalts): U.S. Geological Survey Miscellaneous Field Studies Map MF-710, 2 sheets, scale 1:1,000,000.
- Fairchild, L. M.; Cowan, D. S., 1982, Structure, petrology, and tectonic history of the Leech River complex northwest of Victoria, Vancouver Island: *Canadian Journal of Earth Sciences*, v. 19, p. 1817-1835.
- Frizzell, V. A., Jr.; Tabor, R. W.; Booth, D. B.; Ort, K. M.; Waitt, R. B., Jr., 1984, Preliminary geologic map of the Snoqualmie Pass 1:100,000 quadrangle, Washington: U.S. Geological Survey Open-File Report 84-693, 42 p., 1 pl., scale 1:100,000.
- Gower, H. D., 1980, Bedrock geologic and Quaternary tectonic map of the Port Townsend area, Washington: U.S. Geological Survey Open-File Report 80-1174, 9 p., 1 pl., scale 1:100,000.
- Hansen, B. S.; Easterbrook, D. J., 1974, Stratigraphy and palynology of late Quaternary sediments in the Puget Lowland, Washington: *Geological Society of America Bulletin*, v. 85, no. 4, p. 587-602.
- Hobbs, S. W.; Pecora, W. T., 1941, Nickel-gold deposit near Mount Vernon, Skagit County, Washington: U.S. Geological Survey Bulletin 931-D, p. 57-78.
- Hopkins, W. S., Jr., 1966, Palynology of Tertiary rocks of the Whatcom basin, southwestern British Columbia and northwestern Washington: University of British Columbia Doctor of Philosophy thesis, 184 p., 4 pl.

- Hopkins, W. S., Jr., 1968, Subsurface Miocene rocks, British Columbia-Washington—A palynological investigation: Geological Society of America Bulletin, v. 79, no. 6, p. 763-768.
- Jackson, E. J., compiler, 1976, Generalized geological map of the Canadian Cordillera. In Sutherland-Brown, Atholl, editor, 1976, Porphyry deposits of the Canadian Cordillera: Canadian Institute of Mining and Metallurgy Special Volume 15, Map A, scale 1:2,500,000.
- Jenkins, O. P., 1924, Geological investigation of the coal fields of Skagit County, Washington: Washington Division of Geology Bulletin 29, 63 p., 7 pl.
- Johnson, S. Y., 1978, Sedimentology, petrology, and structure of Mesozoic strata in the northwestern San Juan Islands, Washington: University of Washington Master of Science thesis, 105p., 4 pl.
- Johnson, S. Y., 1981, The Spieden Group—An anomalous piece of the Cordilleran paleogeographic puzzle: Canadian Journal of Earth Sciences, v. 18, no. 11, p. 1694-1707.
- Johnson, S. Y., 1982, Stratigraphy, sedimentology, and tectonic setting of the Eocene Chuckanut Formation, northwest Washington: University of Washington Doctor of Philosophy thesis, 221 p.
- Johnson, S. Y., 1984a, Stratigraphy, age, and paleogeography of the Eocene Chuckanut Formation, northwest Washington: Canadian Journal of Earth Sciences, v. 21, no. 1, p. 92-106.
- Johnson, S. Y., 1984b, Evidence for a margin-truncating transcurrent fault (pre-late Eocene) in western Washington: Geology, v. 12, no. 9, p. 538-541.
- Jones, R. W., 1959, Geology of the Finney Peak area, northern Cascades of Washington: University of Washington Doctor of Philosophy thesis, 186 p.
- Lawrence, R. D.; Rosenfeld, C. L., 1978, Tectonic analysis of SLAR mosaic of northwest Cascade Range and adjacent lowlands, Washington [abstract]: EOS (American Geophysical Union Transactions), v. 59, no. 4, p. 233.
- Lovseth, T. P., 1975, The Devils Mountain fault zone, northwestern Washington: University of Washington Master of Science thesis, 29 p.
- MacLeod, N. S.; Tiffin, D. L.; Snively, P. D., Jr.; Currie, R. G., 1977, Geologic interpretation of magnetic and gravity anomalies in the Strait of Juan de Fuca, U.S.-Canada: Canadian Journal of Earth Sciences, v. 14, no. 2, p. 223-238.
- Marcus, K. L., 1981, The rocks of Bulson Creek—Eocene-Oligocene sedimentation and tectonics in the Lake McMurray area, Washington: Western Washington University Master of Science thesis, 84 p., 1 pl.
- Miller, G. M.; Misch, Peter, 1963, Early Eocene angular unconformity at western front of Northern Cascades, Whatcom County, Washington: American Association of Petroleum Geologists Bulletin, v. 47, p. 163-174.
- Miller, R. B., 1982, Geology of the Rimrock Lake pre-Tertiary inlier, southern Washington Cascades [abstract]: Geological Society of America Abstracts with Programs, v. 14, no. 4, p. 217.
- Minard, J. P., 1980a, Distribution and description of the geologic units in the Arlington East quadrangle, Washington: U.S. Geological Survey Open-File Report 80-460, 8 p., 1 pl., scale 1:24,000.
- Minard, J. P., 1980b, Distribution and description of the geologic units in the Arlington West quadrangle, Washington: U.S. Geological Survey Open-File Report 80-461, 8 p., 1 pl., scale 1:24,000.
- Minard, J. P., 1980c, Distribution and description of the geological units in the Lake Stevens quadrangle, Washington: U.S. Geological Survey Open-File Report 80-463, 7 p., 1 pl., scale 1:24,000.
- Minard, J. P., 1980d, Distribution and description of the geologic units in the Snohomish quadrangle, Washington: U.S. Geological Survey Open-File Report 80-2013, 8 p., 1 pl., scale 1:24,000.
- Minard, J. P., 1981a, Distribution and description of the geologic units in the Maltby quadrangle, Washington: U.S. Geological Survey Open-File Report 81-100, 5 p., 1 pl., scale 1:24,000.
- Minard, J. P., 1981b, Distribution and description of the geologic units in the Bothell quadrangle, Washington: U.S. Geological Survey Open-File Report 81-106, 5 p., 1 pl., scale 1:24,000.
- Minard, J. P., 1981c, Distribution and description of the geologic units in the Everett quadrangle, Washington: U.S. Geological Survey Open-File Report 81-248, 6 p., 1 pl., scale 1:24,000.
- Misch, Peter, 1977, Dextral displacements at some major strike faults in the North Cascades [abstract]: Geological Association of Canada Programme with Abstracts, v. 2, p. 37.
- Noson, L. L.; Crosson, R. S., 1980, Compilation of earthquake hypocenters in western Washington, 1978: Washington Division of Geology and Earth Resources Information Circular 72, 18 p.
- Pacht, J. A., 1980, Sedimentology and petrology of the Late Cretaceous Nanaimo Group in the Nanaimo Basin, Washington and southern British Columbia—Implications for Late Cretaceous tectonics: Ohio State University, Doctor of Philosophy thesis, 380 p.
- Palmer, Pamela, 1977, Investigation of tectonic deformation in the Puget Lowland, Washington (final technical report): Washington Division of Geology and Earth Resources Open File Report 77-6, 36 p., 3 pl., scale 1:24,000.
- Price, R. A.; Monger, J. W. H.; Muller, J. E., 1981, Cordilleran cross-section—Calgary to Victoria: Geological Association of Canada 1981 Annual Meeting, Field Guides to Geology and Mineral Deposits, p. 261-334.
- Riddihough, R. P., and Hyndman, R. D., 1976, Canada's active western margin—The case for subduction: Geoscience Canada, v. 3, no. 4, p. 269-278.
- Roddick, J. A., Muller, J. E., and Okulitch, A. V., 1979, Fraser River, British Columbia—Washington: Geological Survey of Canada Map 1386A, 3 sheets, scale 1:1,000,000.
- Rogers, G. C., 1983, Some comments on the seismicity of the northern Puget Sound—southern Vancouver Island region. In Yount, J. C.; Crosson, R. S., editors, 1983, Proceedings of workshop XIV, Earthquake hazards of the Puget Sound region, Washington: U.S. Geological Survey Open-File Report 83-19, p. 19-39.
- Rogers, G. C.; Hasegawa, H. S., 1978, A second look at the British Columbia earthquake of June 23, 1946: Seismological Society of America Bulletin, v. 68, no. 3, p. 653-675.
- Rogers, W. P., 1970, A geological and geophysical study of the central Puget Lowland: University of Washington Doctor of Philosophy thesis, 123 p., 9 pl.
- Schriener, Alexander, Jr., 1976, Gravity study of the southern Skagit River delta, Washington: University of Washington Bachelor of Science thesis, 26 p.
- Seraphim, R. H.; Hollister, V. F.; Jackson, E. V., 1976, Faults, porphyry deposits and showings, and tectonic belts of the Canadian Cordillera. In Sutherland-Brown, Atholl, editor, 1976, Porphyry deposits of the Canadian Cordillera: Canadian Institute of Mining and Metallurgy Special Volume 15, Map B, scale 1:2,500,000.
- Slawson, W. F.; Savage, J. C., 1979, Geodetic deformation associated with 1946 Vancouver Island, Canada earthquake: Seismological Society of America Bulletin, v. 69, no. 5, p. 1487-1496.
- Tabor, R. W., 1972, Age of the Olympic metamorphism, Washington—K-Ar dating of low-grade metamorphic rocks: Geological Society of America Bulletin, v. 83, no. 6, p. 1805-1816.

- Tabor, R. W.; Cady, W. M.; and others, 1978, Geologic Map of the Olympic Peninsula, Washington: U.S. Geological Survey Map I-994, 2 sheets, scale 1:125,000.
- Tabor, R. W.; Frizzell, V. A., Jr.; Booth, D. B.; Whetten, J. T.; Waitt, R. B., Jr.; Zartman, R. E., 1982, Preliminary geologic map of the Skykomish River 1:100,000 quadrangle, Washington: U.S. Geological Survey Open-File Map OF 82-747, 25 p., scale 1:100,000.
- Tabor, R. W.; Frizzell, V. A., Jr.; Vance, J. A.; Naeser, C. W., 1984, Ages and stratigraphy of lower and middle Tertiary sedimentary and volcanic rocks of the central Cascades, Washington—Application to the tectonic history of the Straight Creek fault: *Geological Society of America Bulletin*, v. 95, p. 26-44.
- Thorson, R. M., 1981, Isostatic effects of the last glaciation in the Puget Lowland, Washington: U.S. Geological Survey Open-File Report 81-370, 100 p., 1 pl.
- Tipper, H. W.; Woodsworth, G. J.; Gabrielse, H., 1981, Tectonic assemblage map of the Canadian Cordillera and adjacent parts of the United States of America: Geological Survey of Canada Map 1505A, 2 sheets, scale 1:2,000,000.
- Vance, J. A., 1975, Bedrock geology of San Juan County: Washington Department of Ecology Water Supply Bulletin 46, p. 3-20.
- Vance, J. A., 1977, The stratigraphy and structure of Orcas Island, San Juan Islands. In Brown, E. H.; Ellis, R. C., editors, 1977, *Geological excursions in the Pacific Northwest*; Geological Society of America (Guidebook), 1977: Western Washington University, p. 170-203.
- Vance, J. A.; Dungan, M. A.; Blanchard, D. P.; Rhodes, J. M., 1980, Tectonic setting and trace element geochemistry of Mesozoic ophiolitic rocks in western Washington: *American Journal of Science*, v. 280-A, Pt. 1, p. 359-388.
- Vance, J. A.; Miller, R. B., 1981, The movement history of the Straight Creek fault in Washington State [abstract]: Geological Association of Canada, Cordilleran Section, Programme and Abstracts, p. 39-41.
- Wagner, H. C.; Wiley, M. C., 1983, Offshore Quaternary geology of the northern Puget Sound—eastern Strait of Juan de Fuca region, Washington. In Yount, J. C.; Crosson, R. S., editors, 1983, *Proceedings of workshop XIV, Earthquake hazards of the Puget Sound region, Washington*: U.S. Geological Survey Open-File Report 83-19, p. 178-267.
- Walsh, T. J., 1984, Coal found on Whidbey Island: *Washington Geologic Newsletter*, v. 12, no. 1, p. 15-17.
- Ward, P. D., 1978, Revisions to the stratigraphy and biochronology of the Upper Cretaceous Nanaimo Group, British Columbia and Washington State: *Canadian Journal of Earth Sciences*, v. 15, no. 3, p. 405-423.
- Whetten, J. T., 1978, The Devil's Mountain fault—A major Tertiary structure in northwest Washington [abstract]: *Geological Society of America Abstracts with Programs*, v. 10, no. 3, p. 153.
- Whetten, J. T.; Jones, D. L.; Cowan, D. S.; Zartman, R. E., 1978, Ages of Mesozoic terranes in the San Juan Islands, Washington. In Howell, D. G.; McDougall, K. A., editors, 1978, *Mesozoic paleogeography of the western United States*, Pacific Coast Paleogeography Symposium no. 2: Society of Economic Paleontologists and Mineralogists, Pacific Section, p. 117-132.
- Whetten, J. T.; Dethier, D. P.; Carroll, P. R., 1979, Preliminary geologic map of the Clear Lake NE quadrangle, Skagit County, Washington: U.S. Geological Survey Open-File Report 79-1468, 10 p., 2 pl., scale 1:24,000.
- Whetten, J. T.; Dethier, D. P.; Carroll, P. R., 1980a, Preliminary geologic map of the Clear Lake NW quadrangle, Skagit County, Washington: U.S. Geological Survey Open-File Report 80-247, 12 p., 2 pl., scale 1:24,000.
- Whetten, J. T.; Zartman, R. E.; Blakely, R. J.; Jones, D. L., 1980b, Allochthonous Jurassic ophiolite in northwest Washington: *Geological Society of America Bulletin*, Part 1, v. 91, no. 6, p. 359-368.
- Wiebe, R. A., 1963, The geology of Mount Pilchuck: University of Washington Master of Science thesis, 52 p.
- Yount, J. C.; Crosson, R. S., editors, 1983, *Proceedings of workshop XIV, Earthquake hazards of the Puget Sound region, Washington*: U.S. Geological Survey Open-File Report 83-19, 306 p.

EARLY CENOZOIC GEOLOGY OF CENTRAL WASHINGTON STATE: I. SUMMARY OF SEDIMENTARY, IGNEOUS, AND TECTONIC EVENTS¹

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ABSTRACT

New field mapping and radiometric dating have made possible the construction of a detailed geologic history for early Cenozoic rocks of central Washington. The lower Eocene Swauk Formation was apparently a widespread unit consisting primarily of fluvial arkosic sandstone; it was folded about east-west axes prior to extrusion of the middle Eocene Teanaway Basalt. Northeast-trending Teanaway dikes are coeval with the northwest-trending bimodal Corbaley Canyon dike swarm located only 50 kilometers away. The northwest-trending Chiwaukum graben was active over about a 6-million-year span during middle Eocene time, and at least 5,800 meters of Chumstick Formation accumulated in it. The Chumstick is dominantly fluvial arkosic sandstone, with interbedded tuff, but the upper part is a shaly lacustrine facies. The region was tectonically and magmatically quiet during the late Eocene and early Oligocene and was beveled to a surface of low relief mantled with deeply weathered bedrock. Two hundred and eighty meters of Wenatchee Formation, a dominantly fluvial unit of quartzose sandstone and shale with minor tuff beds, was deposited on the erosion surface. Compressional deformation within 5 million years of its deposition involved the Wenatchee and older rocks in minor thrust faulting and folding. Hornblende andesite sills and dikes of the Horse Lake Mountain complex intruded all sedimentary units in the southern part of the graben in late Oligocene time; their emplacement was synchronous with deformation by northwest-trending right-lateral shearing. Bentonitic shales of Miocene age were deposited prior to extrusion of the Columbia River Basalt Group.

INTRODUCTION

The Chiwaukum graben and adjacent areas (Fig. 1) encompass a wealth of geologic relationships that are critical to understanding the geologic history of central Washington during the early Cenozoic. Part I of this report introduces new radiometric dates, summarizes old dates, and reviews the sedimentary, igneous, and tectonic history from Paleocene through Miocene time. Part II discusses the implications for plate tectonics and delineates some of the unsolved tectonic problems. Recent work has redefined the stratigraphy of diverse sedimentary units that were previously included under the term "Swauk Formation" (Gresens and others, 1981). For convenience, the Chiwaukum graben, a major northwest-trending structure, is used as a reference area, and the stratigraphy is divided into three parts: the graben interior and contiguous terranes to the southwest and northeast (Fig. 2).

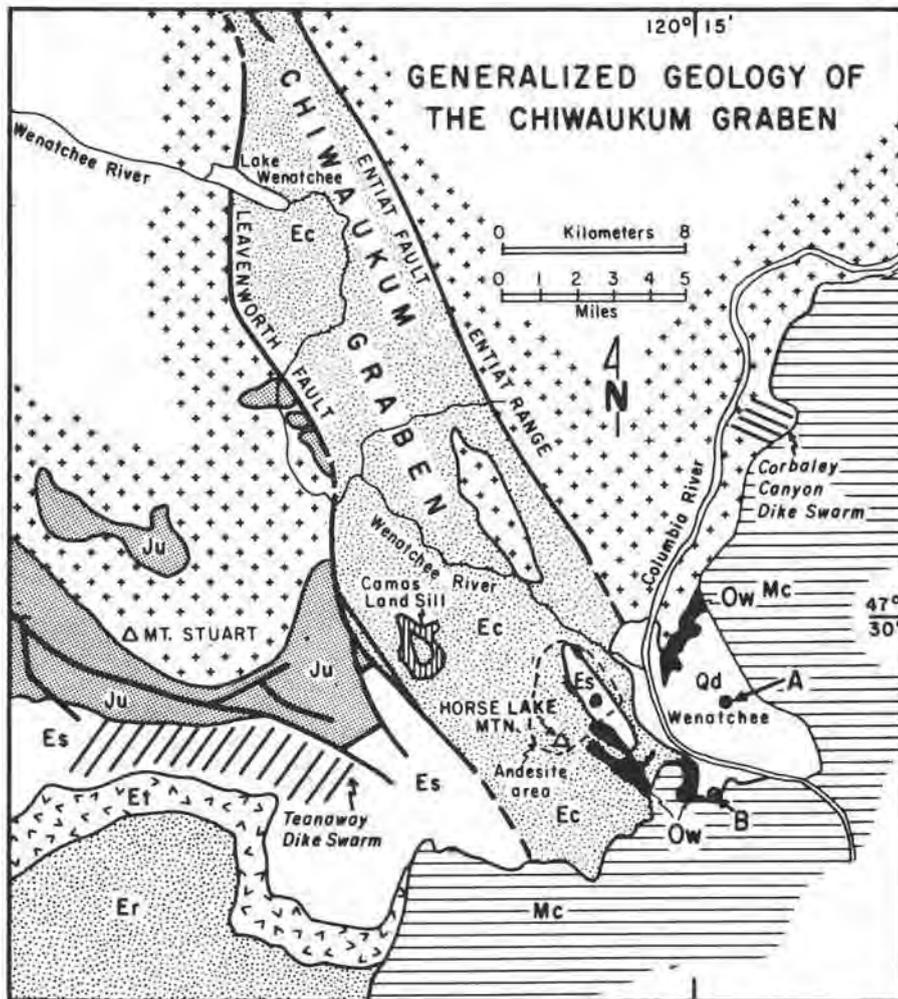
GEOLOGIC HISTORY

The early- to mid-Cenozoic history of the Chiwaukum graben and vicinity is summarized in Table 1. The tripartite division used for the stratigraphic summary is retained. The positions of the Entiat and Leavenworth faults, which bound the graben, respectively, on the east and west, are used to divide the region, even though they did not come into existence until post-Teanaway time.

Eocene Events

South of the Mount Stuart massif, in the area southwest of the graben, the Swauk Formation rests unconformably on crystalline rocks. Its basal section contains ironstones that are interpreted as diagenetically altered laterites developed on underlying serpentine (Lupher, 1944; Lamey and Hotz, 1952). The upper Swauk Formation is reported to interfinger with the 51-m.y.-old Silver Pass volcanics (Tabor and Frizell, 1977). The lower age is not known but could extend into the Paleocene. The Swauk, which consists

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LEGEND

| | | | |
|--|---|--|--|
| | Quaternary--Undifferentiated surficial deposits | | Middle Eocene--Teanaway Basalt |
| | Miocene--Columbia River Basalt | | Lower Eocene--Swauk Fm. |
| | Oligocene--Wenatchee Fm. | | Jurassic--Ultramafic rocks |
| | Middle to Upper Eocene--Roslyn & Chumstick Fms. | | |
| | | | Precambrian(?) to Cretaceous(?) Crystalline basement rocks |

Figure 1.—Generalized geology of the Chiwaukum graben in central Washington. The area outlined as “andesite” (the Horse Lake Mountain complex) is shown in more detail in Figure 5.

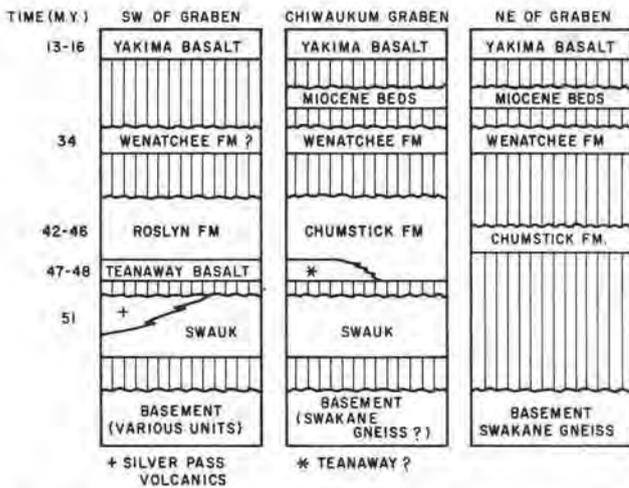


Figure 2.—Stratigraphic columns for terranes within the Chiwaukum graben and on either side of it.

mostly of fluvial and lacustrine arkosic sandstone and shale, was probably a widespread depositional unit and is at least 2,300 m thick near the type locality (Frizzell, 1979). The Swauk is probably correlative with the Chuckanut Formation exposed in the northern Puget Sound region west of the Cascade Range. Both the Swauk and Chuckanut contain abundant fossil palmetto leaves, which are not found in either the Chumstick or Roslyn Formations. Although these fossils are not diagnostic of age, they reflect differences in climate. Frizzell (1979) suggested correlation of Swauk and Chuckanut based on sandstone petrology. Although no Swauk Formation has been observed in the terrane northeast of the graben, it may be absent because of subsequent erosion, rather than because of nondeposition, or it may be offset by the Entiat fault (Gresens, 1982).

The Swauk was folded prior to deposition of Teanaway Basalt; folds have a general east-west trend (Tabor and Frizzell, 1977), indicating north-south compression (Gresens, 1979; Ewing, 1980). This deformation took place between about 51 and 47 m.y.B.P. In contrast, the belt of Swauk Formation within the graben, exposed in the core of Eagle Creek anticline near the city of Wenatchee, has the same northwest structural grain as the trends of the Leavenworth and Entiat faults and the anticline. Evidence of multiple folding, timing unknown, is present in these rocks.

The age of the basalt is about 47 m.y., based on whole-rock K-Ar ages (R. W. Tabor, personal commun., 1981). An extensive swarm of basalt dikes, feeders of Teanaway volcanism, cuts the Swauk Formation in the terrane southwest of the graben (Foster, 1958; Southwick, 1966). The dikes have a general orientation of about N15°E and are nearly vertical

Table 1.—Summary of geologic history of the Chiwaukum graben.

| Time (m.y.) | Leavenworth Fault | | Entiat Fault |
|-------------|---|---|--|
| | Southwest of Graben | Chiwaukum Graben | Northeast of Graben |
| 15-29 | Regional Miocene deposition, including Columbia River Basalts, followed by uplift and erosion, involving ancestral Cascade Range | | |
| 29 | | Intrusion of Horse Lake Mountain Complex—hornblende andesite | |
| 29-34 | | Folding and thrusting of Wenatchee Formation—NE-SW compression | |
| 34 | | Deposition of Wenatchee Fm | |
| 34-40(?) | ← Regional period of tectonic and magmatic quiescence → | | |
| 40(?)–42(?) | | Folding of Chumstick Formation—possible NE-SW compression | |
| 42(?)–46(?) | Deposition of Roslyn Formation conformably on Teanaway Basalt | Deposition of Chumstick Formation in subsiding Chiwaukum graben—NE-SW extension | Active erosion of highland area Local deposition of Chumstick Formation |
| 46(?) | | Initiation of the Chiwaukum graben—NE-SW extension | |
| 47-48 | Emplacement of Teanaway dike swarm in Swauk Formation and extrusion of Teanaway Basalt unconformably over Swauk Formation—WNE-ESE extension | | Emplacement of Corbaley Canyon dike swarm (rhyodacite) into Swakane Gneiss—NNE-SSW extension |
| 47-51 | Folding of Swauk Formation—N-S compression | | |
| 51-?? | Regional deposition of Swauk Formation | | ????? |

(Foster, 1958), which indicates WNW-ESE crustal extension during Teanaway time. Although the composition of the Teanaway is primarily basaltic, minor rhyolite occurs at the top of the unit (Clayton, 1973).

A remarkable dike swarm also is present in the terrane northeast of the graben. The dikes form prominent ridges where they are differentially weathered with respect to the metamorphic rocks (Fig. 3). They are well exposed in road cuts in Corbaley Canyon along U.S. Highway 2 and to the northwest along strike in road cuts on the west side of the Columbia River. The dikes are dated at about 47-48 m.y.B.P.; thus, they are at least broadly coeval with Teanaway volcanism. Corbaley Canyon dikes are bimodal but are dominated by felsic varieties ranging from porphyritic dacite to rhyolite with bipyramidal quartz phenocrysts (Waters, 1926). Mafic rocks include both andesitic varieties and hornblende gabbros. The dikes are clearly pull-apart structures, and one exposure along the highway shows an intrusion of a second dike with chilled borders into an early dike that had been tensionally split apart (Fig. 4). The dikes have a strong N70°W trend, which indicates NNE-SSW crustal extension during their emplacement. Other dike swarms, part of the broad period of Challis-Absaroka volcanism (Armstrong, 1978), are common throughout the region to the north and northwest of the Corbaley Canyon swarm.

Initiation of the Chiwaukum graben as an active structure probably began about 46 m.y.B.P., based on



Figure 3.—View of roadcuts in Corbaley Canyon. The felsic dikes weather differentially relative to the gneiss they intrude, and they show up in striking relief.

the age of the Chumstick Formation. During its subsidence, it accumulated a thick (5,800 m), mainly fluvial and alluvial, sequence of arkosic sandstone and shale (Chumstick Formation) (Whetten, 1977; Gresens, 1981). Toward the end of its activity, sedimentation was dominated by shaly lacustrine facies as reflected in the upper Chumstick Formation (Nahahum Canyon member) (Whetten, 1977). Dating of the Chumstick, and hence of graben activity, is based on a suite of nine fission-track ages from interbedded tuff units that range from 41.9 to 48.8 m.y.B.P. (Whetten, 1977; Gresens and others, 1981). Because of analytical uncertainty in the fission-track method, there is significant variance of the age of the sample population, but the active life of the graben probably spanned the range from about 40 to 46 m.y.B.P.

Along the Leavenworth fault zone in the southern half of the graben, there is an occurrence of chaotic basalt fragments interbedded with the Chumstick Formation. Tabor and Frizzell (1977, Frizzell and Tabor, 1977) have interpreted this as a Teanaway volcanic breccia. They thus have inferred that deposition of the Chumstick and initiation of graben subsidence predate the Teanaway Basalt. The writer has seen the locality and reserves final judgement. However, thick conglomerates with well-rounded clasts, composed nearly exclusively of basalt, occur nearby in blocks of lower Chumstick faulted upward along the Leavenworth fault zone. These must be fanglomerates shed from the Teanaway Formation into the subsiding graben, similar to serpentinite fanglomerates described by Cashman and Whetten (1976) farther north along

the Leavenworth fault. It is possible that the chaotic basalt fragments are associated mass-wasting deposits (perhaps debris flows) shed into the graben from pre-existing Teanaway Formation perhaps a million or more years after its extrusion, rather than volcanic breccia coeval with Teanaway extrusion.

The Chiwaukum graben was flanked by highlands of crystalline basement rock during deposition of Chumstick Formation (Whetten, 1977; Buza, 1979). Apparently the terrane east of the Entiat fault was a persistent highland. The terrane west of the Leavenworth fault was only an intermittent highland, and the Chumstick may have been locally continuous with the Roslyn Formation lying to the west over regions that were subjected to later erosion. The Roslyn Formation has a thickness estimated as 1,900 m (Bressler, 1951). Its equivalent within the graben, the Chumstick Formation, is much thicker; Whetten (1977; Gresens and others, 1981) estimated it conservatively at 5,800 m. Based on the evidence by Buza (1979) and Whetten (1977) and the continuation of Whetten's structural and stratigraphic patterns to the southeast, Gresens (1980) argued that Chumstick Formation lying east of the Entiat fault was emplaced by post-34-m.y.B.P. thrusting out of the graben. Although this interpretation is supported by structural data and remains valid for Chumstick exposures relatively near the Entiat fault, additional fieldwork has revealed poorly exposed Chumstick sufficiently far east of the Entiat fault that it is probably in its original depositional position (Fig. 1). Thus the Entiat highland, although probably tectonically active continually during Chumstick time, may

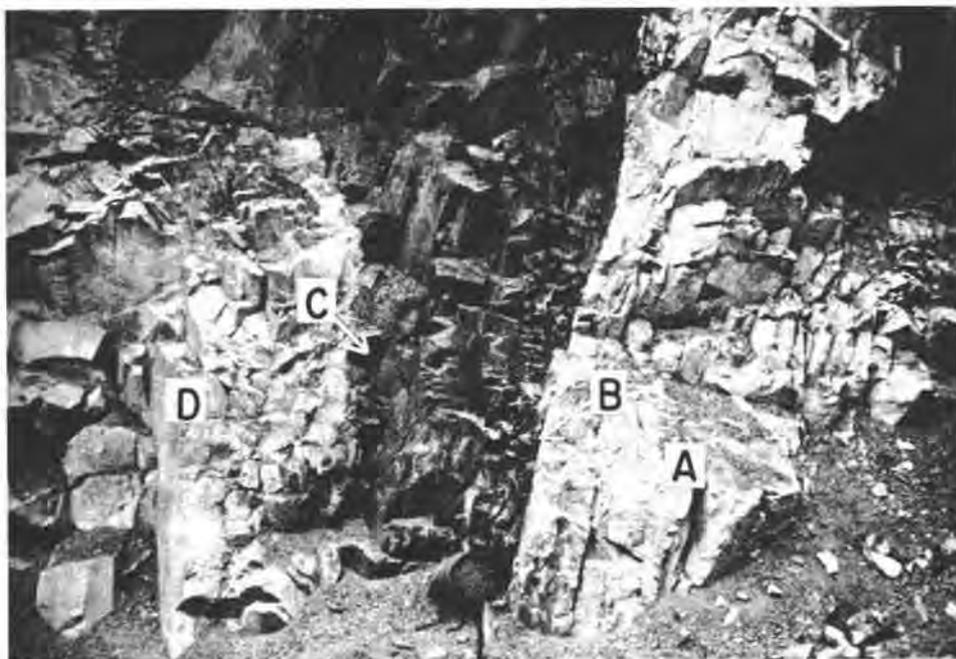


Figure 4.—A tensional dike-within-a-dike structure along a roadcut in Corbaley Canyon. A is the older dike that contacts gneiss to the immediate right of the photo. B is the sharp chilled margin of the inner (younger) dike against the outer dike; the composition of the margin is more mafic than the main part of either dike, as indicated by the darker color. C is the gradational contact of the chilled margin into the main part of the inner dike. D is the main interior part of the inner dike; although D is felsic like A, the two dikes are texturally distinct. The mirror image of the relationships shown in the photo occurs to the immediate left of the photo. Rucksack indicates scale.

have been breached by valleys that passed into adjacent depositional basins now covered by the Columbia River Basalt Group.

It is necessary to assign at least a component of northeast-southwest crustal extension to the Chiuwaukum graben during its active stage. Table 1 indicates (with some uncertainty) a possibility of northeast-southwest compression toward the waning stages of Chumstick deposition. This is based on the recognition of northwest-trending folds in the Chumstick (including the major Eagle Creek anticline) as well as minor northwest-trending thrust faults (Whetten and Laravie, 1976). However, Whetten (1977) and Buza (1979) believe that a bedrock horst was present in the graben during Chumstick deposition at the site now occupied by the Eagle Creek anticline. It is possible that the Eagle Creek structure is a faulted drape fold over the horst and was produced during extension of the graben, rather than by later compression. Smaller folds and thrust faults could have been produced by post-34 m.y.B.P. compressive deformation. Thus the evidence does not demand a pre-Wenatchee period of compressive deformation, which would require a reversal of tectonic stress during the waning

stages of graben activity. Minor intrusive igneous activity continued within the Chiuwaukum graben during its waning stages, as indicated by K-Ar dates of 41-46 m.y.B.P. on both mafic and felsic intrusive rocks.

Between 34 and 40(?) m.y.B.P., the region was in a period of tectonic and perhaps magmatic quiescence. The younger limit is well constrained by the age of the Wenatchee Formation, which rests with profound unconformity on all older units. The older limit has the uncertainty associated with the range of fission-track ages of the Chumstick Formation and the time necessary to deposit the Nahahum (upper) member, which lacks tuff beds. The period of quiescence and deep erosion coincides with the same phenomenon found over most of western North America (Coney, 1971, 1972; Gresens, 1978, 1980, 1981) and is particularly well documented in the southern Rockies (Epis and Chapin, 1975). It is considered to mark the end of Laramide deformation in western North America (Coney, 1971, 1972). During this period topography that existed in Chumstick time was destroyed by erosion.

Oligocene Events

The age of the Wenatchee Formation is defined by two precise fission-track dates averaging 34 m.y.B.P. (Gresens and others, 1981); an Oligocene age also has been determined from palynological data (Newman, 1977). The Wenatchee Formation is about 280 m thick and consists mainly of quartzose sandstone and partly tuffaceous shale and siltstone that are interpreted, respectively, as fluvial sand of braided stream complexes and associated overbank deposits (Gresens and others, 1981). At the end of the late Eocene-early Oligocene erosional episode, the area from central Washington to the Puget Sound region must have been a relatively low-lying plain, perhaps with local monadnocks of more resistant rocks.¹

The Chiwaukum graben was subjected to compressive deformation between 29 and 34 m.y.B.P. (Gresens, 1980). During this episode, Chumstick rocks were thrust out of the graben and across the trace of the Entiat fault along northeast-directed thrust faults; for reasons of scale, they are not shown on Figure 1. The overlying Wenatchee was folded and faulted. The younger age limit is defined by the presence of hornblende andesite that intrudes a thrust plane in Wenatchee Formation but is not itself sheared. The total calculated shortening along the east side of the graben is only 2.5-3.0 km. This deformation documents northeast-southwest crustal compression. Minor thrust faults involving the Chumstick, mapped by Whetten and Laravie (1976) in the north end of the graben, could be related to this episode, but there are no narrow constraints on their age.

Intrusions of hornblende andesite (Horse Lake Mountain complex), mainly in the form of sills with associated dikes, are extensive in the southern part of the Chiwaukum graben. Most are exposed in rather monotonous Chumstick terrane, but some intrude the Wenatchee Formation near the eastern border of the graben. K-Ar ages from various sources give an average age of about 29 m.y.

The Horse Lake Mountain complex was described in part by Bayley (1965), who mapped many of the

sills and dikes and examined their petrography. Complete mapping and interpretation of the complex is presented by Gresens [1983]. The complex ranges in composition from basaltic andesite to andesite. A variety of textural types are represented, but many contain phenocrystic hornblende that ranges to several decimeters in longest axis. Many sills and dikes are altered to chlorite, zeolite, and calcite. Calcite partially replaces hornblende or plagioclase cores in some rocks.

A striking feature of the complex is the thin screens of undisturbed sedimentary rock, sometimes less than one meter thick, between numerous sills. Excellent exposures in some cliffs with 500 m of vertical exposure show sections that are made of greater than 50 percent igneous materials, yet the sedimentary beds show no signs of forceful emplacement.

Figure 5 shows a WNW-trending structure that appears to have cut across at least part of the Chiwaukum graben at about 29 m.y.B.P. The fault zone labeled "A" on Figure 5 is an old fault contemporaneous with graben activity. Fault "B" is a post-34 m.y. B.P. fault that involves the Wenatchee Formation, but its possible relationship to fault zone "C" cannot be determined. At fault zone "C," a belt of Wenatchee Formation occupies a small graben in the Chumstick Formation, and the fault zone is thus post-34 m.y.B.P. The northwest projection of the fault zone cannot be directly mapped because of the lack of markers. However, a broad belt of disturbed beds marks the presumed projection of the fault zone. South of the fault zone, Chumstick bedding strikes northwest, which is the general structural grain of the Chumstick Formation within the graben. North of the fault zone, the Chumstick bedding strikes uniformly north to slightly northeast. As this zone is followed north, the strikes swing back to the northwest through a hinge line. At four places along the line, individual tuff beds can be followed unbroken through the hinge. Angular unconformities between the Wenatchee and Chumstick Formations prove that Chumstick bedding was already tilted prior to post-Wenatchee faulting. The structure shown here suggests deformation by an underlying right-lateral shear couple that has broken through the Chumstick beds to the south but only rotated them about a hinge line to the north. The intervening area has the appearance of a giant "kink band," in which the steeply dipping Chumstick beds were caught. The main center of intrusive activity of the 29-m.y.-old hornblende andesites is within the fault zone, and some intrusions extend south of it into northwest-striking Chumstick beds. But most of the intrusive activity is confined to the area within the giant "kink band," suggesting that tensional stretching within the band allowed contemporaneous passive intrusion of the many sills. One sill (not shown on the scale of

¹ Rocks of late Eocene age on the east side of the Puget Sound region are essentially coal-bearing arkosic sandstone (Tiger Mountain Formation) and volcanogenic and tuffaceous sandstone, siltstone, and shale (Tukwila Formation), both of continental origin (Vine, 1969). The younger Renton Formation, which apparently is broadly coeval with the development of the late Eocene-early Oligocene erosion surface to the east, contains continental sandstones that are more quartz-rich, less plagioclase-rich, and more kaolinitic than the older units (Vine, 1969). The uppermost (early Oligocene) part of the Renton Formation contains fossils diagnostic of brackish water. Within Puget Sound proper, there is an occurrence of marine rocks (the Blakely Formation) that are essentially coeval with the Renton Formation. These have been interpreted as a possible submarine fan complex (McLean, 1977). The strand line during the late Eocene and early Oligocene thus must have been west of the Cascade Range and near the Puget Sound lowland. Transitional rocks deposited on the low-lying plain between the Wenatchee area and the old strand line have been removed by Cascade uplift and erosion.

21 N., R. 21 E. and sec. 31, T. 22 N., R. 21 E. (Gresens and others, 1981) (Fig. 1). These beds are about 50 m thick and are composed primarily of bentonitic shale with minor sand stringers and thin carbonaceous interbeds probably derived from humic material. The unit probably represents shallow swampy lakes or over-bank deposits. Structurally and stratigraphically, the unit near Malaga is positioned between Wenatchee Formation and younger Columbia River Basalt Group. Preliminary palynological data (Estella Leopold, personal commun., 1978) suggest that the age is Miocene.

In the area southwest of the graben, an unnamed unit has a stratigraphic position between the Roslyn Formation and the Columbia River Basalt Group. The generally poor exposures in roadcuts along the Teanaway River drainage do not permit diagnostic assignment of the unit to Wenatchee Formation or to a unit of Miocene age at the time of this writing.

The Wenatchee Formation was eroded before deposition of the Columbia River Basalt Group, and the thinning is progressively greater from east to west (Gresens, 1980). It is not possible to obtain a definitive tectonic interpretation from the scanty data, but the observation would be consistent with the beginning of the present north-south trend of the modern Cascade Range uplift between 34 and 15 m.y B.P.. Similarly, it is not possible to state whether the Miocene beds near Malaga were of significantly greater extent and were stripped along with the underlying Wenatchee Formation. It is possible that they are in part the erosional debris of the stripped Wenatchee Formation; the Wenatchee contains red oxidized siltstone that could be the source for red shales in the Miocene unit.

The geologic record of this area ends with outpouring of the Columbia River Basalt Group. These Miocene basalts and the post-basalt history are not discussed; the reader is referred to Loeschke (1979).

REFERENCES CITED

- Armstrong, R. L., 1978, Cenozoic igneous history of the U.S. cordillera from lat. 42° to 49°N. In Smith, R. B.; Eaton, G. P., editors, 1978, Cenozoic tectonics and regional geophysics of the western cordillera: Geological Society of America Memoir 152, p. 265-282.
- Bayley, E. P., Jr., 1965, Bedrock geology of the Twin Peaks area, and intrusive complex near Wenatchee, Washington: University of Washington Master of Science thesis, 47 p., 2 pl.
- Bressler, C. T., 1951, The petrology of the Roslyn arkose, central Washington: Pennsylvania State College Doctor of Philosophy thesis, 195 p., 11 p.
- Buza, J. W., 1979, Dispersal patterns and paleogeographic implications of lower and middle Tertiary fluvial sandstones in the Chiwaukum graben, east-central Cascade Range, Washington. In Armentrout, J. M.; Cole, M. R.; Ter Best, Harry, Jr., editors, 1979, Cenozoic paleogeography of the western United States; Pacific Coast Paleogeography Symposium 3: Society of Economic Paleontologists and Mineralogists, Pacific Section, p. 63-73.
- Cashman, S. M.; Whetten, J. T., 1976, Low-temperature serpentinization of peridotite fanglomerate on the west margin of the Chiwaukum graben, Washington: Geological Society of America Bulletin, v. 87, no. 12, p. 1773-1776.
- Clayton, D. N., 1973, Volcanic history of the Teanaway Basalt, east-central Cascade Mountains, Washington: University of Washington Master of Science thesis, 55 p., 3 pl.
- Coney, P. J., 1971, Cordilleran tectonic transitions and motion of the North American plate: Nature, v. 233, no. 5320, p. 462-465.
- Coney, P. J., 1972, Cordilleran tectonics and North America plate motion: American Journal of Science, v. 272, no. 7, p. 603-628.
- Epis, R. C.; Chapin, C. E., 1975, Geomorphic and tectonic implications of the post-Laramide, late Eocene erosion surface in the southern Rocky Mountains. In Curtis, B. F., editor, 1975, Cenozoic history of the southern Rocky Mountains: Geological Society of America Memoir 144, p. 45-74.
- Ewing, T. E., 1980, Paleogene tectonic evolution of the Pacific Northwest: Journal of Geology, v. 88, no. 6, p. 619-638.
- Foster, R. J., 1958, The Teanaway dike swarm of central Washington: American Journal of Science, v. 256, no. 9, p. 644-653.
- Frizzell, V. A., Jr., 1979, Petrology and stratigraphy of Paleogene nonmarine sandstones, Cascade Range, Washington: U.S. Geological Survey Open-File Report 79-1149, 151 p., 3 pl.
- Frizzell, V. A., Jr.; Tabor, R. W., 1977, Stratigraphy of Tertiary arkoses and their included monolithologic fanglomerates and breccias in the Leavenworth fault zone, central Cascades, Washington [abstract]: Geological Society of America Abstracts with Programs, v. 9, no. 4, p. 421.
- Gresens, R. L., 1978, Regional significance of a late Eocene/early Oligocene erosion surface exposed in central Washington State [abstract]: Geological Society of America Abstracts with Programs, v. 10, no. 3, p. 107-108.
- Gresens, R. L., 1979, Timing of Cenozoic igneous and tectonic events on the eastern flank of the Cascade Range in central Washington [abstract]: Geological Society of America Abstracts with Programs, v. 11, no. 3, p. 80.
- Gresens, R. L., 1980, Deformation of the Wenatchee Formation and its bearing on the tectonic history of the Chiwaukum graben, Washington, during Cenozoic time: Geological Society of America Bulletin, v. 91, no. 1, p. 4-7 (Part I), p. 115-155 (Part II).
- Gresens, R. L., 1981, Extension of the Telluride erosion surface to Washington State, and its regional and tectonic significance: Tectonophysics, v. 79, no. 3/4, p. 145-64.
- Gresens, R. L., 1982, Early Cenozoic geology of central Washington State; II, Implications for plate tectonics and alternatives for the origin of the Chiwaukum graben: Northwest Science, v. 56, no. 4, p. 259-264.
- Gresens, R. L., 1983, Geology of the Wenatchee and Monitor quadrangles, Chelan and Douglas Counties, Washington: Washington Division of Geology and Earth Resources Bulletin 75, 75 p., 3 pl., scale 1:24,000.
- Gresens, R. L.; Naeser, C. W.; Whetten, J. T., 1981, Stratigraphy and age of the Chumstick and Wenatchee Formations—Tertiary fluvial and lacustrine rocks, Chiwaukum graben, Washington: Geological Society of America Bulletin, v. 92, no. 5, p. 233-236 (Part I), p. 841-875 (Part II).
- Lamey, C. A.; Hotz, P. E., 1952, The Cle Elum River nickeliferous iron deposits, Kittitas County, Washington: U.S. Geological Survey Bulletin 978-B, p. 27-67.

- Loeschke, Jorge, 1979, Basalts of Oregon (U.S.A.) and their geotectonic environment; I, Petrochemistry of Tertiary basalts of the Oregon Coast Range: *Neues Jahrbuch für Mineralogie Abhandlungen*, v. 134, no. 3, p. 225-247.
- Lupher, R. L., 1944, Stratigraphic aspects of the Blewett-Cle Elum iron ore zone, Chelan and Kittitas Counties, Washington: Washington Division of Geology Report of Investigations 11, 63 p., 2 pl.
- McLean, Hugh, 1977, Lithofacies of the Blakeley Formation, Kitsap County, Washington—A submarine fan complex?: *Journal of Sedimentary Petrology*, v. 47, no. 1, p. 78-88.
- Newman, K. R., 1977, Palynologic biostratigraphy of some early Tertiary nonmarine formations in central and western Washington [abstract]: *Geological Society of America Abstracts with Programs*, v. 9, no. 7, p. 1113.
- Southwick, D. L., 1966, Petrography, chemistry, and possible correlations of the Camas Land sill and Teanaway dike swarm, central Washington: *Northwest Science*, v. 40, no. 1, p. 1-16.
- Tabor, R. W.; Frizzell, V. A., Jr., 1977, Tertiary stratigraphy of the central Cascades Mountains, Washington State; Part II, Stratigraphy of the Swauk, Silver Pass, Teanaway, and Roslyn Formations. In Brown, E. H.; Ellis, R. C., editors, 1977, *Geological excursions in the Pacific Northwest*: Western Washington University, p. 88-100.
- Vance, J. A., 1977, Fission track geochronology of the Tertiary volcanic rocks of the central Cascade Mountains, Washington [abstract]: *Geological Society of America Abstracts with Programs*, v. 9, p. 250.
- Vance, J. A., 1979, Early and middle Cenozoic arc magmatism and tectonics in Washington State [abstract]: *Geological Society of America Abstracts with Programs*, v. 11, no. 3, p. 132.
- Vance, J. A.; Naeser, C. W., 1977, Fission track geochronology of the Tertiary volcanic rocks of the central Cascade Mountains, Washington [abstract]: *Geological Society of America Abstracts with Programs*, v. 9, no. 4, p. 520.
- Vine, J. D., 1969, Geology and coal resources of the Cumberland, Hobart, and Maple Valley quadrangles, King County, Washington: U.S. Geological Survey Professional Paper 624, 67 p.
- Waters, A. C., 1926, The petrography of Corbaley Canyon: University of Washington Bachelor of Science thesis, 77 p., 1 pl.
- Whetten, J. T., 1977, Tertiary stratigraphy of the central Cascades Mountains, Washington State. Part III. Rocks of the Chumstick Creek and Nahahum Canyon areas. In Brown, E. H.; Ellis, R. C., editors, 1977, *Geological excursions in the Pacific Northwest*: Western Washington University, p. 100-109.
- Whetten, J. T.; Laravie, J. A., 1976, Preliminary geologic map of the Chiwaukum 4 NE quadrangle, Chiwaukum graben, Washington: U.S. Geological Survey Miscellaneous Field Studies Map MF-794, 1 sheet, scale 1:24,000.

EARLY CENOZOIC GEOLOGY OF CENTRAL WASHINGTON STATE: II. IMPLICATIONS FOR PLATE TECTONICS AND ALTERNATIVES FOR THE ORIGIN OF THE CHIWAUKUM GRABEN¹

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ABSTRACT

Reconstruction of tectonic events of central Washington prior to about 50 million years before the present is tenuous. The early Eocene Swauk Formation may have been deposited after a period of tectonic quiescence. Folding of Swauk on east-west axes between 51 and 47 million years before the present cannot be readily correlated with presumed plate motions. Coeval emplacement (at 47-48 million years ago) of two tensional dike swarms (Teaway and Corbaley Canyon), of contrasting lithologies and extensional directions, presents a tectonic paradox that is not easily solved. The paradox may be resolved by assuming sufficient right-lateral strike-slip motion on the east-bounding fault of the Chiwaukum graben (the Entiat fault) so that the counterpart of the Teaway dikes lies somewhere southeast below the Columbia Plateau. Two models for development of the Chiwaukum graben from about 46 to 40 million years ago by strike-slip motion are possible. The first requires a wedge-shaped graben; this is the "ball-bearing" model that reconciles the Chiwaukum graben with documented clockwise rotation of Eocene rocks that lie to the west. A preferred model for the graben is as a rhombochasm ("pull-apart" structure), such as has been proposed for Tertiary basins of California. By this model, a primary kink in a strike-slip fault opens into a diamond-shaped depression during active movement. The exposed Chiwaukum graben is thus interpreted as the northern end of a rhombochasm that projects beneath the Columbia Plateau. Cessation of graben activity at about 40 million years ago agrees with the end of Laramide deformation over western North America, and it is correlated with global reorganization of plate motions. A period of tectonic and magmatic quiescence from 40 to 34 million years ago correlates with development of a major erosion surface over western North America during plate reorganization. Deformation of the Wenatchee Formation during 34-29 million years ago marks the onset of renewed tectonism and magmatism as subduction was initiated in the Pacific Northwest. Deformation of the southern part of the Chiwaukum graben by right-lateral shearing and associated emplacement of the Horse Lake Mountain complex may be related to the initiation of strike-slip motion elsewhere along the Pacific margin as the spreading center between the Farallon and Pacific plates encountered a subduction zone at the continental margin. Erosional stripping of the Wenatchee Formation in the region now occupied by the Cascade Range occurred before deposition of the Columbia River Basalt Group in Miocene time and may mark the beginning of the ancestral Cascades.

INTRODUCTION

Carlson (1976) correlated Cenozoic plate motions with the geology of the continental margin of the Pacific Northwest (Coast Range and Puget-Willamette Lowland) and the Cascade Range of Washington-Oregon. The summary of geologic history presented in Part I (Gresens, 1982) supplements the correlation of plate motion with the geology of the central Cascade Range of Washington.

PLATE TECTONIC IMPLICATIONS

Early Eocene

The lateritic paleosol present at the base of the Swauk Formation and the probable widespread occurrence of the unit suggest that it may have been deposited on an erosion surface developed during a time of tectonic and magmatic quiescence. It is difficult to attempt correlation with plate motions because the age of the presumed erosion surface is not well documented, lying between 88 m.y. (age of the Mt. Stuart granodiorite; Engels and Crowder, 1971) and some-

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what less than the 51-m.y. age of upper Swauk Formation (Tabor and Frizzell, 1977). Ericson and Williams (1976) used fission-track apparent ages and annealing temperatures of apatite to suggest that the Mount Stuart batholith was uplifted and unroofed at 55 ± 6 m.y.B.P.. This date could reflect the onset of renewed tectonism after a period of crustal stability, but considerable uncertainty remains.

East-west fold axes, suggestive of north-south compression, pre-date Teanaway Basalt and therefore were formed in the interval of about 51-47 m.y.B.P. (Gresens, 1979; Ewing, 1980). This compressive direction does not fit well into a plate tectonic model. During the Cretaceous and into the early Tertiary, the Kula plate was moving northward with respect to the Pacific plate (Hayes and Pitman, 1970) as the Pacific plate itself was moving northward relative to a presumably fixed "hot-spot" (Clague and Jarrard, 1973). Interaction of either plate with North America could produce north-south compression and east-west fold axes. However, North America, moving relatively westward, was presumably interacting with the Farallon plate moving relatively eastward (Atwater, 1970; Hayes and Pitman, 1970); these movements should have produced north-south fold axes. It is possible that this area was involved in crustal rotation, as observed for Eocene rocks to the west and southwest (Simpson and Cox, 1977; Beck, 1976), so that they no longer reflect the true direction of tectonic stresses during early Eocene folding.

Armstrong (1978), Vance (1979), and Ewing (1980) suggested that a broad northwest-southeast subduction-related magmatic arc, the Challis arc, was active in early and middle Eocene time in British Columbia, Washington, central Oregon, Idaho, Montana, and Wyoming. Much of the igneous activity of the Chiwaukum graben area, including the Teanaway Basalt and the Corbaley Canyon dike swarm, would be included in Challis volcanism. Definition of the magmatic arc is based on regional plots of occurrences of plutonic or volcanic rocks and their ages. Failure to examine the tectonic setting and the tectonic implications of the petrology of individual occurrences leads to over-simplification in broad overviews of this type. For example, the Corbaley Canyon tensional bimodal dike swarm is better explained by deep crustal rifting than by a subduction mechanism.

Ewing (1980) attempted to explain east-west fold axes in the Swauk Formation and the northeast trend of the tensional Teanaway dikes by a single stress regime having a principal stress direction oriented slightly east of north. The change from north-south compression to WNW-ESE extension was presumed to reflect an exchange of the intermediate and least prin-

cipal stresses at about 47 m.y.B.P. But such an exchange should produce NNW dextral or NNE sinistral strike-slip faults, not the extensional rifting indicated by the Teanaway dikes. Clearly, much work remains before a definitive correlation between early Eocene plate motions and igneous and tectonic events is possible for this region.

Middle Eocene

The Chiwaukum graben became active immediately after or during the close of the early Eocene events. Geologists familiar with the Pacific Northwest have long speculated informally on the possibility of strike-slip motion on the Entiat fault, and the suggestion was recently formalized (Gresens, 1979; Ewing, 1980). There is no evidence that the fault existed prior to about 46 m.y.B.P. Features suggestive of strike-slip movement include (1) the linearity of the fault, (2) the failure of the Jurassic ultramafic rocks to appear east of the Chiwaukum graben, and (3) the failure of the northeast-trending Teanaway dike swarm to appear east of the Chiwaukum graben. Ewing (1980) considered the Entiat to be part of a network of strike-slip faults that were active in the Eocene.

The new radiometric dates obtained during this investigation emphasize another geologic incompatibility across the Entiat fault. The Teanaway and Corbaley Canyon dike swarms are essentially coeval. Yet they represent markedly different compositions and crustal extension in directions nearly 90° apart, even though they are geographically only 50 km apart. These observations present a fundamental tectonic problem if they are in their original positions with respect to each other. Two possibilities for the origin of the graben, each calling for strike-slip motion on the Entiat fault, are presented.

Model One—Rotation and Strike-Slip Movement

Figure 1a shows the outline of the graben. There is a basic wedge shape, although the irregular Leavenworth fault zone must be given a "best fit." Figure 1c shows the schematic development of the graben by a combination of strike-slip motion and clockwise rotation. Although the wedge shape could be explained by rotation alone, restoration of the rotation would make the Teanaway-Corbaley paradox an even greater tectonic problem (Fig. 1b).

Clockwise rotation of crustal rocks of Eocene age along the Oregon and Washington coasts is well documented by paleomagnetic data (Beck, 1976; Simpson and Cox, 1977; Globberman, 1979; Beck and Burr, 1979). It could be argued that there must be a transi-

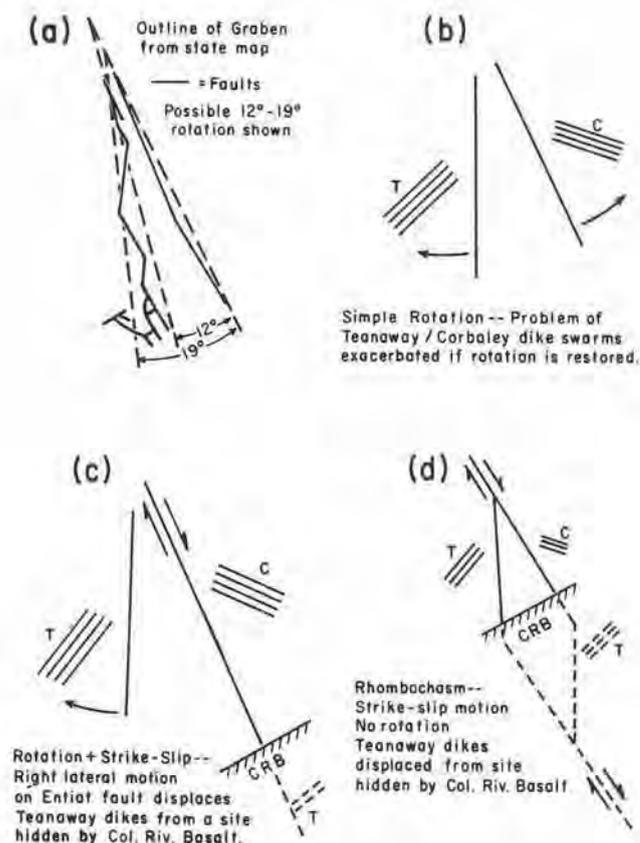


Figure 1.—Possible origins for the Chiwaukum graben. T and C on the diagrams indicate the Teanaway and Corbaley Canyon dike swarms, respectively. CRB refers to Columbia River Basalt Group, which covers the southern projection of the Chiwaukum graben. Further explanation is in the text.

tion, either gradually over a broad zone or abruptly over a narrow zone, from the marginal part of the continent that is strongly affected by rotation into the stable continental interior that has not been rotated. The Entiat fault zone and the Chiwaukum graben could be an abrupt transitional area. This model follows the "ball-bearing" model for combined right-lateral strike-slip motion and clockwise rotation as proposed by Beck (1976).

Model Two—Rhombochasm

The geometry of the Chiwaukum graben also lends itself to interpretation as a rhombochasm (Carey, 1958), with the southern half covered by the Columbia River Basalt Group (Fig. 1d). This model does not require a rotational component, but merely strike-slip motion along a fault having an original "kink." This is, for example, a plausible model for the origin of late Cenozoic basins of California (Crowell, 1974). The Chiwaukum graben has the correct geome-

try for this type of "pull-apart basin" as defined by Crowell. The side bounded by strike-slip movement is straight (Entiat fault), whereas the tensionally rifted "kink" boundary is irregular (Leavenworth fault).

Either of these models would explain the mismatch of geology across the Chiwaukum graben and would alleviate the tectonic paradox posed by the coeval Teanaway and Corbaley dike swarms; sufficient separation by restoration of strike-slip movement might make separate stress regimes plausible. Either model would allow the gradual opening of a depression that could ultimately accommodate 5,800 m of Chumstick sediment. The author prefers the second model, but either is possible. The models differ in terms of the possibility of differential rotation across the graben, but this may be tested by paleomagnetic work on the Teanaway Basalt and Corbaley Canyon rocks. The required paleomagnetic work is now in progress (M. E. Beck, Jr., personal commun., 1979, 1980).

Late Eocene–Early Oligocene

The Chiwaukum graben became inactive at about 40 m.y.B.P., and a period of tectonic quiescence ensued. This period coincides with the end of Laramide deformation over the entire western Cordillera, which is in turn correlated with changes in motion of the North American and Pacific plates (Coney, 1971, 1972; Epis and Chapin, 1975; Gresens, 1978, 1980) and probably with a global reorganization of plate motion during the Eocene (Rona and Richardson, 1978; Gresens, [1981]).

Late Oligocene

Post-34-m.y.B.P. compression of the region must correlate with renewed tectonism and magmatism over much of the western Cordillera at about 35 m.y.B.P. (Epis and Chapin, 1975) and, more specifically, with the renewal of subduction in the Pacific Northwest (Vance, 1979). Compressive stress was transmitted to the continental interior.

The spreading center between the Pacific plate and the Farallon plate encountered a subduction zone at the western margin of North America at about 29 m.y.B.P. (Atwater, 1979), and this initiated right-lateral strike-slip deformation at the continental margin near San Francisco. The passive emplacement of the 29-m.y. Horse Lake Mountain complex is associated with right-lateral strike-slip motion, and it is suggested that this may be a more northerly expression of renewed strike-slip motion on the North American continent (Gresens, 1980). Central Washington is admittedly rather far from San Francisco, but the timing and sense of motion coincide.

CONCLUDING REMARKS

The part of the North American continental margin now represented by the state of Washington has a long history of right-lateral strike-slip faulting, including the Late Cretaceous (Davis and others, 1978), as well as the middle Eocene and late Oligocene events described above. These apparently are interspersed with "conventional" continental margin subduction and related igneous activity and with at least one period of tectonic quiescence. However, any regional tectonic synthesis must take into account the factual chronology of geologic events presented in Part I of this paper.

ACKNOWLEDGMENTS

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REFERENCES CITED

- Armstrong, R. L., 1978, Cenozoic igneous history of the U.S. cordillera from lat. 42° to 49° N. *In* Smith, R. B.; Eaton, G. P., editors, 1978, Cenozoic tectonics and regional geophysics of the western Cordillera: Geological Society of America Memoir 152, p. 265-282.
- Atwater, Tanya, 1970, Implications of plate tectonics for the Cenozoic tectonic evolution of western North America: Geological Society of America Bulletin, v. 81, no. 12, p. 3513-3535.
- Beck, M. E., Jr., 1976, Discordant paleomagnetic pole positions as evidence of regional shear in the western cordillera of North America: American Journal of Science, v. 276, no. 6, p. 694-712.
- Beck, M. E., Jr.; Burr, C. D., 1979, Paleomagnetism and tectonic significance of the Goble volcanic series, southeastern Washington: Geology, v. 7, no. 4, p. 175-179.
- Carey, S. W., 1958, The tectonic approach to continental drift. *In* Carey, S. W., convener, 1958, Continental drift—A symposium: University of Tasmania Geology Department, p. 177-355.
- Carlson, R. L., 1976, Cenozoic plate convergence in the vicinity of the Pacific Northwest, a synthesis and assessment of plate tectonics in the northeastern Pacific: University of Washington Doctor of Philosophy thesis, 129 p.
- Clague, D. A.; Jarrard, R. D., 1973, Tertiary Pacific plate motion deduced from the Hawaiian-Emperor chain: Geological Society of America Bulletin, v. 84, no. 4, p. 1135-1154.
- Coney, P. J., 1971, Cordilleran tectonic transitions and motions of the North American plate: Nature, v. 233, no. 5320, p. 462-465.
- Coney, P. J., 1972, Cordilleran tectonics and North America plate motion: American Journal of Science, v. 272, no. 7, p. 603-628.
- Crowell, J. C., 1974, Origin of Late Cenozoic basins in southern California. *In* Dickinson, W. R., editor, Tectonics and Sedimentation: Society of Economic Paleontologists and Mineralogists Special Publication 22, p. 190-204.
- Davis, G. A.; Monger, J. W. H.; Burchfiel, B. C., 1978, Mesozoic construction of the Cordilleran "collage", central British Columbia to central California. *In* Howell, D. G.; McDougall, K. A., editors, 1978, Mesozoic paleogeography of the western United States; Pacific Coast Paleogeography Symposium No. 2: Society of Economic Paleontologists and Mineralogists, Pacific Section, p. 1-32.
- Engels, J. C.; Crowder, D. F., 1971, Late Cretaceous fission-track and potassium-argon ages of the Mount Stuart granodiorite and Beckler Peak stock, north Cascades, Washington: U.S. Geological Survey Professional Paper 750-D, p. D39-D43.
- Epis, R. C.; Chapin, C. E., 1975, Geomorphic and tectonic implications of the post-Laramide, late Eocene erosion surface in the southern Rocky Mountains. *In* Curtis, B. F., editor, 1975, Cenozoic history of the southern Rocky Mountains: Geological Society of America Memoir 144, p. 45-74.
- Erikson, E. H., Jr.; Williams, A. E., 1976, Implications of apatite fission track ages in the Mount Stuart batholith, Cascade Mountains, Washington [abstract]: Geological Society of America Abstracts with Programs, v. 8, no. 3, p. 372.
- Ewing, T. E., 1980, Paleogene tectonic evolution of the Pacific Northwest: Journal of Geology, v. 88, no. 6, p. 619-638.
- Globerman, B. R., 1979, Geochemistry of Eocene basalts from the Black Hills, Washington Coast Range [abstract]: Geological Society of America Abstracts with Programs, v. 11, no. 3, p. 80.
- Gresens, R. L., 1978, Regional significance of a late Eocene/early Oligocene erosion surface exposed in central Washington State [abstract]: Geological Society of America Abstracts with Programs, v. 10, no. 3, p. 107.
- Gresens, R. L., 1979, Timing of Cenozoic igneous and tectonic events on the eastern flank of the Cascade Range in central Washington [abstract]: Geological Society of America Abstracts with Programs, v. 11, no. 3, p. 80.
- Gresens, R. L., 1980, Deformation of the Wenatchee Formation and its bearing on the tectonic history of the Chiwaukum graben, Washington, during Cenozoic time: Geological Society of America Bulletin, v. 91, no. 1, p. 4-7 (Part I); p. 115-155 (Part II).
- Gresens, R. L., 1981, Extension of the telluride erosion surface to Washington State and its regional and tectonic significance: Tectonophysics, v. 79, no. 3/4, p. 145-164.
- Gresens, R. L., 1982, Early Cenozoic geology of central Washington State. I. Summary of sedimentary, igneous, and tectonic events: Northwest Science, v. 56, no. 3, p. 218-229.
- Hayes, D. E.; Pitman, W. C., III, 1970, Magnetic lineations in the North Pacific. *In* Hayes, J. D., editor, 1970, Geological investigations of the North Pacific: Geological Society of America Memoir 126, p. 291-314.
- Rona, P. A.; Richardson, E. S., 1978, Early Cenozoic global plate reorganization: Earth and Planetary Science Letters, v. 40, no. 1, p. 1-11.
- Simpson, R. W., Jr.; Cox, A. V., 1977, Paleomagnetic evidence for tectonic rotation of the Oregon Coast Range: Geology, v. 5, no. 10, p. 585-589.
- Tabor, R. W.; Frizzell, V. A., Jr., 1977, Tertiary stratigraphy of the central Cascades Mountains, Washington State. Part II. Stratigraphy of the Swauk, Silver Pass, Teanaway, and Roslyn Formations. *In* Brown, E. H.; Ellis, R. C., editors, 1977, Geological excursions in the Pacific Northwest: Western Washington University, p. 88-100.
- Vance, J. A., 1979, Early and middle Cenozoic arc magmatism and tectonics in Washington State [abstract]: Geological Society of America Abstracts with Programs, v. 11, no. 3, p. 132.

DISTRIBUTION MAPS OF STRATIGRAPHIC UNITS OF THE COLUMBIA RIVER BASALT GROUP

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INTRODUCTION

The accompanying 19 outline maps have been compiled by the authors, from their work in various parts of the Columbia Plateau. The maps update those published by Swanson and others (1979) and represent our current knowledge of the distribution of the principal mappable stratigraphic units (flows or groups of flows), including known feeder dikes and vents.

Figure 1 shows the basic geographic features of the study area. Figures 2 through 6 show distribution of the Imnaha Basalt and the Grande Ronde Basalt. Figures 7 through 11 show distribution of the Wanapum Basalt. Figures 12 through 15 show distribution of Saddle Mountains Basalt members older than the lower Snake River Canyon. Figures 16 through 19 show distribution of Saddle Mountains Basalt members younger than the lower Snake River Canyon, and Figure 20 shows those basalt units along the southern margin of the Columbia Plateau that have discrete feeder systems and whose affinity to the Columbia River Basalt Group is uncertain. Strati-

graphic relations and approximate ages are given in Hooper and Swanson, Figure 4 (this volume).

The Saddle Mountains Basalt and Wanapum Basalt are mapped as individual members or flows. The Imnaha Basalt and Grande Ronde Basalt contain many flows not easily distinguished in the field; they have been mapped across the plateau as informal magnetopolarity units (R_0 , N_0 , R_1 , N_1 , R_2 , N_2). Drill core from the Yakima-Ellensburg area shows 5,000 ft of Grande Ronde Basalt, the top in magnetopolarity unit R_2 ; the magnetic polarity of the core samples cannot be determined, but the total thickness suggests either an unusual thickness of N_1 or the presence of R_1 forming the base of the basalt pile in the Pasco basin.

The magnetopolarity unit to which individual Grande Ronde or Imnaha dikes belong usually cannot be determined. The many dikes belonging to these two formations and to the Picture Gorge-Monument dike swarm are shown diagrammatically; map symbols for these dikes indicate orientations, geographic extent, and zones of maximum concentration, rather than individual rock bodies.

Petrographic and chemical descriptions of most units are given in Swanson and others (1979). Other flows are described by Uppuluri (1974), McDougall (1976), Camp (1981), Hooper and others (1984), and other publications listed below.

Note Added In Proof:

The text and illustrations for this paper are essentially as prepared in 1983. They therefore do not reflect information about the extent of various flows of the Columbia River Basalt Group that has been gathered by the authors or others since that time.

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Flows of Grande Ronde Basalt exposed in the walls of the Grande Ronde River canyon, Asotin County, Washington. (Photograph by Washington State Department of Commerce and Economic Development)

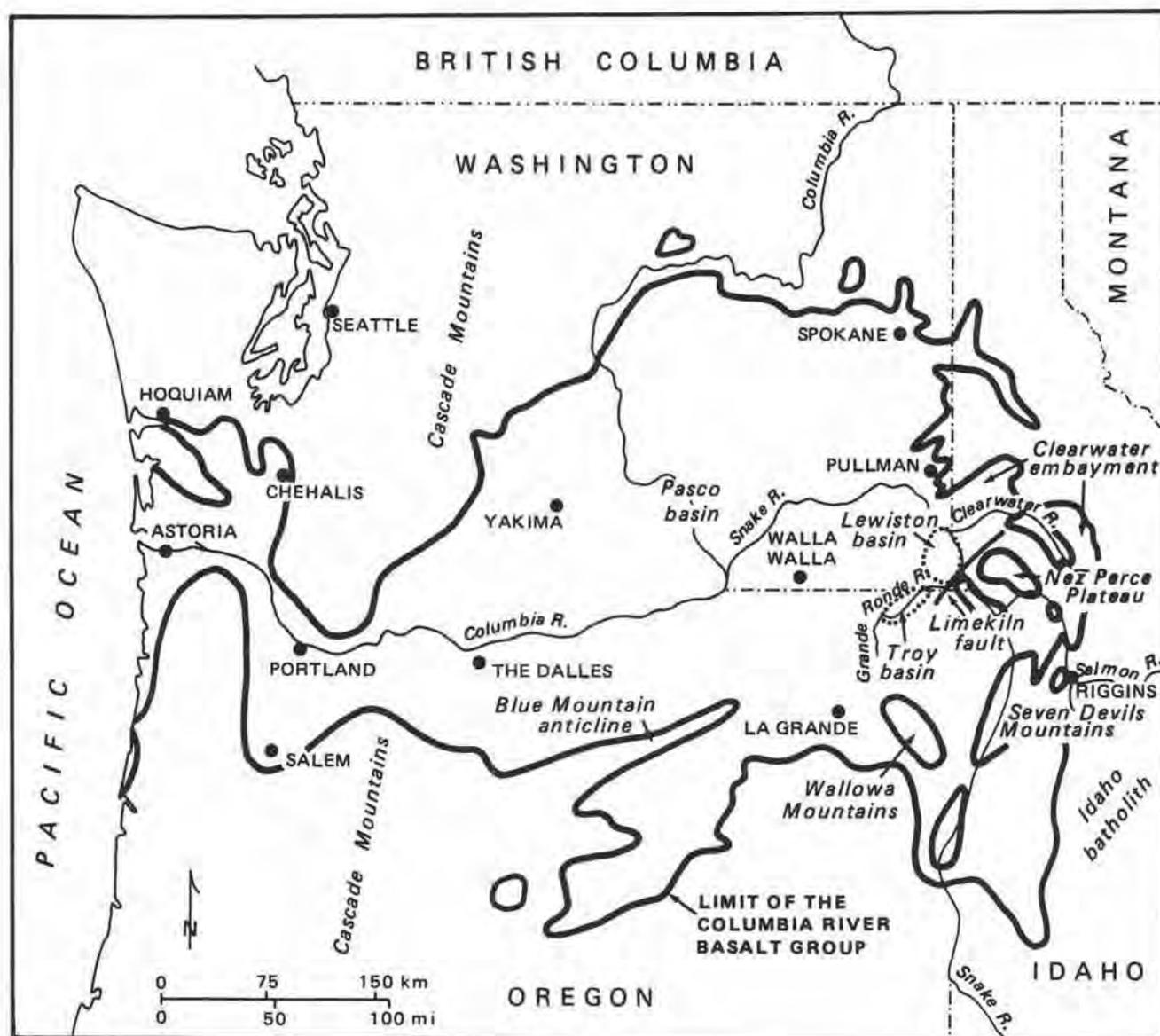


Figure 1.—Outcrop area of the Columbia River Basalt Group, with some related geographic and structural features.

REFERENCES USED IN COMPILATION

- Anderson, J. L., 1978, The stratigraphy and structure of the Columbia River Basalt in the Clackamas River drainage: Portland State University Master of Science thesis, 136 p.
- Anderson, J. L., 1980, Pomona Member of the Columbia River Basalt Group—An intracanyon flow in the Columbia River Gorge, Oregon: *Oregon Geology*, v. 42, no. 12, p. 195-199.
- Anderson, J. L.; Vogt, B. F., 1987, Intracanyon flows of the Columbia River Basalt Group in the southwest part of the Columbia Plateau and adjacent Cascade Range, Oregon and Washington. In Schuster, J. E., editor, 1987, *Selected papers on the geology of Washington*: Washington Division of Geology and Earth Resources Bulletin 77, (this volume).
- Beeson, M. H.; Moran, M. R., 1979, Columbia River Basalt Group stratigraphy in western Oregon: *Oregon Geology*, v. 41, no. 1, p. 11-14.
- Beeson, M. H.; Moran, M. R.; Anderson, J. L.; Vogt, B. F., 1982, The relationship of the Columbia River Basalt Group to the geothermal potential of the Mount Hood area, Oregon. In Priest, G. R.; Vogt, B. F., editors, 1982, *Geology and geothermal resources of the Mount Hood area, Oregon*: Oregon Department of Geology and Mineral Industries Special Paper 14, p. 43-46.
- Camp, V. E., 1976, Petrochemical stratigraphy and structure of the Columbia River basalt, Lewiston Basin area, Idaho-Washington: Washington State University Doctor of Philosophy thesis, 201 p.

(References continued following figures.)

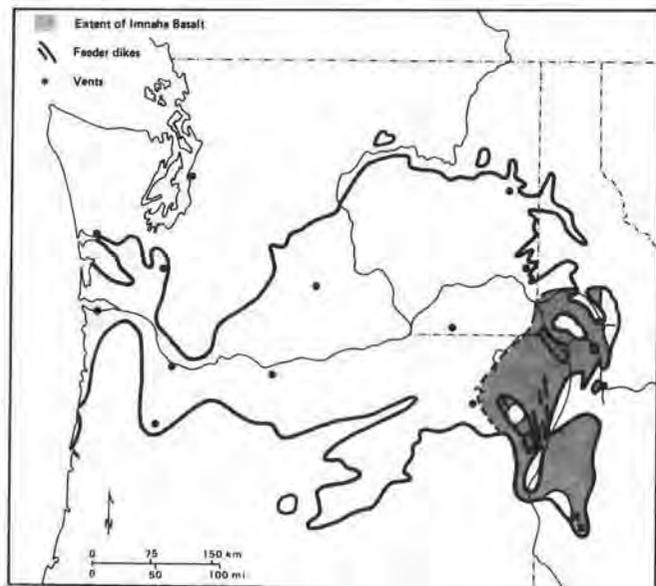


Figure 2.—Imnaha Basalt. All flows have normal polarity except a few flows at the base (marked by X) and a few flows at the top with transitional to reversed magnetic polarity. Thickness decreases from southeast (700 m) to northwest (100 m). Imnaha Basalt flows typically are coarsely phyrlic. Two chemical types (American Bar and Rock Creek) are recognized.

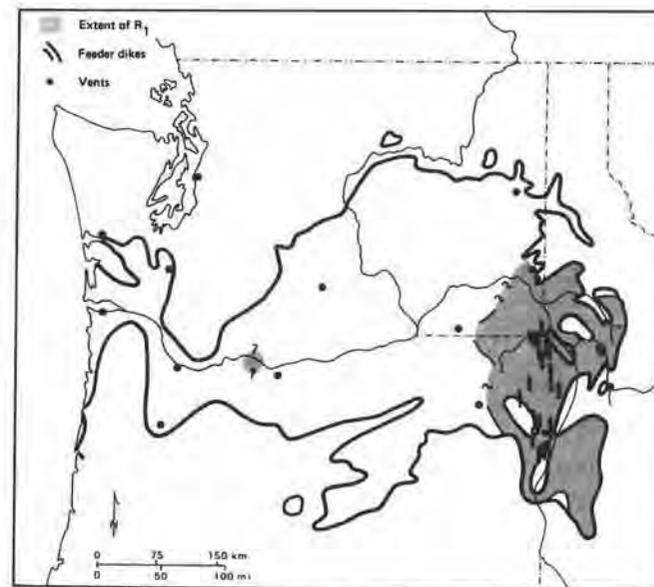


Figure 3.— R_1 , the lower of two units of the Grande Ronde Basalt with reversed magnetic polarity. Thickness greatest (400 m) near Wallowa Mountains and Seven Devils Mountains.

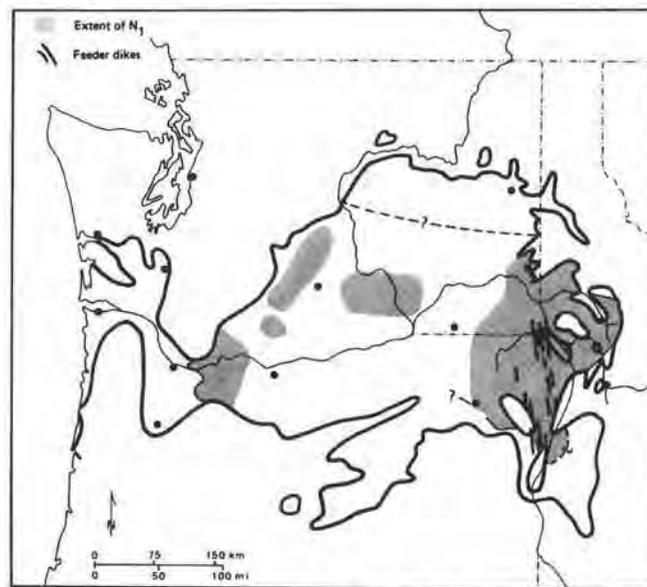


Figure 4.— N_1 , the lower of two units of the Grande Ronde Basalt with normal magnetic polarity. In the southeast sector, thickness increases toward the northwest. Areas with no data left blank.

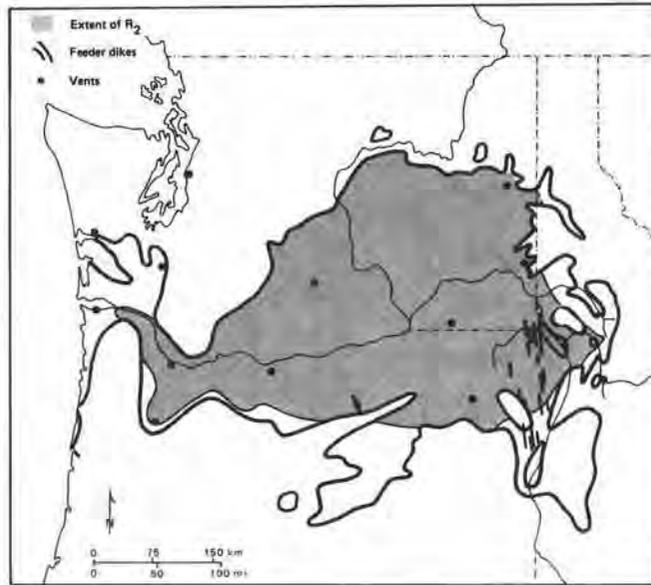


Figure 5.— R_2 , the higher of two units of the Grande Ronde Basalt with reversed magnetic polarity. Laps out on the northwest side of the Wallowa and Seven Devils Mountains and extends down the Columbia River almost as far as Astoria.

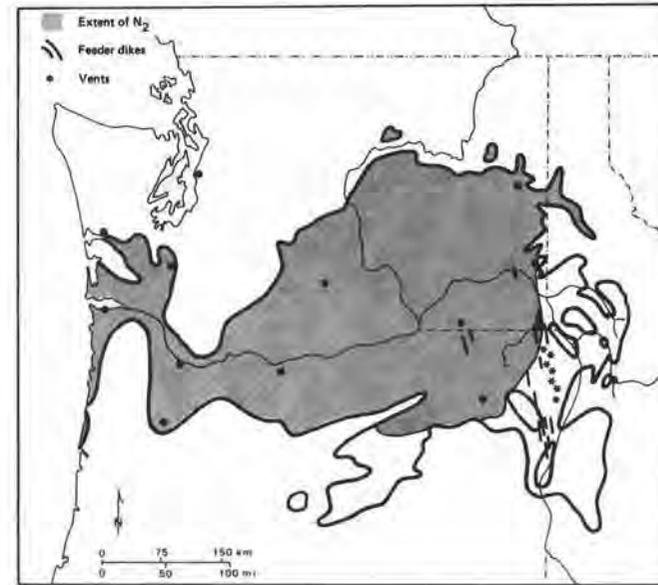


Figure 6.— N_2 , the higher of two units of the Grande Ronde Basalt with normal magnetic polarity. Vents are present in the elevated block southeast of the Limekiln fault, but flows are limited to the northwest side of that structure. Flows of this group extend to the northern limits of the Columbia Plateau and reach the Pacific Ocean.

Figures 2-6.—Imnaha Basalt and Grande Ronde Basalt. Each unit represents many flows; their total volume represents more than 90 percent of the total Columbia River Basalt Group volume. Shaded areas in these figures indicate the extent of the described flows. Specific flow-dike correlations normally cannot be made for these units, so dikes are shown diagrammatically (as short straight lines), illustrating general distribution, maximum concentration, and orientation. Vents are indicated by asterisks. (See Fig. 1 for location and identification of geographic and structural features.)

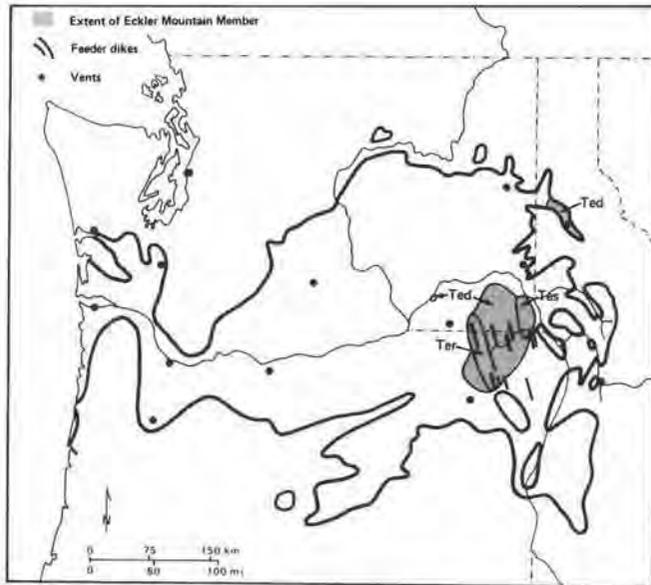


Figure 7.—Eckler Mountain Member. Three major flows are recognized: Robinette Mountain flow (Ter), with two well-developed feeder dikes; Dodge flow (Ted), with many feeder dikes and a probable vent in northern Idaho; Shumaker Creek flow (Tes), with feeder dike (short straight lines) on the south of outcrop area. The Eckler Mountain Member has normal magnetic polarity and was erupted during a hiatus in volcanic activity over most of the Columbia Plateau, corresponding to the development of the sedimentary Vantage Member of the Ellensburg Formation.

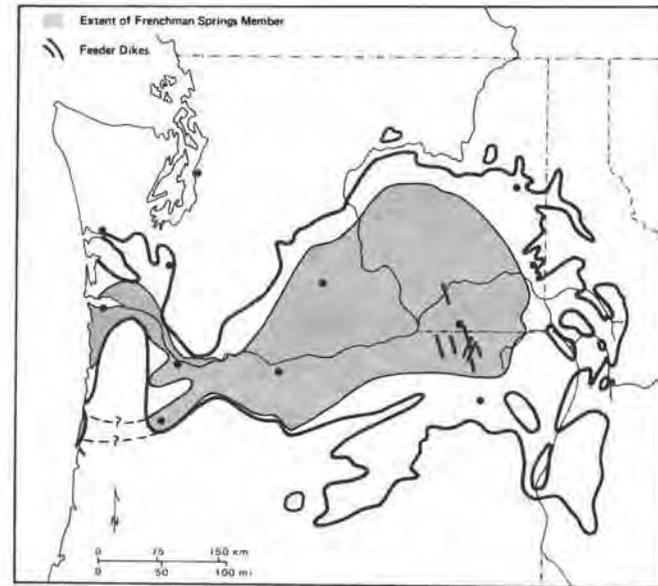


Figure 8.—Frenchman Springs Member. A number of flows of similar chemical composition, all with normal magnetic polarity. Individual dikes are shown.

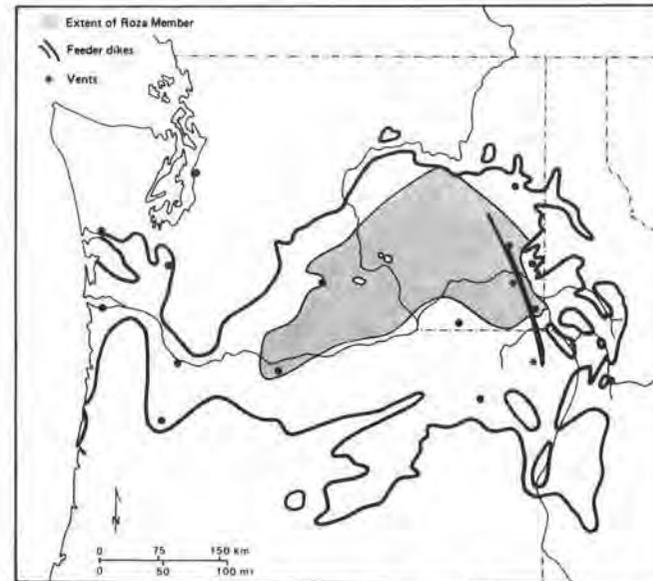


Figure 9.—Roza Member. At least two major flows with transitional to reversed polarity. The fissure-vent system is particularly well documented. More than 20 vent areas have been identified. Marked vents are diagrammatic only.

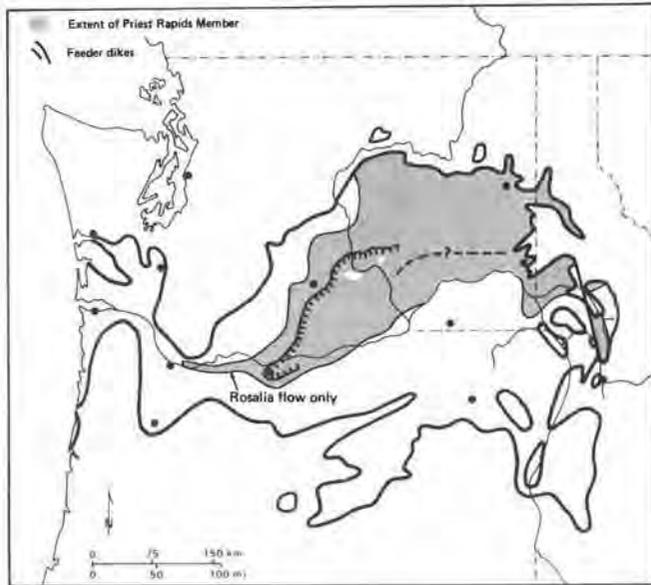


Figure 10.—Priest Rapids Member. Two major flows (Rosalia and Lolo), with reversed magnetic polarity. Feeder dikes located in the Clearwater embayment. Dashed line represents the southern limit of the Rosalia flow; hachured line represents the northern and western limit of the Lolo flow.

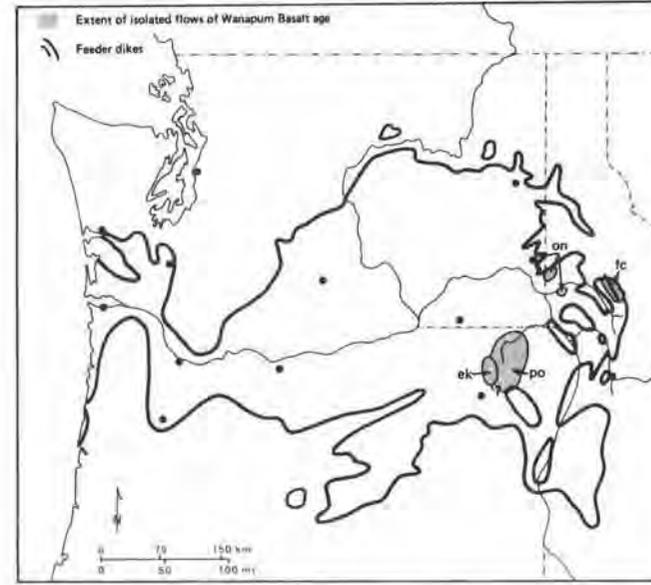


Figure 11.—Isolated flows of Wanapum Basalt age: on, Onaway flow; fc, Feary Creek flow (with feeder dike); po, Powatka flow; ek, Looking Glass flow. Exact stratigraphic positions are not known.

Figures 7-11.—Wanapum Basalt. Shaded areas in these figures indicate the extent of the described member or flows. Dikes and vents shown as heavy lines and asterisks, respectively. (See Fig. 1 for location and identification of geographic and structural features.)

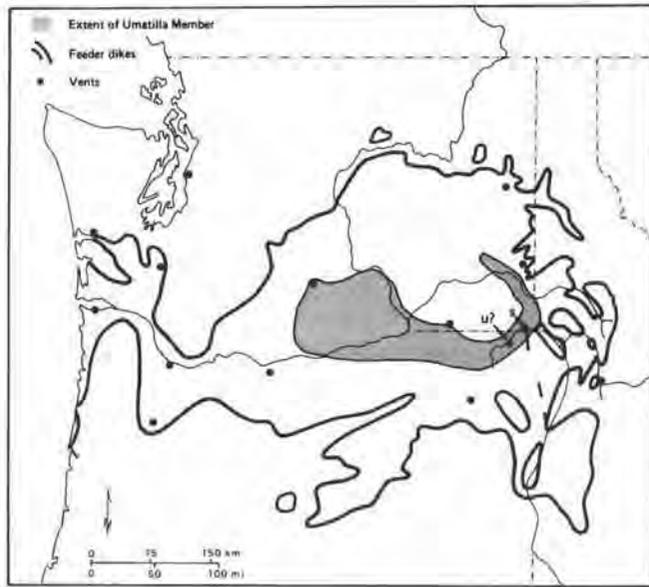


Figure 12.—Umatilla Member. Two flows (Umatilla and Sillusi), with very similar distribution. A Sillusi vent (s) at Puffer Butte and feeder dikes are shown. A possible Umatilla vent (u?) is present in the Troy basin. The Umatilla lavas flowed west, entering the Pasco basin from the southeast.

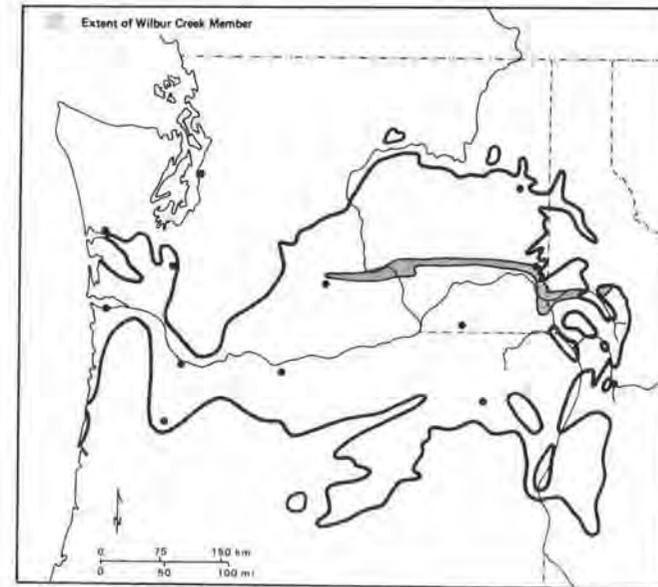


Figure 13.—Wilbur Creek Member (including the Wahluke flow in the Saddle Mountains area in the Pasco basin and the Lapwai flow on the east). Flows of the Asotin Member and Wilbur Creek Member, like the Umatilla and Sillusi flows, filled the Lewiston basin and then followed a shallower drainage to the northwest to enter the Pasco basin on its northeast corner; neither dikes nor vents have been certainly identified, but their presence in the Clearwater embayment is probable.

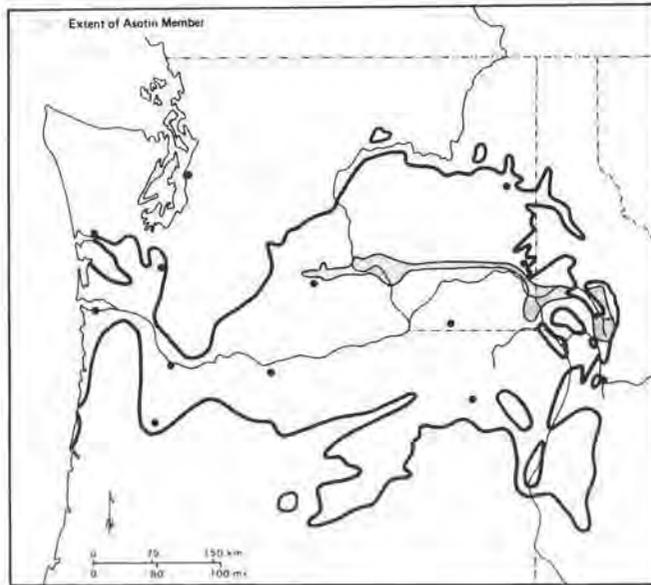


Figure 14.—Asotin Member (including the Huntzinger flow in north Pasco basin). Flows of the Asotin Member and Wilbur Creek Member, like the Umatilla and Sillusi flows, filled the Lewiston basin and then followed a shallower drainage to the northwest to enter the Pasco basin on its northeast corner; neither dikes nor vents have been certainly identified, but their presence in the Clearwater embayment is probable.

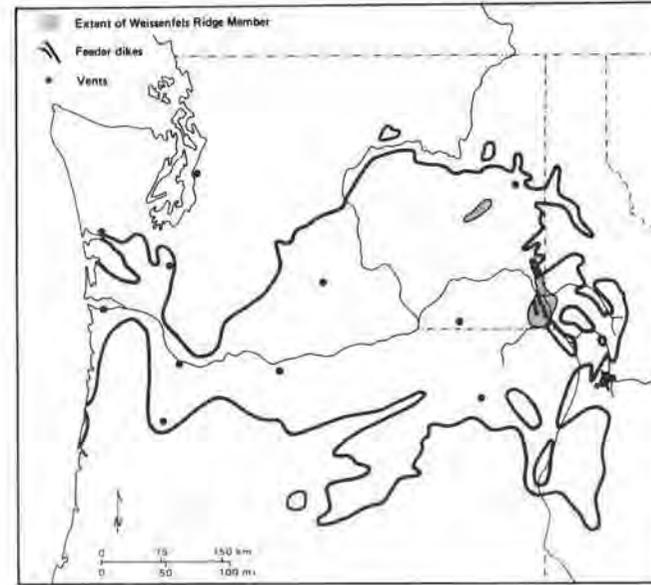


Figure 15.—Weissenfels Ridge Member. Four flows recognized in the Lewiston basin. Only the Lewiston Orchards flow extends beyond the basin. Flows of similar composition occur southwest of Spokane and north of the Lewiston basin, and a vent and dike are present near Riggins in Idaho.

Figures 12-15.—Saddle Mountains Basalt (flows older than the development of the lower Snake River Canyon). Shaded areas indicate the extent of the described members; dikes and vents as in preceding figures. (See Fig. 1 for location and identification of geographic and structural features.)

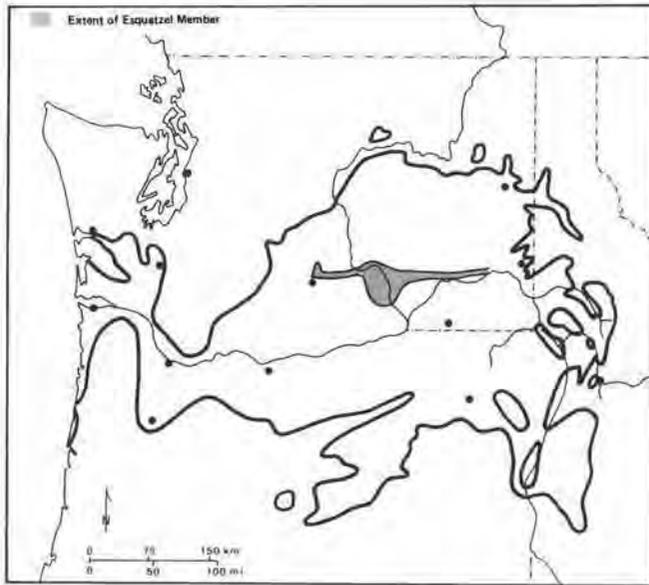


Figure 16.—Esquatzel Member. Fills the lower Snake River Canyon. No identified feeder system.

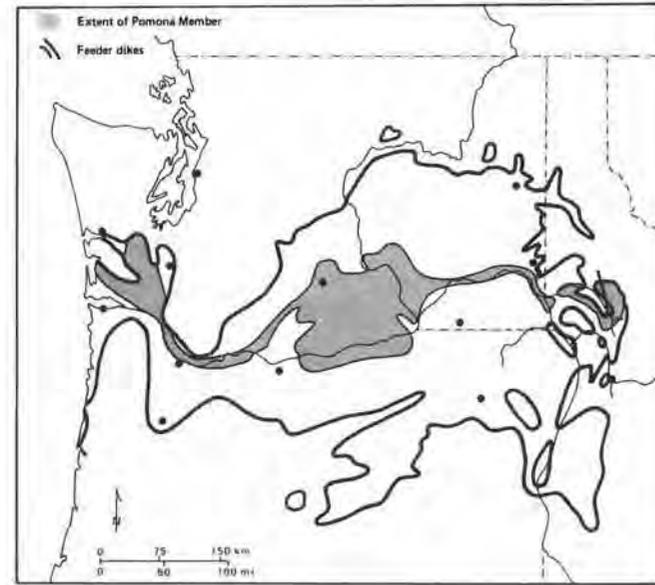


Figure 17.—Pomona Member. Feeder dike in Idaho. Flow followed the Clearwater and Snake River Canyon systems to the Pasco basin, then down the old Columbia River drainage southwest from Yakima to the present Columbia Gorge, reaching the Pacific Ocean between Hoquiam and Astoria.

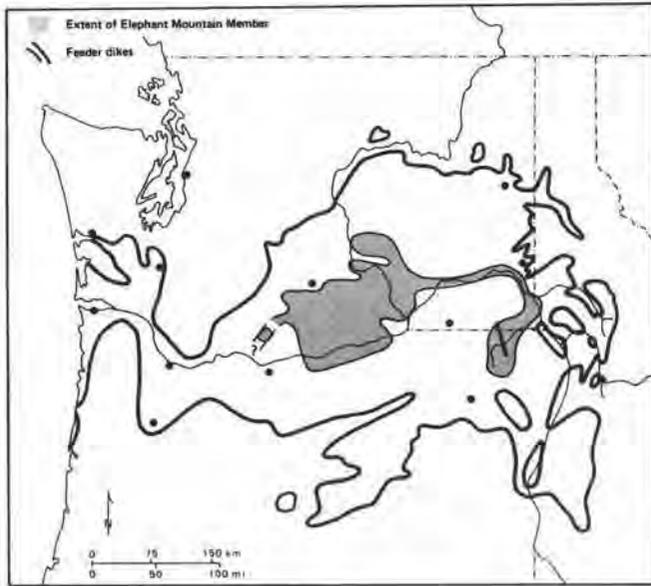


Figure 18.—Elephant Mountain Member. Feeder dike in the Troy basin. Flow moved northeast into the Lewiston basin, followed the Pomona flow down the Snake River Canyon to the Pasco basin, and then apparently followed the same early Columbia River drainage.

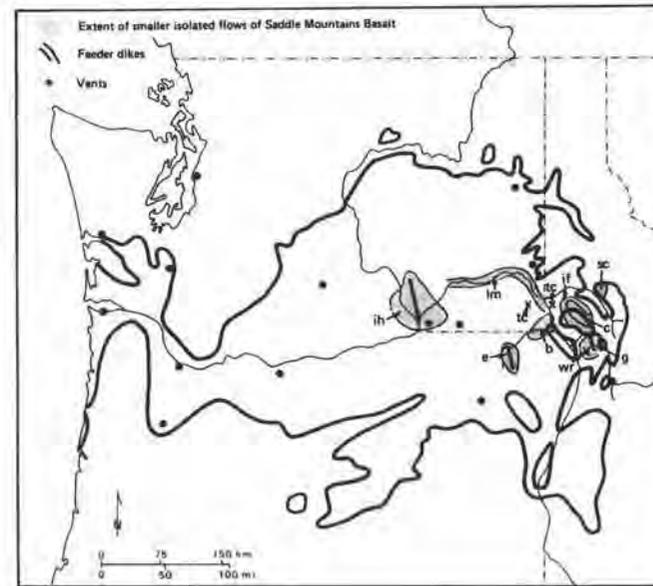


Figure 19.—Smaller, isolated flows of Saddle Mountains Basalt (stratigraphic order largely unknown): ih, Ice Harbor Member (four flows in the Pasco basin with vent and dikes); e, Eden flow; b, Buford Member; sc, Swamp Creek Member; if, Icicle Flat Member; g, Grangeville Member; wr, Windy Ridge flow; tc, Tammany Creek flow; lm, Lower Monumental Member, filling the lower Snake River Canyon; c, Craigmont flow.

Figures 16-19.—Saddle Mountains Basalt (flows younger than the development of the lower Snake River Canyon). Shaded areas indicate extent of described members or flows; dikes and vents as in preceding figures. (See Fig. 1 for location and identification of geographic and structural features.)

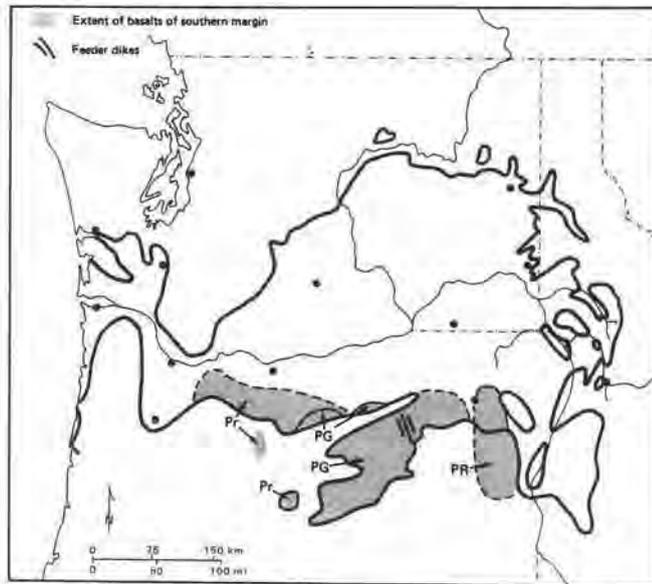


Figure 20.—Basalts with discrete feeder systems developed along the southern margin of the Columbia Plateau. Pr, basalt of Prineville; PG, Picture Gorge Basalt, with Monument dike-swarm feeders; PR, basalts, andesites, and other volcanic rocks of Powder River, associated with development of the La Grande graben. (See preceding figures and Fig. 1 for location and identification of geologic, geographic, and structural features.)

(References, continued)

- Camp, V. E., 1981, Geologic studies of the Columbia Plateau—Part II. Upper Miocene basalt distribution, reflecting source locations, tectonism, and drainage history in the Clearwater embayment, Idaho: *Geological Society of America Bulletin*, v. 92, no. 9, Part I, p. 669-678.
- Camp, V. E.; Hooper, P. R., 1981, Geologic studies of the Columbia Plateau—Part I. Late Cenozoic evolution of the southeast part of the Columbia River Basalt Province: *Geological Society of America Bulletin*, v. 92, no. 9, Part I, p. 659-668.
- Hammond, P. E., 1980, Reconnaissance geologic map and cross-sections of southern Washington Cascade Range, latitude 45°30'–47°15' N., longitude 120°45'–122°22.5' W.: Portland State University Department of Earth Sciences, 31 p., 2 sheets, scale 1:125,000.
- Hooper, P. R.; Swanson, D. A., 1987, Evolution of the eastern part of the Columbia Plateau. In Schuster, J. E., editor, 1987, *Selected papers on the geology of Washington*: Washington Division of Geology and Earth Resources Bulletin 77, (this volume).
- Hooper, P. R.; Webster, G. D., 1982, Geology of the Pullman, Moscow West, Colton and Uniontown 7½' quadrangles, Washington and Idaho: Washington Division of Geology and Earth Resources Geologic Map GM-26, 1 sheet, scale 1:62,500.
- Hooper, P. R.; Kleck, W. D.; Knowles, C. R.; Reidel, S. P.; Thiessen, R. L., 1984, Imnaha Basalt, Columbia River Basalt Group: *Journal of Petrology*, v. 25, part 2, p. 473-500.
- Hooper, P. R.; Webster, G. D.; Camp, V. E., 1985, Geologic map of the Clarkston 15 minute quadrangle, Washington and Idaho: Washington Division of Geology and Earth Resources Geologic Map GM-31, scale 1:48,000.
- Kleck, W. D., 1976, Chemistry, petrography, and stratigraphy of the Columbia River Group in the Imnaha River valley region, eastern Oregon and western Idaho: Washington State University Doctor of Philosophy thesis, 203 p.
- McDougall, Ian, 1976, Geochemistry and origin of basalt of the Columbia River Group, Oregon and Washington: *Geological Society of America Bulletin*, v. 87, no. 5, p. 777-792.
- Murphy, T. M., 1981, Geology of the Nicolai Mountain-Gnat Creek area, Clatsop County, northwest Oregon: Oregon State University Master of Science thesis, 355 p.
- Myers, C. W.; Price, S. M.; and others, 1979, Geologic studies of the Columbia Plateau—A status report: Rockwell Hanford Operations RHO-BWI-ST-4, 541 p., 53 pl.
- Portland General Electric Company, 1974, Pebble Springs Nuclear Plant; Preliminary Safety Analysis Report: U.S. Nuclear Regulatory Commission Docket Nos. 50-514/515, 9 v.
- Reidel, S. P., 1978, The stratigraphy and petrogenesis of the Grande Ronde Basalt in the lower Salmon and adjacent Snake River Canyons: Washington State University Doctor of Philosophy thesis, 415 p.
- Reidel, S. P.; Fecht, K. R., 1981, Wanapum and Saddle Mountains Basalts of the Cold Creek syncline area. In Myers, C. W.; Price, S. M., editors, 1981, *Subsurface Geology of the Cold Creek syncline*: Rockwell Hanford Operations RHO-BWI-ST-14, p. 3-1 - 3-45.
- Reidel, S. P.; Ledgerwood, R. K.; Myers, C. W.; Jones, M. G.; Landon, R. D., 1980, Rate of deformation in the Pasco basin during the Miocene as determined by distribution of Columbia River basalt flows: Rockwell Hanford Operations RHO-BWISA-29, 43 p.
- Reidel, S. P.; Ledgerwood, R. K.; Myers, C. W.; Jones, M. G.; Landon, R. D., 1980, Rate of deformation in the Pasco basin during the Miocene as determined by distribution of Columbia River basalt flows [abstract]: *Geological Society of America Abstracts with Programs*, v. 12, no. 3, p. 149.
- Ross, M. E., 1978, Stratigraphy, structure, and petrology of Columbia River Basalt in a portion of the Grande Ronde River-Blue Mountains area of Oregon and Washington: University of Idaho Doctor of Philosophy thesis, 407 p.
- Snavely, P. D., Jr.; MacLeod, N. S.; Wagner, H. C., 1973, Miocene tholeiitic basalts of coastal Oregon and Washington and their relations to coeval basalts of the Columbia Plateau: *Geological Society of America Bulletin*, v. 84, no. 4, p. 387-424.
- Swanson, D. A.; Anderson, J. L.; Bentley, R. D.; Byerly, G. R.; Camp, V. E.; Gardner, J. N.; Wright, T. L., 1979, Reconnaissance geologic map of the Columbia River Basalt Group in

- eastern Washington and northern Idaho: U.S. Geological Survey Open-File Report 79-1363, 26 p., 12 pl., scale 1:250,000.
- Swanson, D. A.; Anderson, J. L.; Camp, V. E.; Hooper, P. R.; Taubeneck, W. H.; Wright, T. L., 1981, Reconnaissance geologic map of the Columbia River Basalt Group, northern Oregon and western Idaho: U.S. Geological Survey Open-File Report 81-797, 35 p., 6 sheets, scale 1:250,000.
- Swanson, D. A.; Wright, T. L.; Camp, V. E.; Gardner, J. N.; Helz, R. T.; Price, S. M.; Ross, M. E., 1977, Reconnaissance geologic map of the Columbia River Basalt Group, Pullman and Walla Walla quadrangles, southeast Washington and adjacent Idaho: U.S. Geological Survey Open-File Report 77-100, 10 p., 2 sheets, scale 1:250,000.
- Swanson, D. A.; Wright, T. L.; Camp, V. E.; Gardner, J. N.; Helz, R. T.; Price, S. M.; Reidel, S. P.; Ross, M. E., 1980, Reconnaissance geologic map of the Columbia River Basalt Group, Pullman and Walla Walla quadrangles, southeast Washington and adjacent Idaho: U.S. Geological Survey Miscellaneous Investigations Map I-1139, 2 sheets, scale 1:250,000.
- Swanson, D. A.; Wright, T. L.; Helz, R. T., 1975, Linear vent systems and estimated rates of magma production and eruption for the Yakima Basalt on the Columbia Plateau: *American Journal of Science*, v. 275, no. 8, p. 877-905.
- Swanson, D. A.; Wright, T. L.; Hooper, P. R.; Bentley, R. D., 1979, Revisions in stratigraphic nomenclature of the Columbia River Basalt Group: U.S. Geological Survey Bulletin 1457-G, 59 p.
- Sylvester, K. J., 1978, Geophysical investigations of the hydrogeology of the Goldendale-Centerville area, Washington: Washington State University Master of Science thesis, 160 p., 1 pl.
- Timm, Susan, 1979, The structure and stratigraphy of the Columbia River Basalt in the Hood River Valley, Oregon: Portland State University Master of Science thesis, 60 p., 2 pl.
- Tolan, T. L., 1982, The stratigraphic relationships of the Columbia River Basalt Group in the lower Columbia River Gorge of Oregon and Washington: Portland State University Master of Science thesis, 151 p., 1 pl.
- Tolan, T. L.; Beeson, M. H., 1984, Intracanyon flows of the Columbia River Basalt Group in the lower Columbia River Gorge and their relationship to the Troutdale Formation: *Geological Society of America Bulletin*, v. 85, no. 4, p. 463-477.
- Uppuluri, V. R., 1974, Prineville chemical type—A new basalt type in the Columbia River Group: *Geological Society of America Bulletin*, v. 85, no. 8, p. 1315-1318.
- Van Diver-Powell, Lorraine, 1979, The structure and correlation of the Grande Ronde Basalt on Tygh Ridge, Wasco County, Oregon: University of Idaho Master of Science thesis, 57 p.
- Vogt, B. F., 1981, The stratigraphy and structure of the Columbia River Basalt Group in the Bull Run watershed, Multnomah and Clackamas Counties, Oregon: Portland State University Master of Science thesis, 151 p.
- Wells, R. E., 1982, Paleomagnetism and geology of Eocene volcanic rocks in southwest Washington—Constraints on mechanisms of rotation and their regional tectonic significance: University of California, Santa Cruz, Doctor of Philosophy thesis, 229 p., 1 pl.

EVOLUTION OF THE EASTERN PART OF THE COLUMBIA PLATEAU

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ABSTRACT

The eastern Columbia Plateau covers part of the western margin of the Precambrian and early Paleozoic craton of North America. Younger allochthonous terranes of oceanic and island-arc affinities were accreted to the craton from the west along a suture zone that is displaced left-laterally under the present Clearwater River and lower Snake River. Within the relatively thin, accreted oceanic crust immediately west of the suture, sheet-flood flows of basalt erupted from a north-northwest-trending fissure-vent system 17 to 6 million years ago to form the Columbia Plateau.

Eruptions began at the south end of the fissure system in the southeast corner of the plateau, where the coarsely phyrlic Imnaha Basalt filled deep pre-basalt valleys draining south and west. The subsequent, largely aphyric Grande Ronde Basalt erupted farther north as the more southerly valleys were filled to form a lava plain that was then tilted toward the northwest by the continuing rise of the Idaho batholith and associated plutonic rocks in the southeast. The northwesterly slope permitted the Grande Ronde Basalt to cover the western Columbia Plateau and ultimately to continue down the ancestral Columbia River to the Pacific Ocean.

Major eruptions of Wanapum Basalt occurred only northwest of the Limekiln fault (which further raised the southeast corner of the plateau), except in the topographic depression of the Clearwater valley. As the rate of eruptions waned through Wanapum and Saddle Mountains time, contemporaneous deformation became more conspicuous, and most Saddle Mountains flows unconformably fill tectonic and erosional basins and valleys.

Available evidence suggests that the plateau was being deformed throughout the eruptions. In addition to the continuing westerly and northwesterly tilting due to the rise of the southeast part of the Columbia Plateau, a north-northwest horizontal direction of maximum compression and a west-southwest horizontal direction of tension resulted not only in north-northwest-trending tensional feeder fissures, but also in northwest-trending dextral and northeast-trending sinistral strike-slip faulting and in east-trending folds and reverse faults.

The Columbia River Basalt Group is predominantly a quartz-tholeiite sequence rich in incompatible trace elements and with high Fe/Mg ratios, similar to rocks of other continental tholeiite provinces. The group is notably lacking in rocks of more extreme composition (picrites, alkaline rocks, andesites, or rhyolites) except in small volumes in marginal areas, where assignment to the group is uncertain. The voluminous individual flows (as much as 700 cubic kilometers) are homogeneous and almost devoid of xenoliths. A gradual increase in Sr-isotope ratios from Imnaha Basalt to Wanapum Basalt, with higher and more erratic ratios in the Saddle Mountains Basalt, may have been caused by crustal assimilation, but may have resulted also from vertical (but not lateral) heterogeneity in the mantle source. Major- and trace-element mass-balance calculations have failed to provide a convincing model, but they suggest that crystal fractionation and crustal assimilation contributed. In addition, either magma recharge and mixing, or a heterogeneous mantle source, or both, appear to be required.

INTRODUCTION

For 11 million years (17-6 m.y. ago) in the middle and late Miocene, eastern and central Washington were the scene of an immense volcanic event. Huge volumes of basaltic lava, represented by the Columbia River Basalt Group, erupted sporadically from fissures as much as 175 km long to flood a deeply eroded topography and form the Columbia Plateau. This pla-

teau now separates the Rocky Mountains and the Cascade Range, and the basalts, including extensions to the Pacific Ocean, cover about 160,000 km².

Throughout the eruptions, the developing plateau was being gradually deformed. The rise of granitic rocks of the Idaho batholith and associated plutons under the Rocky Mountains to the southeast and the growth of the Cascade Range to the west by arching



Figure 1.—Palouse topography of southeastern Washington: standing dunes of loess resting on a nearly horizontal surface of Columbia River basalt.

and calc-alkaline volcanic activity resulted in the continued re-formation of an asymmetric basin centered near Pasco. The basin had a relatively steep western flank against the Cascades and a long, gentle eastern flank. The WSW extension that caused the fissures was accompanied by a NNW compression that created nearly east-west anticlinal ridges, reverse faults, and northwest-trending dextral and northeast-trending sinistral strike-slip faults. Successive flows followed the lowest ground in this deforming pattern, and their distribution and thickness provide a guide to the changing topographic form of the plateau surface.

The rain shadow created by the rising Cascade Range in the late Miocene restricted erosion of the eastern surface of the plateau, which was subsequently covered by a veneer of loess and, in places, deposits from the Missoula floods (Fig. 1). The large rivers of the Columbia-Snake system, fed by the high rainfall in the Rocky Mountains to the east, have eroded deep canyons across the plateau, which greatly facilitate study of the flat to gently dipping basalt flows (Fig. 2).

Recent study of cuttings from a deep drill hole (Reidel and others, 1982) suggests that the basalt is more than 3,250 m thick in the Pasco area, thinning toward the plateau margin in all directions. If this is correct, then the total basalt volume may be more than 250,000 km³.

Determination of the stratigraphy of the basalt pile has been assisted by the use of precise chemical analyses and by magnetic polarity measured with a

small field fluxgate magnetometer to "fingerprint" individual flows or small groups of flows. The stratigraphy has now been mapped over the greater part of the plateau at a reconnaissance level (Swanson and others, 1979a,b,c; 1980; 1981), and more detailed mapping has been done in selected areas (Myers and others, 1979; Bentley and Campbell, 1983a,b; Hooper and Webster, 1982; Hooper and others, 1985; Swanson and Wright, 1983).

Distribution of stratigraphic units of the Columbia River Basalt Group is summarized in a series of maps by Anderson and others (this volume), and paleodrainage of the Columbia Plateau is summarized by Fecht and others (this volume).

The petrology and chemical composition of the basalt succession provide both a detailed view of the genesis of continental flood basalt and clues to the composition of the underlying crust and mantle. The geometry of the fissure systems and the deformation of the flows are of major significance to models of plate movement as the North American plate progressively overrode the northern extension of the East Pacific ridge-rift system.

THE PRE-BASALT BASEMENT

Beneath the eastern margin of the Columbia Plateau lies the suture zone that separates the Precambrian and early Paleozoic craton of North America to the east from a series of diverse allochthonous terranes to the west. The western terranes were accreted to the craton in Late Triassic to Cretaceous time and have since been rotated clockwise (Wilson and Cox, 1980; Beck and others, 1982; Fox and Beck, 1985).

The suture (Fig. 3) runs almost due south from the Canadian border and passes beneath the Columbia Plateau 65 km northwest of Spokane (Fox and Beck, 1985). Apparently it continues with a similar trend south of Pullman, for the steps of basement rock standing above the basalt surface between Spokane and Pullman (Steptoe, Kamiak, and Bald Buttes) have cratonic lithologies (Hooper and Webster, 1982). The suture reappears in Idaho, where it trends southward along the west side of the Idaho batholith. The apparent 120- to 150-km left-lateral displacement of the craton margin occurs along the lower Snake River and Clearwater River drainages where it is obscured by Columbia River basalt. The cratonic margin is clearly defined by lithologic differences. The margin is also marked by a break in the ⁸⁷Sr/⁸⁶Sr ratio (<0.704- >0.706) in Jurassic and younger plutonic rocks; Armstrong and others (1977) correlated this break with the differences in age and thickness of cratonic and accreted crust.



Figure 2.—Deep canyon of the lower Grande Ronde River in southeastern Washington, eroded through flat-lying flows of Columbia River basalt.

East of the suture, cratonic rocks include the late Precambrian Belt Supergroup and overlying lower Paleozoic metasediments, both of shelf depositional environment. These rocks overlie a much older, poorly exposed pre-Belt metamorphic terrane (Armstrong, 1975a,b, 1976) and are intruded by granitic plutons of Jurassic to Eocene age associated with the Idaho batholith and the Challis magmatic event (Bennett, 1980).

West of the suture, the allochthonous terranes have an oceanic association, with ophiolitic, island-arc, and back-arc-basin deposits (Brooks and Vallier, 1978; Dickinson and Thayer, 1978; Vallier, 1977; Vallier and others, 1977; Fox and Beck, 1985). Post-accretion intrusions of Jurassic to Oligocene age helped to weld the oceanic terranes together. North of the Columbia Plateau, Eocene volcanic rocks (O'Brien Creek Formation, Sanpoil Volcanics, Klondike Mountain Formation, and Scatter Creek dikes) are associated with and largely fill a NNE-trending graben. In northeastern and north-central Oregon, south of the Columbia Plateau, eruption of calc-alkaline volcanic rocks took place through the Eocene (Clarno Formation), Oligocene (John Day Formation), and Miocene (Strawberry volcanic rocks). Andesitic rocks of similar age to the Strawberry volcanic rocks appear to be associated in the La Grande and Baker areas with a NNW- to WNW-trending graben (Barrash and others, 1980; Hooper, unpub. data).

The origin of the accreted terranes remains unclear. The Seven Devils and Huntington arc terranes have similarly oriented magnetic poles, with inclinations indicating their formation $18^\circ \pm 4^\circ$ from the Equator, as does the Wrangellia terrane of Vancouver Island (Hillhouse and others, 1982). These authors considered this latitude to be significantly different

from the latitudes of adjacent cratonic rocks, which were $23^\circ \pm 6^\circ$ from the equator in Permian-Triassic time. This strengthens the association of the Seven Devils rocks with the Wrangellia terrane, as first proposed by Jones and others (1977), despite the typical tholeiitic nature of the Karmutsen Formation (Wrangellia Terrane) of Vancouver Island and the typical island-arc calc-alkaline petrochemistry of the age-equivalent Seven Devils Group (Scheffler, 1983; Sarewitz, 1983). Certainly the rocks of the Seven Devils Group must have been rotated around a vertical axis before the Late Jurassic, or translocated about 40° of latitude from south of the equator since their formation.

Post-Jurassic clockwise rotation of parts of the Blue Mountains province of northeastern Oregon (Wilson and Cox, 1980; Hillhouse and others, 1982) is of the same magnitude as that implied by the structural trend of the area, the southern limb of the "Columbia Arc" (Wise, 1963; Hamilton, 1978). The rotation was probably associated with development of the Columbia arc, and both events may be related to asymmetric subduction along the suture zone to the east. The displacement of the suture zone may then represent transform activity during subduction. The NNW-trending dikes that supplied lavas for the Columbia River Basalt Group are concentrated in the thinner, accreted crust west of the suture zone, close to the apex of the Columbia arc.

COLUMBIA RIVER BASALT GROUP

Stratigraphic Techniques

Knowledge of the stratigraphy of the Columbia River Basalt Group (CRBG) has greatly increased in the last 30 years. Useful summaries of this progress

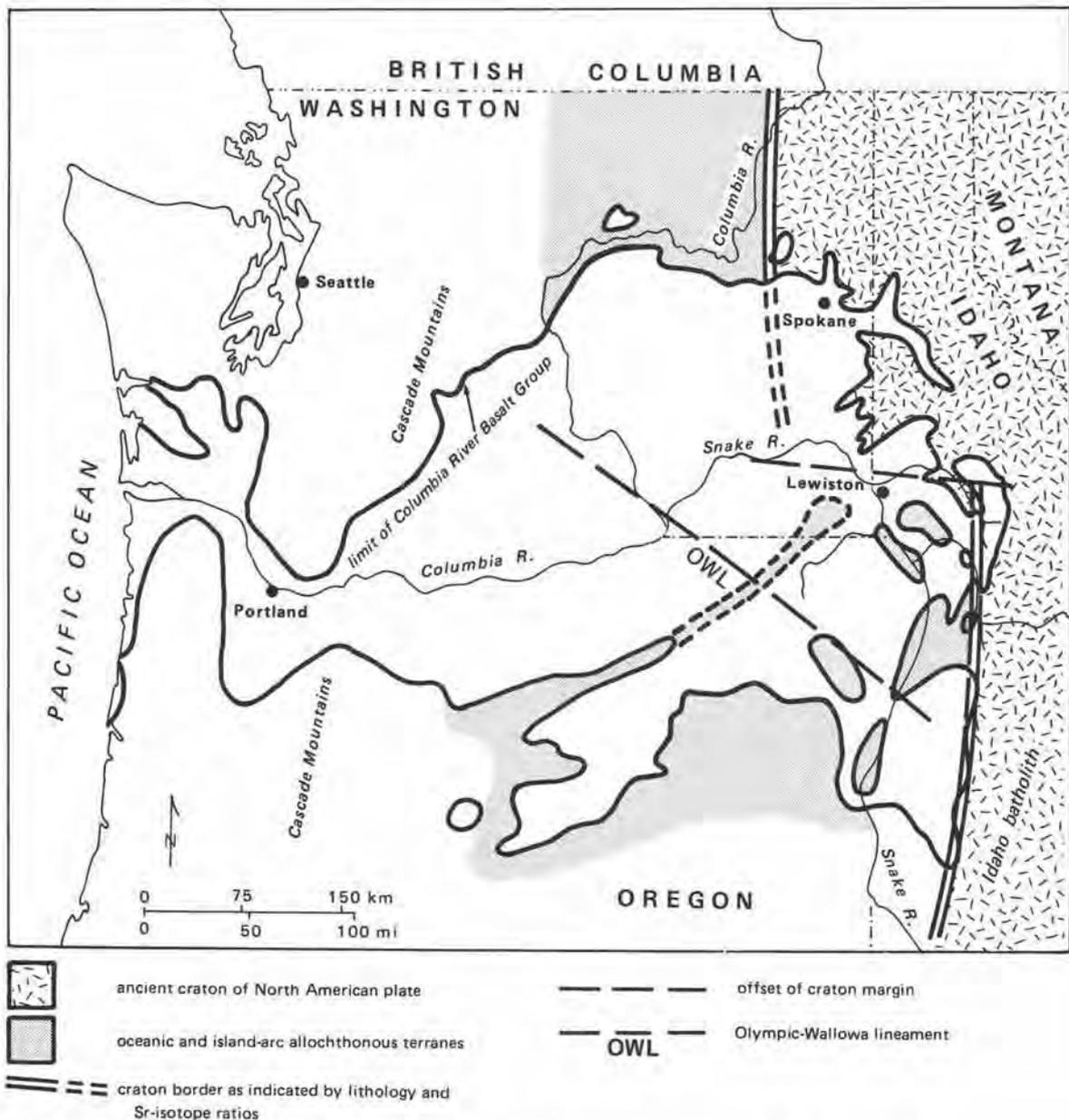


Figure 3.—Major pre-Cenozoic units of the eastern Columbia Plateau area.

are given by Waters (1961), Wright and others (1973), and Swanson and others (1979c). Our knowledge is summarized by Figure 4 and in the flow-distribution maps of Anderson and others (this volume).

Early attempts to correlate flows and establish a stratigraphy by field recognition alone had local significance but failed to provide a reliable stratigraphy for the whole plateau. Chemical analyses, used to fingerprint flows, added a new measure of confidence to flow correlation (Waters, 1961; Schmincke, 1967). Rapid and precise instrumental techniques for obtaining

chemical compositions have since been used to establish an increasingly detailed flow stratigraphy across the entire plateau (Wright and others, 1973; Holden and Hooper, 1976; Hooper, 1981; Swanson and others, 1979c; Swanson and Wright, 1981; Reidel, 1983).

Measurement of magnetic polarity has also proved critical in establishing a stratigraphy, particularly in the Grande Ronde Basalt. This approach was pioneered on the plateau by Campbell and Runcorn (1956) and was used with K-Ar dating to make regional correlations by Watkins and Baksi (1974). The development by Doell and Cox (1962) of a small,

| Epoch | Group | Formation* | Age (m. y.) | Member** | Flows | Magnetic Polarity*** | Some Approximate Correlations | |
|----------------------------|-----------------------------|---|-------------------|-------------------|-------------------|----------------------|--|-----|
| Early to Late Miocene | Columbia River Basalt Group | Saddle Mountains Basalt (<1%) | 6 | Lower Monumental | | N | Craigmont } Grangeville } Windy Ridge } Icicle Flat Member } Eden flow } | |
| | | | | | | Tammany Creek | | |
| | | | 8.5 | Ice Harbor | | Goose Island | | N |
| | | | | | | Martindale | | R |
| | | | | | | Basin City | | N |
| | | | | Buford | | R | | |
| | | | 10.5 | Elephant Mountain | | | | N-T |
| | | | | | | Swamp Creek | | |
| | | | 12 | Pomona | | | | R |
| | | | | | | Esquatzel | | N |
| | | | Weissenfels Ridge | | Slippery Cr. | N | | |
| | | | | | Tenmile Cr. | N | | |
| | | | | | Lewiston Orchards | N | | |
| | | | | | Cloverland | N | | |
| | | | Asotin | | N | | | |
| | | | Wilbur Creek | Lapwai | N | | | |
| | | Umatilla | | Sillusi | N | | | |
| | | | | Umatilla | N | | | |
| | | Wanapum Basalt (5-10%) | Priest Rapids | | Lolo | R | | |
| | | | | | Rosalia | R | | |
| | | | Roza | | T | | | |
| | | | Frenchman Springs | | N | | | |
| | | Eckler Mountain | | Lookingglass | N | | | |
| | Dodge | | N | | | | | |
| | Robinette Mountain | | N | | | | | |
| Grande Ronde Basalt (>85%) | | | | N ₂ | | | | |
| | | | | R ₂ | | | | |
| | | | | N ₁ | | | | |
| | | | | R ₁ | | | | |
| Imnaha Basalt (5-10%) | | Rock Creek and American Bar types interleaved | | | | | | |

*Showing approximate volume as percent of Columbia River Basalt Group.

**Informal magnetopolarity units in Grande Ronde Basalt.

***Magnetic Polarity (N = normal; R = reversed; T = transitional).

Figure 4.—Stratigraphy of the Columbia River Basalt Group of the eastern Columbia Plateau.

portable, fluxgate magnetometer capable of determining the polarity of flows in the field proved particularly useful in the Grande Ronde Basalt. Forming more than 85 percent of the CRBG by volume, these flows are fine grained and aphyric or only sparsely phytic, and show only minor chemical variation. Few flows possess a unique chemical composition (Reidel, 1983), but most have well-developed oxidized flow tops that

provide a clear and reliable measure of their magnetic polarity. Four magnetopolarity units (R₁, N₁, R₂, N₂) are recognized in the Grande Ronde Basalt (Fig. 4) and have been mapped across the plateau; their critical sections have been checked by more detailed paleomagnetic techniques (Choiniere and Swanson, 1979; Hooper and others, 1979; Swanson and others, 1979c; Anderson and others, this volume).

Linear Fissure Feeder System

All flows of the Columbia River Basalt Group, so far as we know, were fed by linear fissure systems (Fig. 4). The vertical dikes that represent feeders for these systems have a remarkably consistent NNW orientation, from the oldest Imnaha Basalt dikes in the southeast corner of the plateau to the youngest known dikes which fed the Ice Harbor Member almost 9 m.y. later. Small spatter and pumice cones and ramparts, cut in places by irregular thin dikes, mark local vents along the fissures (Swanson and others, 1975). The well-documented Roza Member (Fig. 4) (and, by inference, other flows) was formed by eruptions of large volumes of fluid, chemically homogeneous lava along the whole length of its fissure system (about 175 km) in a short time, probably in the order of one to two weeks (Shaw and Swanson, 1970; Swanson and others, 1975).

Exposed feeder dikes are present across the eastern half of the Columbia Plateau, but the greatest concentrations occur in the Cornucopia and Grande Ronde dike swarms of the southeast (Fig. 5; Waters, 1961; Gibson, 1969; Price, 1977; Anderson and others, this volume; Swanson and others, 1979a,b,c, 1980, 1981), together termed the Chief Joseph dike swarm by Taubeneck (1970). The Cornucopia swarm, southeast of the Wallowa Mountains (Fig. 5), comprises Imnaha and Grande Ronde dikes. The Grande Ronde swarm, cutting the eastern end of the Washington-Oregon border, comprises Grande Ronde and younger dikes. The Picture Gorge Basalt was fed by the discrete Monument dike swarm, and both dikes and flows of the Picture Gorge are confined to the John Day basin (Waters, 1961; Fruchter and Baldwin, 1975).

Whether the Columbia River Basalt Group dikes as now exposed represent the original distribution of fissures, or merely reflect the greater depth of erosion of the uplifted southeast part of the province, remains unclear. Certainly, any dikes older than Grande Ronde unit R_2 (Fig. 4) in the Pasco basin would now be covered. No dikes are known in the pre-Columbia River Basalt Group rocks on the northern and southern margins of the plateau, so if dikes are present in the western part of the plateau, they are confined within the borders of the plateau.

The Grande Ronde dike swarm is aligned roughly along the projected northward trend of the Cornucopia swarm, but in detail it is offset by about 20 km eastward (Fig. 5), assuming that the two swarms were once directly along trend. Hooper (1982) suggested that this offset might have resulted from right-lateral strike-slip displacement parallel to the eastern extension of the Olympic-Wallowa lineament (Fig. 3).

Smaller scale faulting of this type is observed in the basalt near Riggins, Idaho; furthermore, recent paleomagnetic work has demonstrated local 17° clockwise rotation of Imnaha dikes and flows south of the Wallowa-Seven Devils divide and 30° clockwise rotation in the southeast corner of the Weiser basin (Basham, 1978; Martin, 1984). One explanation for such rotation is strike-slip deformation on the ball-bearing pattern suggested by Beck (1976). Alternatively, the Grande Ronde and Cornucopia dike swarms may have been displaced at inception by an older structural discontinuity.

Whether displaced or not, a single, broad, vertical, NNW-trending fissure system underlies the eastern half of the plateau, with the greatest concentration of dikes lying in a much narrower zone just west of the Idaho border.

Imnaha Basalt was erupted from the south end of the fissure system, and Grande Ronde and younger basalt units generally were erupted farther north along the system. Rather than invoking a northward migration of the basalt source or of the magma reservoir at depth, Hooper and Camp (1981) ascribed this change to the northwesterly tilting of this part of the plateau, causing the intersection of the fissures with the topographically lowest part of the surface to move northward through time.

Basalt units confined to the southern margin of the Columbia Plateau—the basalt of Prineville, the Picture Gorge Basalt, and the basalt of Powder River—have separate feeder systems. The Picture Gorge dikes (Monument dike swarm) are essentially confined to the John Day basin (Fruchter and Baldwin, 1975) and are not associated with dikes of other basalt units. Feeders for the Prineville flows have not been found, but they must lie close to the Cascade Range (Fig. 5). The basalt of Powder River forms shallow cones apparently associated with marginal faults of the La Grande graben. In each case, the discrete feeder system presumably implies a discrete source, and we consider them separately from the main series of basalt flows.

History of Eruptions

Eruptions of the Columbia River Basalt Group began about 17 m.y. ago (McKee and others, 1981) in the southeast corner of the plateau; this activity is represented by the coarsely phyrlic (plagioclase + olivine ± augite) Imnaha Basalt (Fig. 4). Thick lava flows filled deep valleys north and south of the Seven Devils-Wallowa Mountains divide (Fig. 5) to a depth of nearly 700 m, with only the tops of the original mountains remaining above the lava plain (Anderson and

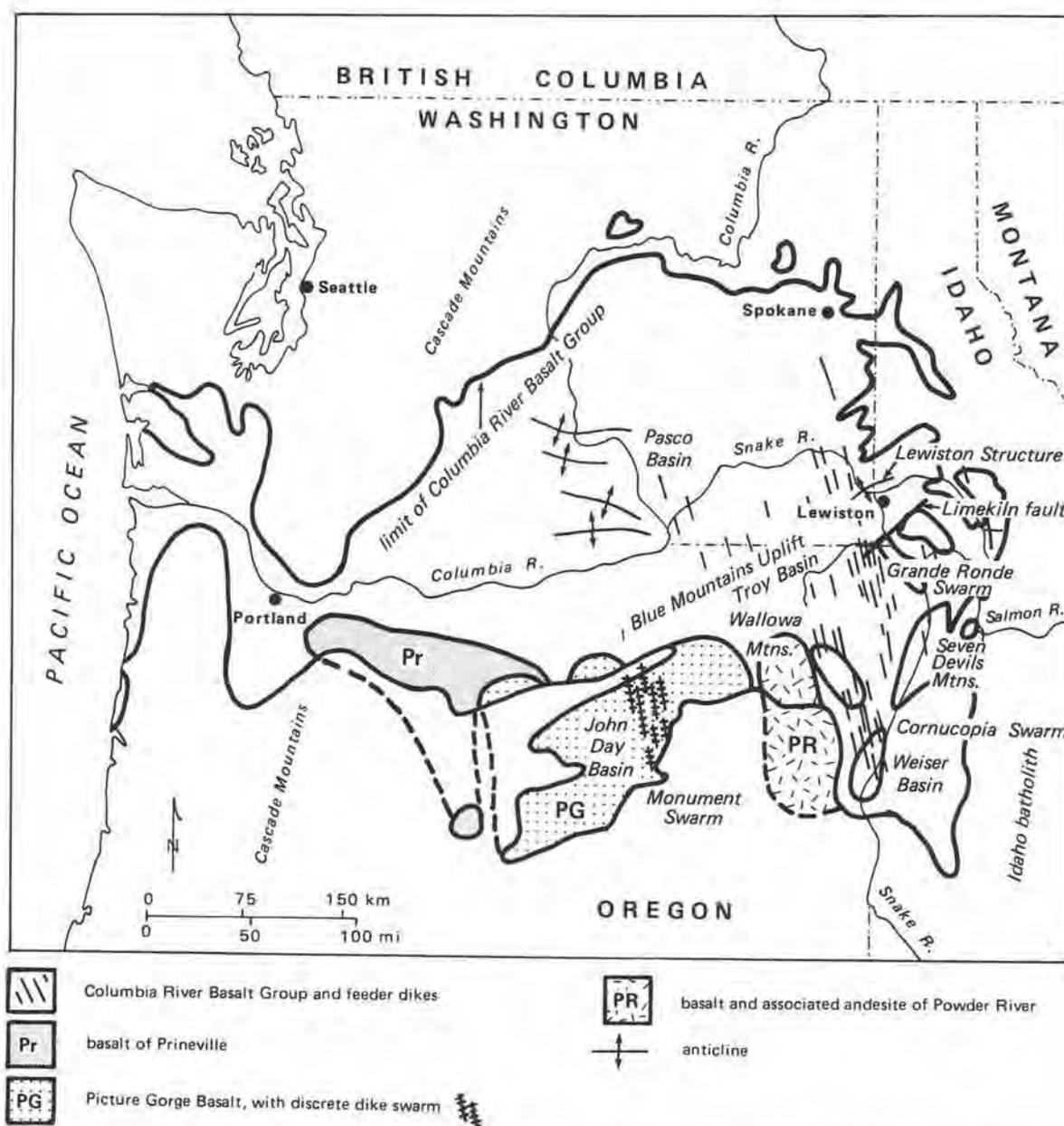


Figure 5.—Areal extent of the Columbia River Basalt Group and related basalt units in Washington, Oregon, and Idaho. Dikes and anticlines are shown diagrammatically, indicating general geographic distribution, relative concentration, and orientation. PG, Picture Gorge Basalt, with discrete dike swarm; Pr, basalts of Prineville; PR, basalts and associated andesites of Powder River.

others, this volume). All known Imnaha dikes and vents occur south of the Seven Devils-Wallowa divide, except for a few in the canyons of the Snake and Imnaha Rivers. The formation is thickest in the Weiser basin (Fitzgerald, 1982), thinning northward as far as the Lewiston basin, where 100 m are exposed in the uplifted block of the Lewiston structure (Fig. 5; Camp, 1976; Hooper and others, 1985). The extent of the Imnaha Basalt to the north and west is unknown. Basalt of Imnaha chemistry was not penetrated by a

3,250-m drill hole on the west side of the Pasco basin (Reidel and others, 1982), and neither Imnaha flows nor dikes have been observed around the north, west, or southwest margins of the plateau.

Grande Ronde Basalt succeeded Imnaha Basalt during the transition from normal to reversed polarity (Hooper and others, 1979). The fine-grained, generally aphyric Grande Ronde Basalt is more silicic than Imnaha Basalt, its quartz-tholeiite composition showing relatively little variation throughout the period of

eruption (~16.5-14.5 m.y. ago). The rate of magma supply, already high for the Imnaha Basalt, probably reached a maximum during this period ($40-60 \times 10^3 \text{ km}^3/\text{m.y.}$), and the Grande Ronde Basalt is by far the most voluminous formation (>85% by volume) of the group.

The oldest unit of the Grande Ronde flows (R_1 magnetopolarity unit; Fig. 4) is thickest (~400 m) immediately adjacent to the Seven Devils-Wallowa divide and thins to less than 200 m where it dips beneath younger flows a few kilometers northwest of Lewiston (Camp, 1976; Camp and Hooper, 1981; Hooper and others, 1985; Anderson and others, this volume). A small area of R_1 has been tentatively recognized in the Columbia Gorge (J. L. Anderson, personal commun., 1983), so some of these early Grande Ronde flows may have advanced across the Pasco basin (Anderson and others, this volume).

The younger units of the Grande Ronde Basalt (N_1 , R_2 , N_2 magnetopolarity units; Fig. 4) all thicken northwestward in the eastern part of the Columbia Plateau and lap out against older rocks to the southeast (Camp and Hooper, 1981; Anderson and others, this volume). The distribution and thickness of these magnetopolarity units illustrate the complex tectonic evolution of the southeast part of the plateau. While the entire area continued to tilt northwestward, individual blocks with granitic cores—the Seven Devils Mountains, the Wallowa Mountains, and, later, the Nez Perce Plateau and the northeastern extension of the Blue Mountains uplift—rose at varying rates. In contrast, the north-trending elongated, synformal graben along the ancient suture zone west of the Idaho batholith continued to sink relative to the general elevation of the area, so that basalt flows partly filled basins at Stites and Riggins in Idaho (Camp, 1981; Hooper, 1982).

Toward the end of R_2 time, the uplift of the southeast part of the plateau was reflected in the relatively sudden development of the Limekiln fault. This northeast-trending normal fault crosses the Snake River near the mouth of the Grande Ronde River (Fig. 5) and is downthrown a maximum of 620 m on the northwest (Reidel, 1978). The fault changes into monoclinical flexures in both directions (Camp and Hooper, 1981). Flows of unit N_2 are confined to the northwest of the fault (Anderson and others, this volume), but a line of N_2 cones without associated N_2 flows lies along the trend of the Grande Ronde dike swarm on the upthrown side of the fault (Anderson and others, this volume). Camp and Hooper (1981) suggested that the N_2 fissures had not moved northward; instead, only the point along the fissures from which magma emerged had moved north as the topo-

graphic low had moved north, due largely to development of the Limekiln fault. Both R_2 and N_2 flows crossed the western Columbia Plateau and advanced through the ancestral Cascade Range. At least one R_2 flow reached a point near Cathlamet, Washington (Wells and others, 1983), and N_2 flows extended to the Pacific Ocean (Anderson and others, this volume).

During Grande Ronde time, the southwest end of the Blue Mountains uplift of north-central Oregon (Fig. 5) formed an effective barrier between Grande Ronde Basalt, erupted east and north of the uplift, and Picture Gorge Basalt, erupted from the Monument dike swarm, mostly in the John Day basin south of the uplift. The older flows of the Picture Gorge Basalt have normal magnetic polarity, and the younger flows have reversed polarity. Younger Picture Gorge flows with reversed polarity locally interfinger with Grande Ronde R_2 flows across a low saddle in the Blue Mountains uplift (Cockerham and Bentley, 1973; Cockerham, 1975; Nathan and Fruchter, 1974), so the Picture Gorge Basalt appears to be contemporaneous with Grande Ronde N_1 and R_2 units. The Picture Gorge Basalt is petrographically similar to the Imnaha Basalt, but small differences in major- and trace-element abundances are evident (Nathan and Fruchter, 1974; McDougall, 1976). The basalt of Prineville (Uppuluri, 1974) was also erupted during Grande Ronde time (N_2 , and probably R_2) east of the ancestral Cascade Range and south of the Columbia Gorge.

The end of Grande Ronde time marked a brief hiatus in volcanic activity over most of the Columbia Plateau; the Vantage Member of the Ellensburg Formation, a thin sedimentary unit that covers much of the western part of the plateau, was deposited during this time. Farther east, a soil horizon developed at the top of the Grande Ronde Basalt (Camp, 1976, 1981; Swanson and Wright, 1976). Apparent interbedding of Wanapum and Grande Ronde flows in a small area just south of Pomeroy, Washington (Swanson and others, 1979c; Anderson and others, this volume), has recently been reinterpreted as due to faulting, following more detailed mapping (Beeson and others, 1985). Elsewhere, chemically diverse flows of the Eckler Mountain Member (Fig. 4) occur at the base of the Wanapum Basalt; this member contains one of the most primitive flows on the plateau (Robinette Mountain flow), as well as one of the most evolved (Shumaker Creek flow).

The Dodge flows (Eckler Mountain Member) are more widespread than either the Robinette Mountain or the Shumaker Creek flows. They are coarsely phyrlic (plagioclase and olivine) and form a useful marker unit in the northeastern Blue Mountains area. Several

dikes with Dodge chemistry and mineralogy occur over a wide zone in northeastern Oregon (Swanson and others, 1981). At least one flow of the Dodge was erupted from an independent vent in the valley of St. Maries River in northern Idaho (Swanson and others, 1979b; Anderson and others, this volume). The age of that vent is constrained only between N_2 and the Priest Rapids Member (Fig. 4), but the normal polarity of the flow suggests that it is essentially coeval with the Dodge flows in the Blue Mountains, about 80 km to the southwest.

Above the Eckler Mountain Member, the Wanapum Basalt is represented by three members of considerable volume, Frenchman Springs (normal magnetic polarity), Roza (transitional to reversed polarity), and Priest Rapids (reversed), each composed of two or more flows in some localities. The Frenchman Springs Member (more than five flows) and Roza Members (two flows) are chemically similar, although distinguishable in the field by the size, abundance, and form of their phenocrysts (Swanson and others, 1979c; Beeson and others, 1985). The Priest Rapids Member includes two flows of distinct chemical composition erupted sequentially (Swanson and others, 1979c).

The source dikes for the Frenchman Springs Member lie northwest of the Blue Mountains uplift, but a single flow of the member extends almost as far south as La Grande, Oregon. Five flows have now been formally recognized (Beeson and others, 1985). The lavas flowed north, west, and in places east, covering the greater part of the Columbia Plateau, and they advanced through the ancestral Cascades to the Pacific Ocean. This distribution (Anderson and others, this volume) implies that the plateau formed a gentle basin at that time, bordered on the south by a very low anticlinal rise (Blue Mountains) and on the southeast by somewhat higher elevations.

The Roza Member is one of the most voluminous (1,500 km³) and best known members on the plateau. Its fissure system and associated vents lie in the center of the main Grande Ronde dike swarm (Shaw and Swanson, 1970; Swanson and others, 1975, 1979c). The length of its fissure system (175 km), the huge volume of a single eruption (>700 km³), the homogeneity of the erupted magma, and the high rate of eruption provide critical constraints to physical models for eruptions of the basalt. The distribution of the Roza Member implies a northeast extension of the Blue Mountains uplift toward Lewiston (Anderson and others, this volume). As with Grande Ronde N_2 fissures, Roza fissures extended southeastward to the upthrown side of the Limekiln fault, but, possibly because of the greater elevation of the upthrown side, only dikes and small vents were developed. The two

major eruptions that supplied the principal lava flows occurred where the fissures intersected the lower topography on the downthrown side of the fault (Anderson and others, this volume).

The last great eruptions of the Wanapum Basalt produced the Priest Rapids Member. Its main source was on the eastern extremity of the Columbia Plateau in central Idaho (Camp, 1981; Anderson and others, this volume). A northern extension of this fissure system may have erupted in the St. Joe and St. Maries valleys (Swanson and others, 1979c; Camp, 1981). Eruptions apparently occurred primarily in the topographic low of the Clearwater River drainage. The lava flows filled the Stites basin and moved west to the incipient Lewiston basin, where some of them apparently encountered lake water and formed pillows. From there, the flows spread across the northern Columbia Plateau (bounded on the south by the rising Blue Mountains uplift) and extended through what is today the Columbia Gorge (Tolan, 1982). The Spokane-Cheney area was flooded by flows apparently erupted in the St. Joe valley or St. Maries valley.

Smaller flows of Wanapum age occur south of the Blue Mountains uplift (Anderson and others, this volume). The Powatka flow (Walker, 1973; Ross, 1978; Swanson and others, 1981) is fairly extensive and appears to outline a very shallow, incipient Troy basin (Fig. 5).

By the end of Wanapum time the northeast end of the Blue Mountains uplift was rising, flanked on either side by shallow synclinal basins at Lewiston and Troy. These structures continued to develop actively during Saddle Mountains time and after the basalt eruptions ceased. Late in Saddle Mountains time the Lewiston basin was greatly deepened by faulting along its northern margin.

The Saddle Mountains Basalt, forming less than 1 percent by volume of the Columbia River Basalt Group, was erupted over a long time. Consequently, the flows rest unconformably on a deformed and eroded surface. They include the intracanyon flows of Luper and Warren (1942) and are generally confined to, and hence preserve, the outlines of structural and erosional lows. The flows show even greater chemical and mineralogical variation than the Wanapum Basalt and are characterized by $^{87}\text{Sr}/^{86}\text{Sr}$ ratios significantly higher than those of any earlier flows (McDougall, 1976; Nelson, 1980; Carlson and others, 1981).

The Umatilla Member comprises two main flows, the Umatilla and Sillusi. A vent for the Sillusi occurs at Puffer Butte (Price, 1977), near the eastern end of the Oregon-Washington boundary; feeder dikes are found in this area and as far south as Oxbow Dam (Anderson and others, this volume). Both flows

entered the developing Lewiston basin and continued some distance northwest, but most of the lava moved west through the Troy basin syncline, following an older Grande Ronde drainage across the Blue Mountains uplift into the southern part of the Pasco basin (D. A. Swanson and T. L. Wright, unpub. data; Anderson and others, this volume). This distribution suggests that the major drainage of the southern part of the Columbia Plateau used the present lower Grande Ronde River and Wenaha River drainage in reverse. The lack of deep channels filled with basalt north of Lewiston suggests that little water occupied the lower Snake-Clearwater drainage. The ancestral Grande Ronde River drainage may have included the Salmon River, middle Snake River, and possibly even the upper Clearwater River, feeding into the Salmon River near Grangeville. Eruptions of the Umatilla Member may have played a major part in blocking this early drainage system.

The two succeeding Saddle Mountains members, Wilbur Creek and Asotin, together with a small flow of intermediate composition that lies between them (Lapwai flow of Camp, 1981), have no known feeder dikes. Camp (1981) suggested a source well up the North Fork of the Clearwater River for one or both of these flows. The Wilbur Creek Member may have another local source near Colton in southeastern Washington (Hooper and Webster, 1982). Both Wilbur Creek and Asotin flows filled the Lewiston basin, where, like older flows of the Umatilla Member, they encountered and often invaded wet sediment already filling the basin. From Lewiston they flowed northwest along a shallow drainage corresponding to the present Union Flat Creek, keeping north of the present Snake River and entering the Pasco basin in the vicinity of the Saddle Mountains (Swanson and others, 1979c; Anderson and others, this volume).

Three of the four flows of the Weissenfels Ridge Member are confined to the Lewiston basin. At least two of these flows have local feeder dikes within the basin (Hooper and others, 1985), and dikes chemically similar to the other two flows lie either within or just along the southern margin of the basin (Swanson and others, 1979c). The Lewiston Orchards flow appears to be more extensive than the other three. It is compositionally indistinguishable from the basalt of Sprague Lake (Swanson and others, 1979b,c) and from a dike, vent, and small flow just southwest of Riggins, Idaho (Hooper, 1982; Anderson and others, this volume). Traces of this flow also have been found north of the Lewiston basin near Colton, Washington, and Moscow, Idaho. The distribution of the Weissenfels Ridge Member again demonstrates the lack of major westward drainage from the Lewiston basin at that time.

An unnamed flow and basalt of the Esquatzel Member were the first to follow part of the present lower Snake River Canyon (Swanson and others, 1979c, 1980; Anderson and others, this volume), implying a major river draining west from the Lewiston basin. For the next 6 m.y., successive basalt flows partly filled this deep canyon only to be largely eroded by the river.

The Pomona Member (12 m.y. old; McKee and others, 1977) is the most voluminous of these flows. A possible feeder dike lies in the valley of the Clearwater River in central Idaho (Camp, 1981). The flow can be traced down the Clearwater River and lower Snake River (Swanson and others, 1979a, 1980) to the Pasco basin, where it spread out before following the southwest-flowing ancestral Columbia River out of the basin into the lower Columbia Gorge (Anderson, 1980; Tolan, 1982) and onward to the Pacific Ocean, where it has been called the basalt of Packsack Lookout (Snively and others, 1973; Anderson and others, this volume)—a remarkable and well-documented journey of more than 600 km.

The succeeding Elephant Mountain Member (10.5 m.y. old; McKee and others, 1977) had its source in the Troy basin, on the south side of the Blue Mountains uplift (Walker, 1973; Ross, 1978; Swanson and others, 1979c). Because direct drainage to the west was then blocked by the Umatilla flow or by the Blue Mountains uplift, the basalt of the Elephant Mountain Member flowed northeastward from the Troy basin, down the southwest slope of the Lewiston basin, and into the Snake River Canyon (Swanson and others, 1980; Hooper and others, 1985). It then followed the Snake River into the Pasco basin, just as the Pomona had done. Isolated outcrops imply that the Elephant Mountain basalt reached as far west as The Dalles (Anderson and others, this volume).

Between the eruptions of the Weissenfels Ridge Member and that of the Pomona Member, deep canyons developed, similar to those of the present Snake River draining west from Lewiston. This was a major, relatively sudden change in the region's drainage. It appears to have been a consequence of the blocking by the Umatilla Member of the drainage up the Grande Ronde and Wenaha valleys, and the subsequent overflow of the dammed water across the shallow divide to the north, into the Lewiston basin. Rapid erosion of the new channel along the Limekiln fault no doubt contributed to the consolidation of the new drainage pattern in which the whole Salmon, Clearwater, and Grande Ronde drainage was funneled for the first time through the Lewiston basin.

At this time (12 to 10 m.y. ago) the Lewiston basin remained a shallow east-west syncline. Continued

NNW-SSE compression, however, resulted in the northern limb of the syncline breaking to form the Lewiston structure (Camp, 1976) in which an anticlinal block is thrust upward to steepen the northern edge of the Lewiston basin. This probably occurred between eruption of the Elephant Mountain Member (10.5 m.y.) and eruption of the Lower Monumental Member (6 m.y.).

Several less voluminous basalt eruptions occurred during Saddle Mountains time. Only two of these have been dated, and the stratigraphic position of the others is poorly constrained because of their local occurrence. The most important of these are the Buford Member (Walker, 1973), the Ice Harbor Member (8.5 m.y. old; Swanson and others, 1975; McKee and others, 1977), and the Lower Monumental Member, which—at 6 m.y. old—is the youngest flow known on the Columbia Plateau (McKee and others, 1977; Swanson and others, 1979c). The distribution of these flows is shown in Anderson and others (this volume).

TECTONIC EVOLUTION OF THE EASTERN COLUMBIA PLATEAU

Each flow faithfully sought out the lowest topography. The extent and the variable thickness of flows thus provide a graphic illustration of the evolving plateau surface during eruptions of the basalt. From this we can develop a well-constrained structural model. Much of the available evidence may be found in Camp (1976), Swanson and Wright (1976), Vallier and Hooper (1976), Ross (1978), Reidel (1978), Camp (1981), and Camp and Hooper (1981) and is summarized by Hooper and Camp (1981).

Three tectonic factors appear to be responsible for the structural evolution:

- (1) The vertical rise of the southeast corner of the Columbia Plateau relative to the main part of the plateau and the Pasco basin;
- (2) A stress regime with σ^1 NNW (horizontal), and with σ^2 and σ^3 interchangeably ENE (horizontal) and vertical;
- (3) Use of an older fracture pattern in the basement rocks with northwest-, northeast-, and approximately north-trending, vertical planes of weakness to accommodate factors (1) and (2).

The vertical rise of the southeast part of the Columbia Plateau appears to be a continuation of the steady rise of the Idaho batholith and associated plutonic bodies during the early Tertiary (Axelrod, 1968). The rise presumably was isostatic; it continued after cessation of basalt eruptions 6 m.y. ago, and some movement on most of the observed faults postdates the youngest flows involved.

Rise of the southeast part of the Columbia Plateau relative to the central plateau was not uniform. The Seven Devils Mountains, Wallowa Mountains, Nez Perce Plateau, and Blue Mountains uplift southwest of Lewiston apparently rose at different rates. The Seven Devils and Wallowa blocks were already topographic highs before 17 m.y. ago, with a relief similar to that of today. Both blocks were almost covered by the early flows of Imnaha Basalt, which rapidly filled the intervening valleys. Although the rate of lava eruption increased in Grande Ronde time (Hooper, 1981), the thickness of the basalt pile increased less rapidly as flows spread widely over the lava plain and advanced downslope to the northwest. As a consequence of the less rapid buildup of lava in this area, the Seven Devils Mountains and Wallowa Mountains began to re-emerge. Successive units of Grande Ronde Basalt covered less and less of the rising blocks, each younger unit dipping off the blocks at a lower angle as the lower flows were tilted (Hooper and Camp, 1981).

The Nez Perce Plateau also remained above the Imnaha Basalt, but it was submerged by flows of the Grande Ronde Basalt (R_1 , N_1 , R_2). However, toward the end of R_2 time, the Nez Perce block broke along the northeast-trending Limekiln fault (Fig. 5) and rose 620 m relative to the downthrown northwest side of the fault. Grande Ronde N_2 flows are confined to the northwest side of the fault.

Behavior of the Blue Mountains uplift was more complex. Southwest of Walla Walla it acted as a barrier separating Grande Ronde Basalt to the north from most of the contemporaneous Picture Gorge Basalt to the south. Farther northeast, a low ridge may have been a barrier to Imnaha Basalt (Swanson and others, 1980; Camp and Hooper, 1981), but it was submerged by flows of the Grande Ronde Basalt, and no evidence of further upward movement is seen until Wanapum time (Ross, 1978; Swanson and Wright, 1983). Rise of the northeast part of the Blue Mountains uplift was accompanied by formation of complex series of folds and faults during late Wanapum and Saddle Mountains time, including synclines in the Lewiston and Troy basins (Ross, 1978; Swanson and others, 1980; Camp and Hooper, 1981; Swanson and Wright, 1983). The complexity of the Blue Mountains uplift suggests that it is not simply a product of isostatic movement, but formed in part by horizontal tectonic stresses.

Successively younger stratigraphic units in the Imnaha and Grande Ronde Basalts attain their maximum thickness progressively farther northwest. Wanapum and Saddle Mountains Basalts flowed onto an irregular topographic surface and were confined almost exclusively to the area northwest of the Lime-

kiln fault and associated monoclines, except in the topographic low of the upper Clearwater drainage. This distribution pattern is interpreted as the result of northwestward tilting of this part of the Columbia Plateau (Swanson and others, 1980; Camp, 1981; Hooper and Camp, 1981).

Specific evidence of horizontal compression is provided by the Lewiston structure (Camp, 1976; Camp and Hooper, 1981), which developed on the north side of the Lewiston basin syncline during the latter half of Saddle Mountains time (Fig. 5). The predominantly east-west structure comprises an uplifted anticlinal block or wedge bounded on its north side by a reverse fault. The uplift parallels an older structural ridge to the north (Bald Butte-Granite Point ridge; Hooper and Rosenberg, 1970), along the trace of the old cratonic margin. It seems probable that this early zone of weakness in the basement was a factor in the location of the Lewiston structure. The anticline and associated reverse fault suggest north-south compression, similar to that responsible for folds in the western part of the plateau (Price, 1982).

Perhaps the most significant structural feature in the eastern Columbia Plateau is the dike system itself. The remarkably consistent orientation of these dikes during an 11-m.y. period not only testifies to the tectonic origin of the original fissures, but also specifically implies a regional stress regime at the base of the crust with σ^3 ENE and horizontal, and σ^1 approximately equal to σ^2 throughout those 11 m.y.

Finally, local strike-slip displacement along northwest-trending (dextral), north-trending (dextral), and northeast-trending (sinistral) faults has been reported (Ross, 1978; Shubat, 1979; Hooper and Camp, 1981; Swanson and Wright, 1983), the displacements being either directly observed or inferred from prominent horizontal slickensides (Fig. 6). Near Riggins, Idaho, such strike-slip movement apparently has led to clockwise rotation (Hooper, 1982). Evidence of clockwise rotation of the Imnaha Basalt along the southeast extension of the Olympic-Wallowa lineament in the Oxbow and Weiser areas was reported by Basham (1978) and confirmed by Martin (1984). No rotation of basalt flows is evident farther north.

A model regional stress regime explaining each of these parameters includes σ^1 oriented NNW and horizontal, σ^2 vertical, and σ^3 ENE and horizontal. Vertical stress increases with depth and the NNW-trending dikes would originally develop deep in the crust or upper mantle under a tensional regime with σ^1/σ^2 perhaps approaching unity. At lesser depths, σ^1 (NNW) would be greater than σ^2 (vertical), which in turn would be greater than σ^3 (ENE)—a configuration that would permit development of northwest-trending



Figure 6.—Horizontal slickensides in the Lewiston structure, southeastern Washington. The curved, slickensided face is more than 1 m long.

(dextral) and the northeast-trending (sinistral) strike-slip faults. Near the surface, the vertical direction would become σ^3 , which would encourage development of east-trending folds and reverse faults.

The little available evidence suggests that this regional stress regime has been in effect from the Eocene to the present. Both R. J. Kuhns (1980) and Youngs (1981) have mapped Challis dikes oriented northwest and northeast within the Idaho batholith. Some dikes postdate northwest-trending (dextral) and northeast-trending (sinistral) strike-slip displacements along these vertical planes, whereas earlier dikes are sheared and displaced. R. J. Kuhns and Youngs demonstrated that this probably resulted from σ^1 oriented approximately north-south and horizontal. In the Lewiston basin, small reverse faults parallel the major reverse fault and displace gravel younger than the 6-m.y.-old Lower Monument Member (M. J. P. Kuhns, 1980). Recent measurement of contemporary stress in the Pasco basin (Caggiano and others, 1983) indicates a regime with σ^1 north-south and horizontal.

The onset of flood-basalt volcanism in the southeast part of the Columbia Plateau may have reflected changes in plate motion off the Oregon and Washington coasts about 17 m.y. ago (Carlson, 1976). There is little evidence to suggest further changes in the stress regime in the eastern Columbia Plateau that might correspond to subsequent changes in plate motions (Barrash and Venkatakrishnan, 1982; Robyn and Hooper, 1982).

BASALT GENESIS

A number of physical constraints should be emphasized before discussing compositional variations of the Columbia River Basalt Group and models that attempt to account for those variations by such processes as partial melting, crystal fractionation, and crustal assimilation.

Individual flows were erupted in unusually large volumes in a very short time. The linear fissure systems were exceptionally long and retained a remarkably consistent trend for 11 m.y. For example, the fissure systems of six units for which good evidence is available are 70 km or more long (Frenchman Springs, Roza, Priest Rapids, Umatilla, Lewiston Orchards, and Ice Harbor units; Shaw and Swanson, 1970; Swanson and others, 1975; Camp, 1981; Hooper, unpub. data). The Roza fissure system (>175 km long) twice provided more than 700 km³ of basalt in a few days, with no discernable variation in composition. It appears inevitable that fissures more than 70 km long that provided homogeneous magma to the surface throughout their entire length would have extended to the base of the 25- to 30-km-thick crust (Hill, 1978).

The large volume of lava and its homogeneity require an equally large reservoir of well-mixed magma no higher than the base of the crust, and the rapid rise of magma from that reservoir to the surface. Any significant crustal assimilation would have involved only lower crustal material, which may be more mafic than upper crustal material, and must have been followed by very effective mixing (McDougall, 1976). That significant volumes of silicic upper-crust material were not assimilated by the basaltic magma is indicated by the rarity of crustal xenoliths in the many hundreds of dikes and flows examined in the field. All physical evidence argues against the possibilities of magma storage in the upper crust, upper-crust contamination, or crystal fractionation under near-surface conditions.

The amount of major-element data available for the Columbia River Basalt Group is now prodigious. While many of the data remain unpublished, useful lists are available in Brock and Grolier (1973), Hooper and others (1976b), Wright and others (1979, 1980,

1982), and Reidel and others (1981). Trace-element analyses have been published by Osawa and Goles (1970), Nathan and Fruchter (1974), McDougall (1976), Wright and others (1979, 1980, 1982), Reidel (1983), and Hooper and others (1984). Wright and others (1979, 1980, 1982) also include glass analyses. Averaged flow analyses were provided by Swanson and others (1979c), Wright and others (1979, 1980, 1982), Reidel (1983), and Hooper and others (1984).

As demonstrated many years ago by Washington (1922) and emphasized by subsequent workers (for example, Waters, 1961), the Columbia River Basalt Group is tholeiitic; the greatest volume is represented in the Grande Ronde Basalt, a quartz tholeiite with SiO₂ between 52 and 57 percent. The earliest flows (Imnaha Basalt) are coarsely phyrlic (plagioclase + olivine ± augite), with SiO₂ between 47 and 52 percent; these rocks are olivine tholeiites (transitional to alkali basalt) in which olivine continued to crystallize, accompanied by a single Ca-rich clinopyroxene and by considerable Fe enrichment in local segregation veins. However, the residual glass within the flows is consistently enriched in silica (Hooper, 1974; Hooper and others 1976a). Very similar mineralogy and chemical trends occur in the Picture Gorge Basalt (Lindsley, 1960; Lindsley and others, 1971).

The predominant quartz tholeiites of the Grande Ronde Basalt are almost entirely aphyric, but a few contain plagioclase phenocrysts and fewer still carry rare, partly resorbed phenocrysts of Al-poor orthopyroxene (Reidel, 1978, 1983). These rocks have less than 3 percent olivine, and they carry microphenocrysts of plagioclase, pigeonite, and augite (Reidel, 1983). Wanapum and Saddle Mountains flows have variable compositions within the basalt family. Most flows are phyrlic (plagioclase, olivine, augite), but the Umatilla flow, for example, is aphyric and exceptionally fine grained.

Like continental flood basalts of other provinces, the Columbia River Basalt Group is relatively "evolved", with low Mg/(Mg + Fe²⁺) values, low abundances of trace elements compatible with early crystallizing phases (Cr, Ni, Co), and relatively high abundances of the more incompatible elements (K, P, Ti, Ba, Sr, Rb, Zr, and rare-earth elements). Such patterns are inconsistent with a magma in equilibrium with a garnet lherzolite, and they imply either a mantle source of different composition (Wilkinson and Binns, 1977; Wright and Helz, 1981) or extensive fractionation of olivine and pyroxene from the magma after separation from its source (McDougall, 1976).

In the following discussion of chemical constraints on the origin and evolution of the group, the Imnaha Basalt, Grande Ronde Basalt, Wanapum Basalt, and

Saddle Mountains Basalt are referred to as the "main series" all derived from the same, admittedly very broad, Chief Joseph dike swarm (Grande Ronde and Cornucopia swarms combined) (Taubeneck, 1970). Other units of the Columbia River Basalt Group and contemporaneous units of its southern margin (Figs. 4 and 5) are omitted from this discussion because they have separate feeder systems and can be assumed to have separate sources.

Two principal types of chemical variation can be distinguished within the main series. First, sharp discontinuities in major-element composition are accompanied by significant changes in mineralogy. Most obvious are the changes in SiO_2 , K_2O , and associated trace-element contents (Rb, Ba). These discontinuities occur at stratigraphic breaks, such as those between the Imnaha Basalt and Grande Ronde Basalt, between the Grande Ronde Basalt and Wanapum Basalt, and between many individual flows in the Wanapum Basalt and Saddle Mountains Basalt (Fig. 4). This type of variation also separates the two contemporaneous types of Imnaha Basalt (Rock Creek and American Bar; Hooper and others, 1984).

The second type of chemical variation is the more gradational change within the larger stratigraphic units in the abundances of incompatible minor and trace elements (Ti, P, Zr, rare-earth elements). The abundances of these elements vary inversely with the abundances of the most compatible trace elements (Cr, Ni, Co) and with the $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ ratio (Hooper and others, 1984).

The two types of variation are largely independent. Two flows with similar concentrations of minor and trace elements may have very different concentrations of SiO_2 and K_2O , and different mineralogies. This decoupling of the two types of chemical variation cannot result from any combination of crystal fractionation and partial melting. Some other factor is required. Obvious candidates are a heterogeneous source (including a variable volatile component), crustal assimilation, and either crystal fractionation or partial melting at varying depths. None of these processes excludes any other, and any or all could have played some part.

Quantitative modeling of compositional variations between flows and groups of flows has achieved very limited success (Goles and Leeman, 1976; Helz, 1978; Wright and others, 1976; Swanson and Wright, 1981; Reidel, 1983; Hooper and others, 1984). With the use of major- and trace-element abundances and the observed phenocryst phases, it has not yet proved possible to derive one flow composition from another, or to adequately explain the larger compositional differences between groups of flows. Swanson and Wright

(1981) have shown (1) that variation between the bulk composition of a flow and its chilled margins is not significant, with the implication that either the phenocrysts are not intratelluric, or, if intratelluric, have not undergone substantial physical separation from the residual liquids; and (2) that chemical-variation trends between the bulk composition and the residual glass within a flow do not correspond to the chemical variation between two closely associated flows, which might otherwise have been linked by simple crystal fractionation. The coarsely phyrlic Imnaha Basalt contains intratelluric phenocrysts of plagioclase and olivine, with or without augite, but quantitative models of the fractionation of these phases again prove inadequate to derive one flow composition from another (Hooper and others, 1984).

Wright and others (1976) and Swanson and Wright (1981) concluded that crystal fractionation within the crust was not a significant process in the development of the Grande Ronde and younger formations of the Columbia River Basalt Group. They agree with Helz's (1978) conclusions, based on experimental studies, that the magmas represent separate melts of heterogeneous Fe-rich pyroxenites devoid of garnet. Modeling of Imnaha Basalt (Hooper and others, 1984), while in many ways consistent with the work of Swanson and Wright (1981), suggests that crystal fractionation and crustal contamination may have had significant roles, although inadequate by themselves to account for all the chemical variations. For example, the best crystal-fractionation mass-balance solutions to major-element variations show a close correlation between the proportions of phases subtracted from the parent and the proportions of phenocrysts observed in the rocks. Reidel (1983) reached a similar conclusion for the Grande Ronde Basalt. However, trace-element concentrations belie such a simple model for both Imnaha Basalt and Grande Ronde Basalt. Crustal assimilation in addition to crystal fractionation can reduce the trace-element discrepancies considerably, but the degree of required assimilation appears too great in the light of isotopic data (McDougall, 1976; Carlson and others, 1981). Furthermore, these models do not adequately account for the changes between the subtypes of Imnaha Basalt and between the Imnaha Basalt and the Grande Ronde Basalt (Hooper and others, 1984). Either recharge by a magma more "primitive" than any that reached the surface, or considerable variation in the composition of the assimilated crust, or a heterogeneous source appears to be required—in addition to crystal fractionation combined with lower-crust assimilation.

The degree of assimilation involved with the Columbia River Basalt Group has been discussed by

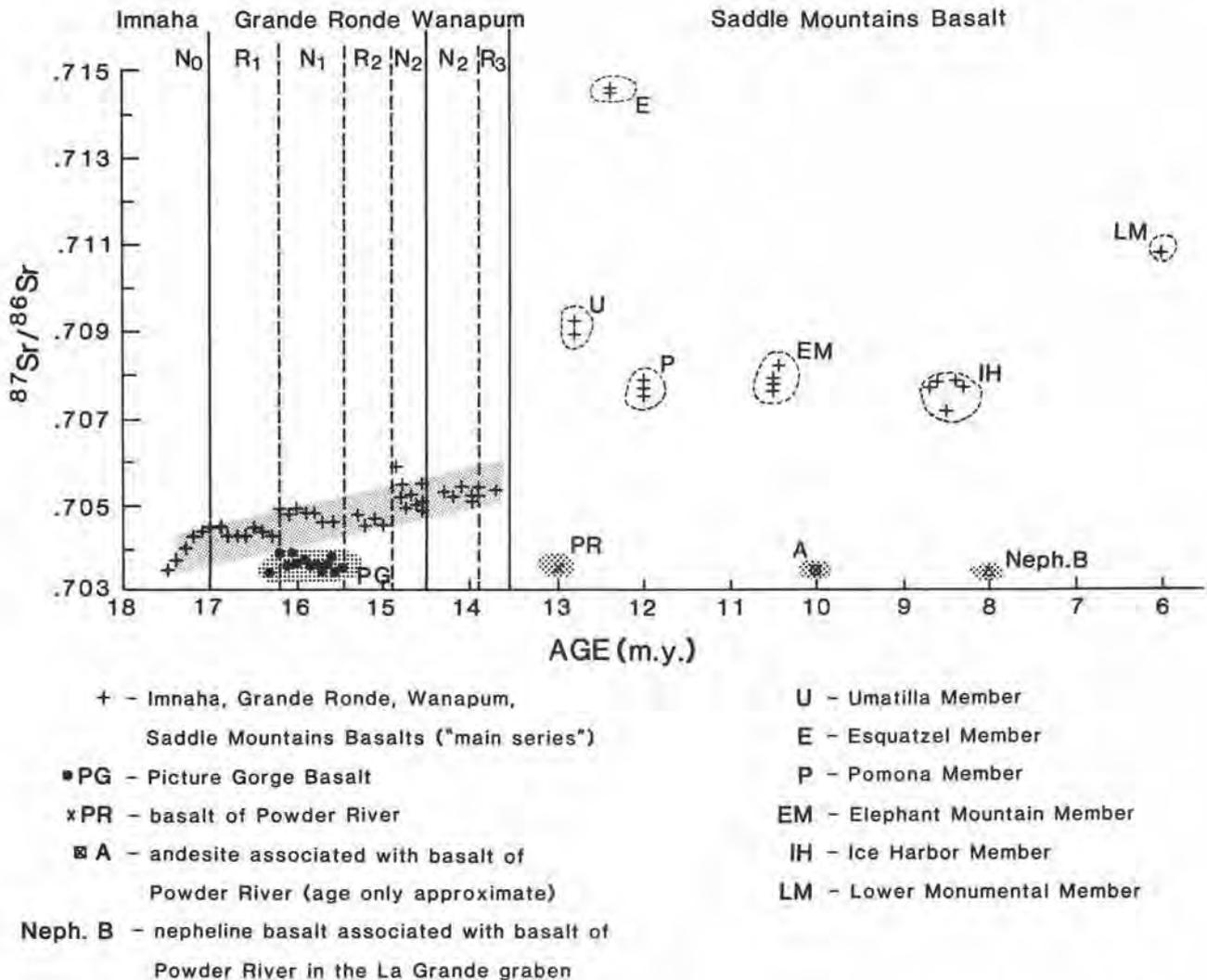


Figure 7.—Plot of $^{87}\text{Sr}/^{86}\text{Sr}$ versus age for Columbia River Basalt Group and some associated rocks. Age cannot be determined as accurately as suggested by the plotted points, but within each group the points have been spaced according to stratigraphic positions of analyzed samples.

McDougall (1976), Carlson and others (1981, 1983), Nelson (1983), and De Paolo (1983), who considered Sr, Nd, and O isotopic data in addition to trace-element abundances. Their conclusions vary from 20-percent crustal contamination (Carlson and others, 1981, 1983) to very little assimilation of crustal material for the main series of Imnaha Basalt, Grande Ronde Basalt, and Wanapum Basalt (McDougall, 1976; Nelson, 1978, 1980, 1983; Nelson and others, 1980; De Paolo, 1983). The high $^{87}\text{Sr}/^{86}\text{Sr}$ values for the Saddle Mountains Basalt (McDougall, 1976; Carlson and others, 1981; Nelson, 1983) were taken by DePaolo (1983) to indicate crustal contamination for this very small volume of magma, but Carlson and others (1981) and Nelson (1983) remained unsure.

Published Sr-isotope data are plotted against time in Figure 7, a plot made possible by increasingly

precise knowledge of the regional stratigraphic relations of the analyzed samples. From this we note the following:

- (1) The Sr-isotope ratio generally increases through time in the main series from Imnaha time to the end of Wanapum time;
- (2) In the Grande Ronde and Wanapum, Sr-isotope ratios remain about the same for significant periods of time; that is, all R_1 samples are about the same, as are all N_1 - R_2 samples and N_2 -Wanapum samples;
- (3) All sampled individual flows of the Saddle Mountains Basalt have Sr-isotope ratios significantly higher than those of the older formations, and these ratios show no correlation with time;

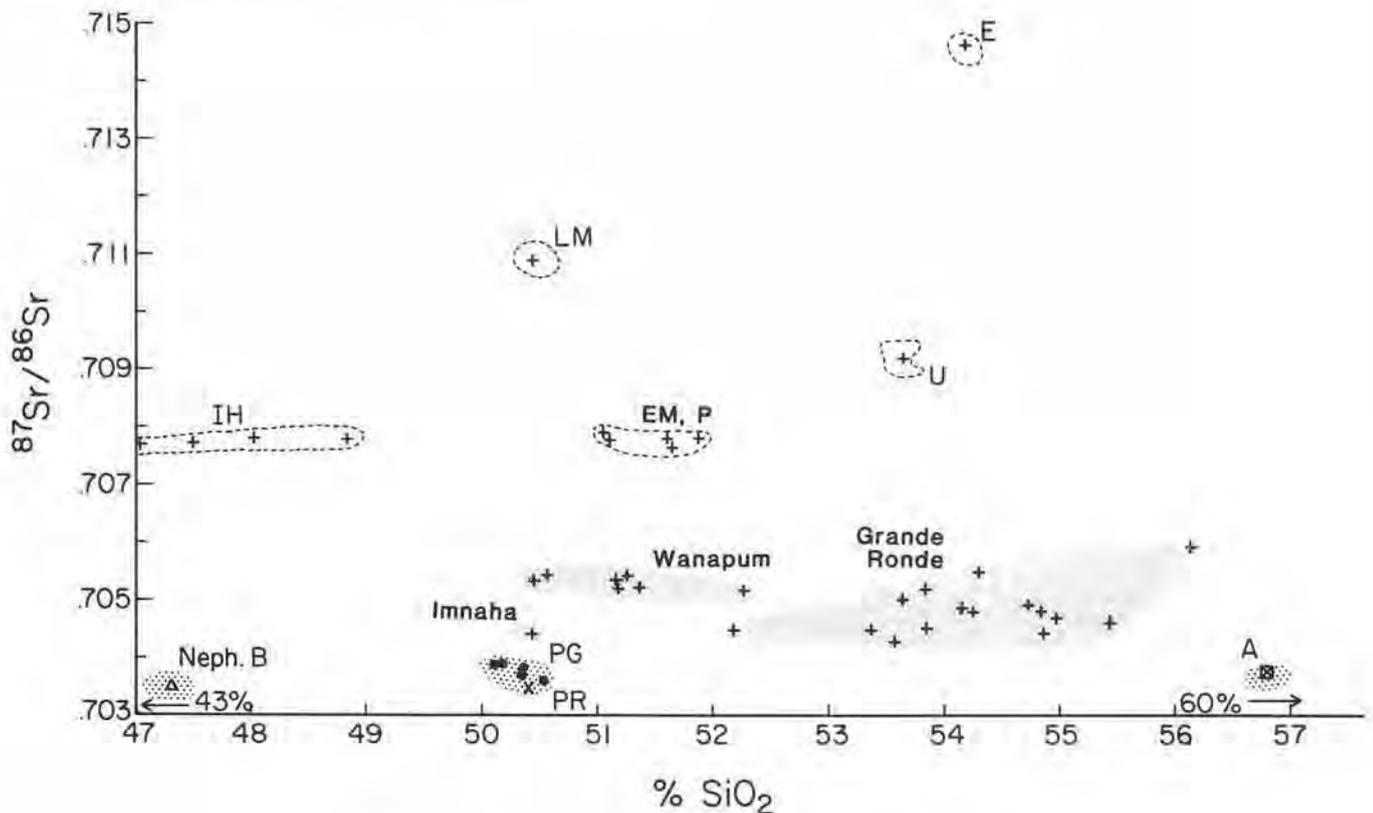


Figure 8.—Plot of $^{87}\text{Sr}/^{86}\text{Sr}$ versus SiO_2 (symbols as in Fig. 7).

- (4) The Picture Gorge Basalt, the basalt of Powder River, and the flows specifically associated with the La Grande graben (nepheline basalt and andesite; Carlson and others, 1981) all retain the same low Sr-isotope ratio despite an age range equivalent to that of the main series; and
- (5) Sr-isotope ratios obtained for the same flow by different laboratories display little variation, so that significant analytical error can be discounted.

A plot of Sr-isotope ratios against SiO_2 (Fig. 8) shows no correlation, and these ratios plotted against a relatively incompatible element such as phosphorus (P_2O_5 ; Fig. 9) show at best a poor positive correlation.

Many authors have suggested a correlation between isotopic composition and geographic location (Carlson and others, 1981; De Paolo, 1983) and implied that a laterally heterogeneous mantle is responsible in part for the variation in the Sr-isotope ratio. The location of the feeder system, not the location of the flows, is critical to such a discussion. We have noted some evidence for the point of eruption of flows moving northward through time along the trend

of the main dike swarm, but the field evidence makes it clear that this was an artifact of a progressively tilting topographic surface, and there is no good evidence to suggest that the fissure system, and hence the source, moved in such a manner.

There are many well-documented examples of flows of similar composition erupted from fissures separated widely from east to west across the fissure system. Such flows might be expected to have significantly different trace-element compositions and Sr-isotope ratios if they were derived from a heterogeneous mantle. This is exemplified by the Imnaha and Picture Gorge Basalts, whose feeder dikes are well documented (Anderson and others, this volume). Of slightly different age, the two formations were both formed by the first eruptions in their respective areas, some 200 km apart. They are similar in almost all aspects of field appearance, mineralogy, and composition. They can be separated by discriminant analyses using elemental abundances, but the differences are small. The two formations probably reflect slightly different source compositions, yet their Sr-isotope ratios are similar (Fig. 7).

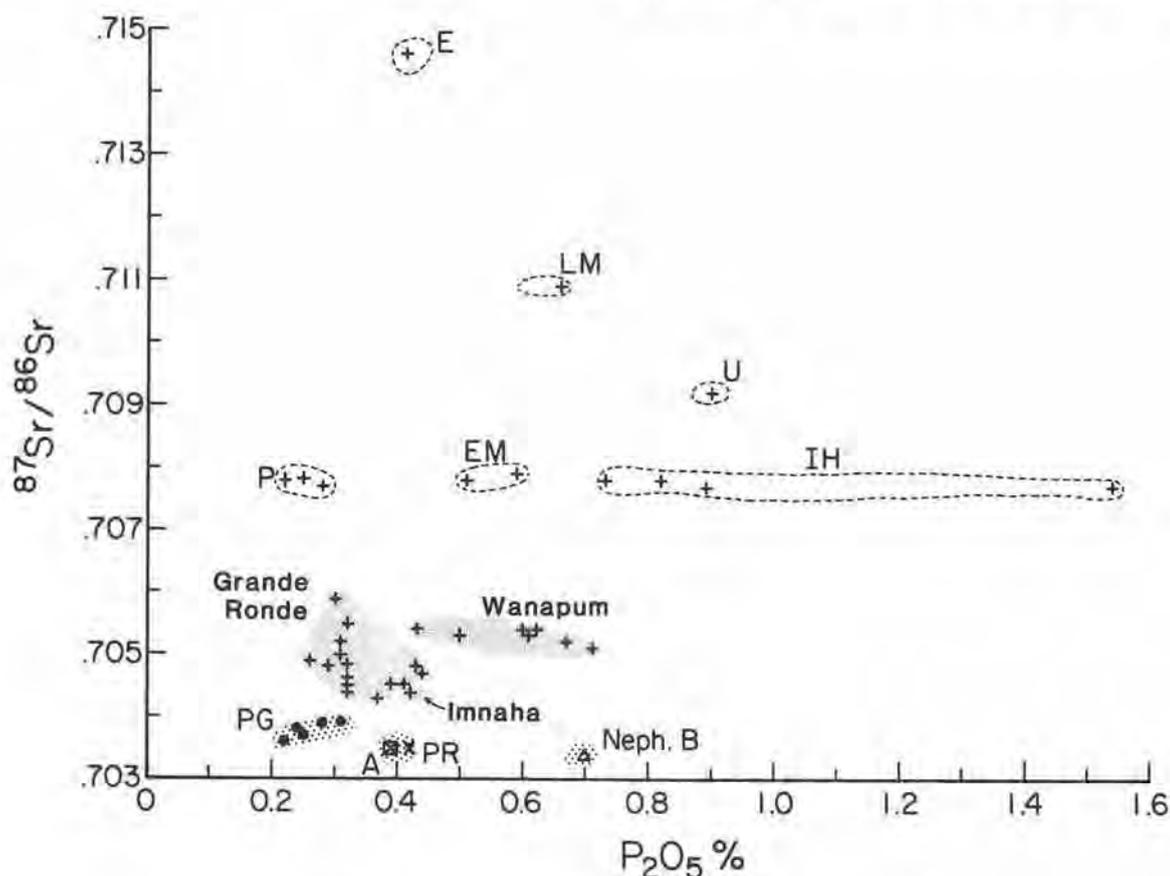


Figure 9.—Plot of $^{87}\text{Sr}/^{86}\text{Sr}$ versus P_2O_5 (symbols as in Fig. 7).

The three major members of the Wanapum Basalt provide another example. The Frenchman Springs, Roza, and Priest Rapids Members were erupted respectively from the west, middle, and east parts of the aggregate fissure system (Anderson and others, this volume), the distance across the system being more than 200 km. Whereas these flows are rather easily distinguished in the field by the different sizes and configurations of their phenocrysts, their chemical compositions are generally similar. Not even discriminant analysis adequately separates the Roza from the Frenchman Springs. Nor is there a significant difference in their Sr-isotope ratios (Fig. 7).

Within the Saddle Mountains Basalt, the Pomona, Elephant Mountain, and Ice Harbor Members were erupted from sources respectively in the east, middle, and west parts of the fissure system (Anderson and others, this volume). Chemically these three flows are indeed very different, yet their Sr-isotope ratios are similar (Fig. 7). We conclude that lateral mantle heterogeneity is not a significant factor in the variation of Sr-isotope ratios among flows on the Columbia Plateau.

In conclusion, an adequate model for the origin and evolution of the Columbia River basalts still eludes us. No single process is adequate. Crystal fractionation with assimilation in an "open reservoir" system, periodically recharged by more primitive magma from below, seems possible. Perhaps the most crucial question is the relative importance of crystal fractionation and assimilation on the one hand and the tapping of a vertically heterogeneous source on the other.

REFERENCES CITED

- Anderson, J. L., 1980, Pomona Member of the Columbia River Basalt Group—An intracanyon flow in the Columbia River Gorge: *Oregon Geology*, v. 42, no. 12, p. 195-199.
- Anderson, J. L.; Beeson, M. H.; Bentley, R. D.; Fecht, K. R.; Hooper, P. R.; Niem, A. R.; Reidel, S. P.; Swanson, D. A.; Tolan, T. L.; Wright, T. L., 1987, Distribution maps of stratigraphic units of the Columbia River Basalt Group. In Schuster, J. E., editor, 1987, Selected papers on the geology of Washington: Washington Division of Geology and Earth Resources Bulletin 77, (this volume).
- Armstrong, R. L., 1975a, Precambrian (1,500 m.y. old) rocks of central Idaho—The Salmon River Arch and its role in Cordilleran sedimentation and tectonics: *American Journal of Science*, v. 275-A, p. 437-467.

- Armstrong, R. L., 1975b, The geochronometry of Idaho: Isochron/West, no. 14, p. 1-50.
- Armstrong, R. L., 1976, The geochronometry of Idaho: Isochron/West, no. 15, p. 1-33.
- Armstrong, R. L.; Taubeneck, W. H.; Hales, P. O., 1977, Rb-Sr and K-Ar geochronometry of Mesozoic granitic rocks and their Sr isotopic composition, Oregon, Washington and Idaho: Geological Society of America Bulletin, v. 88, no. 3, p. 397-411.
- Axelrod, D. I., 1968, Tertiary floras and topographic history of the Snake River Basin, Idaho: Geological Society of America Bulletin, v. 79, no. 6, p. 713-734.
- Barrash, Warren; Bond, J. G.; Kauffman, J. D.; Venkatakrishnan, Ramesh, 1980, Geology of the La Grande area, Oregon: Oregon Department of Geology and Mineral Industries Special Paper 6, 47 p.
- Barrash, Warren; Venkatakrishnan, Ramesh, 1982, Timing of late Cenozoic volcanic and tectonic events along the western margin of the North American plate: Geological Society of America Bulletin, v. 93, no. 10, p. 977-989.
- Basham, W. L., 1978, Paleomagnetic evidence for structural rotation in the Pacific Northwest: University of Oklahoma Doctor of Philosophy thesis, 180 p.
- Beck, M. E., Jr., 1976, Discordant paleomagnetic pole positions as evidence of regional shear in the western Cordillera of North America: American Journal of Science, v. 276, no. 6, p. 694-712.
- Beck, M. E., Jr.; Faxon, M. F.; Fox, K. F., Jr., 1982, Post early Eocene tectonic rotation of northeastern Washington, and the emplacement history of "Quesnellia" [abstract]: Geological Society of America Abstracts with Programs, v. 14, no. 7, p. 441.
- Beeson, M. H.; Fecht, K. R.; Reidel, S. P.; Tolan, T. L., 1985, Regional correlations within the Frenchman Springs Member of the Columbia River Basalt Group—New insight into the middle Miocene tectonics of northwestern Oregon: Oregon Geology, v. 47, p. 87-96.
- Bennett, E. H., 1980, Granitic rocks of Tertiary age in the Idaho Batholith and their relation to mineralization: Economic Geology, v. 75, no. 2, p. 278-288.
- Bentley, R. D.; Campbell, N. P., 1983a, Geologic map of the Yakima quadrangle, Washington: Washington Division of Geology and Earth Resources Geologic Map GM-29, 1 sheet, scale 1:62,500.
- Bentley, R. D.; Campbell, N. P., 1983b, Geologic map of the Ellensburg quadrangle, Washington: Washington Division of Geology and Earth Resources Geologic Map GM-28, 1 sheet, scale 1:62,500.
- Brock, M. R.; Grolier, M. J., 1973, Chemical analyses of 305 basalt samples from the Columbia River Plateau, Washington, Oregon, and Idaho: U.S. Geological Survey Open-File Report 73-33, 35 p.
- Brooks, H. C.; Vallier, T. L., 1978, Mesozoic rocks and tectonic evolution of eastern Oregon and western Idaho. In Howell, D. G.; McDougall, K. A., editors, 1978, Mesozoic paleogeography of the western United States: Society of Economic Paleontologists and Mineralogists, Pacific Section, Pacific Coast Paleogeography Symposium 2, p. 133-146.
- Caggiano, J. A.; Duncan, D. W., editors; and others, 1983, Preliminary interpretation of the tectonic stability of the reference repository location, Cold Creek syncline, Hanford Site: Rockwell Hanford Operations Report RHO-BW-ST 19 P, 175 p., 5 pl.
- Camp, V. E., 1976, Petrochemical stratigraphy and structure of the Columbia River basalt, Lewiston Basin area, Idaho—Washington: Washington State University Doctor of Philosophy thesis, 201 p., 1 pl.
- Camp, V. E., 1981, Geologic studies of the Columbia Plateau—Part II. Upper Miocene basalt distribution, reflecting source locations, tectonism, and drainage history in the Clearwater embayment, Idaho: Geological Society of America Bulletin, Part I, v. 92, no. 9, p. 669-678.
- Camp, V. E.; Hooper, P. R., 1981, Geologic studies of the Columbia Plateau—Part I. Late Cenozoic evolution of the southeastern part of the Columbia River Basalt Province: Geological Society of America Bulletin, Part I, v. 92, no. 9, p. 659-668.
- Campbell, C. D.; Runcorn, S. K., 1956, Magnetization of the Columbia River Basalts in Washington and northern Oregon: Journal of Geophysical Research, v. 61, no. 3, p. 449-458.
- Carlson, R. L., 1976, Cenozoic plate convergence in the vicinity of the Pacific Northwest—A synthesis and assessment of plate tectonics in the northeastern Pacific: University of Washington Doctor of Philosophy thesis, 129 p.
- Carlson, R. W.; Lugmair, G. W.; Macdougall, J. D., 1981, Columbia River volcanism—The question of mantle heterogeneity or crustal contamination: Geochimica et Cosmochimica Acta, v. 45, no. 12, p. 2483-2499.
- Carlson, R. W.; Lugmair, G. W.; Macdougall, J. D., 1983, "Columbia River volcanism—The question of mantle heterogeneity or crustal contamination" (reply to a comment by D. J. De Paolo): Geochimica et Cosmochimica Acta, v. 47, no. 4, p. 845-846.
- Choiniere, S. R.; Swanson, D. A., 1979, Magnetostratigraphy and correlation of Miocene basalts of the northern Oregon coast and Columbia Plateau, southeast Washington: American Journal of Science, v. 277, no. 7, p. 755-777.
- Cockerham, R. S., 1975, Stratigraphy of the Columbia River Group, north-central Oregon: Western Washington University Master of Science thesis, 56 p.
- Cockerham, R. S.; Bentley, R. D., 1973, Picture Gorge and Yakima Basalt between Clarno and Butte Creek, Oregon [abstract]: Geological Society of America Abstracts with Programs, v. 5, no. 1, p. 23.
- De Paolo, D. J., 1983, Comment on "Columbia River volcanism—The question of mantle heterogeneity or crustal contamination" by R. W. Carlson, G. W. Lugmair, and J. D. Macdougall: Geochimica et Cosmochimica Acta, v. 47, no. 4, p. 841-844.
- Dickinson, W. R.; Thayer, T. P., 1978, Paleogeographic and paleotectonic implications of Mesozoic stratigraphy and structure in the John Day inlier of central Oregon. In Howell, D. G.; McDougall, K. A., editors, 1978, Mesozoic paleogeography of the western United States: Society of Economic Paleontologists and Mineralogists, Pacific Section, Pacific Coast Paleogeography Symposium 2, p. 147-161.
- Doell, R. R.; Cox, Allan, 1962, Determination of the magnetic polarity of rock samples in the field: U.S. Geological Survey Professional Paper 450-D, p. 105-108.
- Fecht, K. R.; Reidel, S. P.; Tallman, A. M., 1987, Paleodrainage of the Columbia River system on the Columbia Plateau of Washington State—A summary. In Schuster, J. E., editor, 1987, Selected papers on the geology of Washington: Washington Division of Geology and Earth Resources Bulletin 77, (this volume).
- Fitzgerald, J. F., 1982, Geology and basalt stratigraphy of the Weiser embayment, west-central Idaho. In Bonnicksen, Bill; Breckenridge, R. M., editors, 1982, Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 103-128.
- Fox, K. F.; Beck, M. E., Jr., 1985, Paleomagnetic results for Eocene volcanic rocks from northeastern Washington and the Tertiary tectonics of the Pacific Northwest: Tectonics, v. 4, p. 323-341.
- Fruchter, J. S.; Baldwin, S. F., 1975, Correlations between dikes of the Monument Swarm, central Oregon, and Picture Gorge Basalt flows: Geological Society of America Bulletin, v. 86, no. 4, p. 514-516.

- Gibson, I. L., 1969, A comparative account of the flood basalt volcanism of the Columbia Plateau and eastern Iceland: *Bulletin Volcanologique*, v. 33, no. 2, p. 419-437.
- Goles, G. G.; Leeman, W. P., 1976, Petrogenesis of Picture Gorge Basalts—A Preliminary Report [abstract]: *Geological Society of America Abstracts with Programs*, v. 8, no. 3, p. 376.
- Hamilton, Warren, 1978, Mesozoic tectonics of the western United States. In Howell, D. G.; McDougall, K. A., editors, 1978, Mesozoic paleogeography of the western United States: *Society of Economic Paleontologists and Mineralogists, Pacific Section, Pacific Coast Paleogeography Symposium 2*, p. 33-70.
- Helz, R. T., 1978, The petrogenesis of the Ice Harbor Member, Columbia Plateau, Washington—A chemical and experimental study: *Pennsylvania State University Doctor of Philosophy thesis*, 284 p.
- Hill, D. P., 1978, Seismic evidence for the structure and Cenozoic tectonics of the Pacific Coast states. In Smith, R. B.; Eaton, G. P., editors, 1978, Cenozoic tectonics and regional geophysics of the western Cordillera: *Geological Society of America Memoir 152*, p. 145-174.
- Hillhouse, J. W.; Grommé, C. S.; Vallier, T. L., 1982, Paleomagnetism and Mesozoic tectonics of the Seven Devils volcanic arc in northeastern Oregon: *Journal of Geophysical Research*, v. 87, no. 85, p. 3777-3794.
- Holden, G. S.; Hooper, P. R., 1976, Petrology and chemistry of a Columbia River Basalt section, Rocky Canyon, west-central Idaho: *Geological Society of America Bulletin*, v. 87, no. 2, p. 215-225.
- Hooper, P. R., 1974, Petrology and chemistry of the Rock Creek flow, Columbia River Basalt, Idaho: *Geological Society of America Bulletin*, v. 85, no. 1, p. 15-26.
- Hooper, P. R., 1981, The role of magnetic polarity and chemical analyses in establishing the stratigraphy, tectonic evolution, and petrogenesis of the Columbia River Basalt. In Subbarao, K. V.; Sukheswala, R. N., editors, 1981, Deccan volcanism and related basalt provinces in other parts of the world: *Geological Society of India Memoir 3*, p. 362-376.
- Hooper, P. R., 1982, Structural model for the Columbia River Basalt near Riggins, Idaho. In Bonnicksen, Bill; Breckenridge, R. M., editors, 1982, Cenozoic Geology of Idaho: *Idaho Bureau of Mines and Geology Bulletin 26*, p. 129-136.
- Hooper, P. R.; Camp, V. E., 1981, Deformation of the southeast part of the Columbia Plateau: *Geology*, v. 9, no. 7, p. 323-328.
- Hooper, P. R.; Kleck, W. D.; Knowles, C. R., 1976a, Composition and differentiation of the Imnaha Basalt, Columbia River Group [abstract]: *Geological Society of America Abstracts with Programs*, v. 8, no. 3, p. 383.
- Hooper, P. R.; Kleck, W. D.; Knowles, C. R.; Reidel, S. P.; Thiessen, R. L., 1984, Imnaha Basalt, Columbia River Basalt Group: *Journal of Petrology*, v. 25, part 2, p. 473-500.
- Hooper, P. R.; Knowles, C. R.; Watkins, N. D., 1979, Magnetostratigraphy of the Imnaha and Grande Ronde basalts in the southeast part of the Columbia Plateau: *American Journal of Science*, v. 279, no. 7, p. 737-754.
- Hooper, P. R.; Reidel, S. P.; Brown, J. C.; Holden, G. S.; Kleck, W. D.; Sundstrom, C. E.; Taylor, T. L., 1976b, Major element analyses of Columbia River Basalt, Part I: *Washington State University Geology Department Open-File Report*, 59 p.
- Hooper, P. R.; Rosenberg, P. E., 1970, The petrology of Granite Point, southeastern Washington: *Northwest Science*, v. 44, no. 2, p. 131-142.
- Hooper, P. R.; Webster, G. D., 1982, Geology of the Pullman, Moscow West, Colton, and Uniontown 7 1/2-minute quadrangles, Washington and Idaho: *Washington Division of Geology and Earth Resources Geologic Map GM-26*, 1 sheet, scale 1:62,500.
- Hooper, P. R.; Webster, G. D.; Camp, V. E., 1985, Geologic map of the Clarkston, 15-minute quadrangle, Washington and Idaho. *Washington Division of Geology and Earth Resources Geologic Map GM-31*, 1 sheet, scale 1:48,000.
- Jones, D. L.; Silberling, N. J.; Hillhouse, J., 1977, Wrangellia—A displaced terrane in northwestern North America: *Canadian Journal of Earth Sciences*, v. 14, no. 11, p. 2565-2577.
- Kuhns, M. J. P., 1980, Late Cenozoic deposits of the lower Clearwater Valley, Idaho and Washington: *Washington State University Master of Science thesis*, 71 p.
- Kuhns, R. J., 1980, Structural and chemical aspects of the Lochsa geothermal system near the northern margin of the Idaho batholith: *Washington State University Master of Science thesis*, 103 p.
- Lindsley, D. H., 1960, Geology of the Spray quadrangle, Oregon, with special emphasis on the petrography and magnetic properties of the Picture Gorge Basalt: *The Johns Hopkins University Doctor of Philosophy thesis*, 226 p.
- Lindsley, D. H.; Smith, Douglas; Haggerty, S. E., 1971, Petrography and mineral chemistry of a differentiated flow of Picture Gorge Basalt near Spray, Oregon—Geologic setting and mechanism of differentiation: *Carnegie Institution of Washington Yearbook*, no. 69, p. 264-285.
- Lupher, R. L.; Warren, W. C., 1942, The Asotin stage of the Snake River Canyon near Lewiston, Idaho: *Journal of Geology*, v. 50, no. 7, p. 866-881.
- Martin, B. S., 1984, Paleomagnetism of basalts in northeastern Oregon and west-central Idaho: *Washington State University Master of Science thesis*, 151 p.
- McDougall, Ian, 1976, Geochemistry and origin of basalt of the Columbia River Group, Oregon and Washington: *Geological Society of America Bulletin*, v. 87, no. 5, p. 777-792.
- McKee, E. H.; Hooper, P. R.; Kleck, W. D., 1981, Age of Imnaha Basalt—Oldest basalt flows of the Columbia River Basalt Group, northwest United States: *Isochron/West*, no. 31, p. 31-33.
- McKee, E. H.; Swanson, D. A.; Wright, T. L., 1977, Duration and volume of Columbia River basalt volcanism, Washington, Oregon and Idaho [abstract]: *Geological Society of America Abstracts with Programs*, v. 9, no. 4, p. 463-464.
- Myers, C. W.; Price, S. M.; and others, 1979, Geologic studies of the Columbia Plateau—A status report: *Rockwell Hanford Operations RHO-BWI-ST-4*, 541 p., 53 pl.
- Nathan, S.; Fruchter, J. S., 1974, Geochemical and paleomagnetic stratigraphy of the Picture Gorge and Yakima Basalts (Columbia River Group) in central Oregon: *Geological Society of America Bulletin*, v. 85, no. 1, p. 63-76.
- Nelson, D. O., 1978, Strontium isotopic and trace element geochemistry of the Grande Ronde basalts of the Columbia River Group—Evidence for a heterogeneous source region [abstract]: *Geological Society of America Abstracts with Programs*, v. 10, no. 5, p. 235.
- Nelson, D. O., 1980, Strontium isotopic and trace element geochemistry of the Saddle Mountains and Grande Ronde Basalts of the Columbia River Group: *Oregon State University Doctor of Philosophy thesis*, 224 p.
- Nelson, D. O., 1983, Implications of oxygen-isotope data and trace-element modeling for a large scale mixing model for the Columbia River Basalt: *Geology*, v. 11, no. 4, p. 248-251.

- Nelson, D. O.; Cocker, J. D.; Dasch, E. J., 1980, Strontium and oxygen isotopic studies of radiogenic basalts of the Saddle Mountains Basalt—A case for contamination [abstract]: Geological Society of America Abstracts with Programs, v. 12, no. 3, p. 144.
- Osawa, Masumi; Goles, G. G., 1970, Trace element abundances in Columbia River Basalts. In Gilmour, E. H., Stradling, D. F., editors, 1970, Proceedings of the Second Columbia River Basalt Symposium: Eastern Washington State College Press, p. 55-71.
- Price, E. H., 1982, Structural geometry, strain distribution, and mechanical evolution of eastern Untanum Ridge and a comparison with other selected localities within Yakima fold structures, south central Washington: Washington State University Doctor of Philosophy thesis, 196 p.
- Price, S. M., 1977, An evaluation of dike-flow correlations indicated by geochemistry, Chief Joseph swarm, Columbia River Basalt: University of Idaho Doctor of Philosophy thesis, 320 p.
- Reidel, S. P., 1978, The stratigraphy and petrogenesis of the Grande Ronde Basalt in the lower Salmon and adjacent Snake River canyons: Washington State University Doctor of Philosophy thesis, 415 p., 1 pl.
- Reidel, S. P., 1983, Stratigraphy and petrogenesis of the Grande Ronde Basalt from the deep canyon country of Washington, Oregon, and Idaho: Geological Society of America Bulletin, v. 94, no. 4, p. 519-542.
- Reidel, S. P.; Camp, V. E.; Ross, M. E., 1981, Major element analyses of Columbia River Basalt, Part II: Washington State University Geology Department Basalt Research Group, 90 p.
- Reidel, S. P.; Long, P. E.; Myers, C. W.; Mase, J., 1982, New evidence for greater than 3.2 km of Columbia River basalt beneath the central Columbia Plateau [abstract]: EOS (American Geophysical Union Transactions), v. 63, no. 8, p. 173.
- Robyn, T. L.; Hoover, J. D., 1982, Late Cenozoic deformation and volcanism in the Blue Mountains of central Oregon—Microplate interactions?: Geology, v. 10, no. 11, p. 572-576.
- Ross, M. E., 1978, Stratigraphy, structure and petrology of Columbia River basalt in a portion of the Grande Ronde River-Blue Mountains area of Oregon and Washington: University of Idaho Doctor of Philosophy thesis, 407 p., 1 pl.
- Sarewitz, Daniel, 1983, Seven Devils terrane—Is it really a piece of Wrangellia?: Geology, v. 11, no. 11, p. 634-637.
- Scheffler, J. M., 1983, A petrologic and tectonic comparison of the Hells Canyon area, Oregon-Idaho, and Vancouver Island, British Columbia: Washington State University Master of Science thesis, 98 p.
- Schmincke, H. -U., 1967, Stratigraphy and petrography of four upper Yakima Basalt flows in south-central Washington: Geological Society of America Bulletin, v. 78, no. 11, p. 1385-1422.
- Shaw, H. R.; Swanson, D. A., 1970, Speculations on the fluid mechanical history of the Yakima basalt flows [abstract]. In Gilmour, E. H.; Stradling, D. F., editors, 1970, Proceedings of the Second Columbia River Basalt Symposium: Eastern Washington State College Press, p. 330.
- Shubat, M. A., 1979, Stratigraphy, petrochemistry, petrography and structural geology of the Columbia River Basalt in the Minam-Wallowa River area, northeast Oregon: Washington State University Master of Science thesis, 156 p.
- Snively, P. D., Jr.; MacLeod, N. J.; Wagner, H. C., 1973, Miocene tholeiitic basalts of coastal Oregon and Washington and their relations to coeval basalts of the Columbia Plateau: Geological Society of America Bulletin, v. 84, no. 2, p. 387-424.
- Swanson, D. A.; Anderson, J. L.; Bentley, R. D.; Byerly, G. R.; Camp, V. E.; Gardner, J. N.; Wright, T. L., 1979a, Reconnaissance geologic map of the Columbia River Basalt Group in eastern Washington and northern Idaho: U.S. Geological Survey Open-File Report 79-1363, 26 p., 12 pl., scale 1:250,000.
- Swanson, D. A.; Anderson, J. L.; Camp, V. E.; Hooper, P. R.; Taubeneck, W. H.; Wright, T. L., 1981, Reconnaissance geologic map of the Columbia River Basalt Group, northern Oregon and western Idaho: U.S. Geological Survey Open-File Report 81-797, 35 p., 5 pl., scale 1:250,000.
- Swanson, D. A.; Bentley, R. D.; Byerly, G. R.; Gardner, J. N.; Wright, T. L., 1979b, Preliminary reconnaissance geologic maps of the Columbia River Basalt Group in parts of eastern Washington and northern Idaho: U.S. Geological Survey Open-File Report 79-534, 40 p., 9 pl., scale 1:250,000.
- Swanson, D. A.; Wright, T. L., 1976, Guide to field trip between Pasco and Pullman, Washington, emphasizing stratigraphy, vent areas, and intracanyon flows of Yakima Basalt: Geological Society of America, Cordilleran Section Meeting, Pullman, Washington, Field Guide 1, 33 p.
- Swanson, D. A.; Wright, T. L., 1981, The regional approach to studying the Columbia River Basalt Group. In Subbarao, K. V.; Sukheswala, R. N., editors, 1981, Deccan volcanism and related basalt provinces in other parts of the world: Geological Society of India Memoir 3, p. 58-80.
- Swanson, D. A.; Wright, T. L., 1983, Geologic map of the Wenaha-Tucannon Wilderness, Washington and Oregon: U.S. Geological Survey Miscellaneous Field Studies Map MF-1536, 1 sheet, scale 1:48,000.
- Swanson, D. A.; Wright, T. L.; Camp, V. E.; Gardner, J. N.; Helz, R. T.; Price, S. M.; Reidel, S. P.; Ross, M. E., 1980, Reconnaissance geological map of the Columbia River Basalt Group, Pullman and Walla Walla quadrangles, southeast Washington and adjacent Idaho: U.S. Geological Survey Miscellaneous Investigations Series Map I-1139, 2 sheets, scale 1:250,000.
- Swanson, D. A.; Wright, T. L.; Hooper, P. R.; Bentley, R. D., 1979c, Revisions in stratigraphic nomenclature of the Columbia River Basalt Group: U.S. Geological Survey Bulletin 1457-G, 59 p.
- Swanson, D. A.; Wright, T. L.; Helz, R. T., 1975, Linear vent systems and estimated rates of magma production and eruption for the Yakima Basalt on the Columbia Plateau: American Journal of Science, v. 275, no. 8, p. 877-905.
- Taubeneck, W. H., 1970, Dikes of Columbia River basalt in northeastern Oregon, western Idaho, and southeastern Washington. In Gilmour, E. H.; Stradling, D. F., editors, 1970, Proceedings of the Second Columbia River Basalt Symposium: Eastern Washington State College Press, p. 73-96.
- Tolan, T. L., 1982, The stratigraphic relationships of the Columbia River Gorge of Oregon and Washington: Portland State University Master of Science thesis, 151 p., 1 pl.
- Uppuluri, V. R., 1974, Prineville chemical type—A new basalt type in the Columbia River Group: Geological Society of America Bulletin, v. 85, no. 8, p. 1315-1318.
- Vallier, T. L., 1977, The Permian and Triassic Seven Devils Group, western Idaho and northeastern Oregon: U.S. Geological Survey Bulletin 1437, 58 p.
- Vallier, T. L.; Brooks, H. C.; Thayer, T. P., 1977, Paleozoic rocks of eastern Oregon and western Idaho. In Stewart, J. H.; Stevens, C. H.; Fritsche, A. E., editors, 1977, Paleozoic paleogeography of the western United States: Society of Economic Paleontologists and Mineralogists, Pacific Section, Pacific Coast Paleogeography Symposium 1, p. 455-466.
- Vallier, T. L.; Hooper, P. R., 1976, Geologic guide to Hells Canyon, Snake River: Geological Society of America, Cordilleran Section Meeting, 72nd Annual Meeting, Field guide 5: Washington State University, 38 p.

- Walker, G. W., 1973, Contrasting compositions of the youngest Columbia River basalt flows in Union and Wallowa Counties, northeastern Oregon: *Geological Society of America Bulletin*, v. 84, no. 2, p. 425-430.
- Washington, H. S., 1922, Deccan traps and other plateau basalts: *Geological Society of America Bulletin*, v. 33, no. 4, p. 765-804.
- Waters, A. C., 1961, Stratigraphic and lithologic variations in the Columbia River Basalt: *American Journal of Science*, v. 259, no. 8, p. 538-611.
- Watkins, N. D.; Baksi, A. K., 1974, Magnetostratigraphy and oroclinal folding of the Columbia River, Steens and Owyhee basalts in Oregon, Washington and Idaho: *American Journal of Science*, v. 274, no. 2, p. 148-189.
- Wells, R. E.; Simpson, R. W.; Kelley, M. M.; Beeson, M. H.; Bentley, R. D., 1983, Columbia River Basalt stratigraphy and rotation in southwest Washington [abstract]: *EOS (American Geophysical Union Transactions)*, v. 64, no. 45, p. 687.
- Wilkinson, J. F. G.; Binns, R. A., 1977, Relatively iron-rich lherzolite xenoliths of the Cr-diopside suite—A guide to the primary nature of anorogenic tholeiitic andesite magmas: *Contributions to Mineralogy and Petrology*, v. 65, no. 2, p. 199-212.
- Wilson, D.; Cox, A., 1980, Paleomagnetic evidence for tectonic rotation of Jurassic plutons in Blue Mountains, eastern Oregon: *Journal of Geophysical Research*, v. 85, no. B7, p. 3681-3689.
- Wise, D. U., 1963, An outrageous hypothesis for the tectonic pattern of the North American cordillera: *Geological Society of America Bulletin*, v. 74, no. 3, p. 357-362.
- Wright, T. L.; Black, K. N.; Swanson, D. A.; O'Hearn, Tim, 1980, Columbia River Basalt—1978-1979 sample data and chemical analyses: U.S. Geological Survey Open-File Report 80-921, 99 p.
- Wright, T. L.; Black, K. N.; Swanson, D. A.; O'Hearn, Tim, 1982, Columbia River Basalt—1980-1981 sample data and chemical analyses: U.S. Geological Survey Open-File Report 82-532, 62 p.
- Wright, T. L.; Grolier, M. J.; Swanson, D. A., 1973, Chemical variation related to the stratigraphy of the Columbia River basalt: *Geological Society of America Bulletin*, v. 84, no. 2, p. 371-385.
- Wright, T. L.; Helz, R. T., 1981, Generation of tholeiitic basalt—A comparative study of three intraplate provinces. In Subbarao, K. V.; Sukhwala, R. N., editors, 1981, Deccan volcanism and related basalt provinces in other parts of the world: *Geological Society of India Memoir* 3, p. 419-421.
- Wright, T. L.; Swanson, D. A.; Fruchter, J., 1976, Petrogenesis of Yakima Basalt in southeastern Washington [abstract]: *Geological Society of America Abstracts with Programs*, v. 8, no. 3, p. 422.
- Wright, T. L.; Swanson, D. A.; Helz, R. T.; Byerly, G. R., 1979, Major oxide, trace element, and glass chemistry of Columbia River basalt samples collected between 1971 and 1977: U.S. Geological Survey Open-File Report 79-711, 146 p.
- Youngs, S. W., 1981, Geology and geochemistry of the Running Springs geothermal area, southern Bitterroot lobe, Idaho batholith: Washington State University Master of Science thesis, 125 p.

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PALEODRAINAGE OF THE COLUMBIA RIVER SYSTEM ON THE COLUMBIA PLATEAU OF WASHINGTON STATE— A SUMMARY¹

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ABSTRACT

Over the last 17 million years, stream courses of the Columbia River system on the Columbia Plateau in Washington State have evolved primarily as a result of the effects of volcanism and deformation. Repeated eruptions of Columbia River basalt in early to middle Miocene time formed a vast basaltic plain and obliterated the prebasalt drainage system. Major streams during this interval were flowing westward, apparently near the periphery of the evolving plateau and through a broad lowland in the ancestral Cascade Range generally south of the present-day Columbia River Gorge. Stream courses were controlled regionally by the west-dipping Palouse slope and the constructional topography of the Grande Ronde flows and locally by emerging structural topography. In middle Miocene time, a similar drainage system on the plateau was periodically altered by Wanapum Basalt. Middle to late Miocene major streams formed a well-integrated river system that flowed in a centripetal pattern from the surrounding highlands, across regional paleoslopes and into the central plateau. In the central plateau, the stream courses were controlled by the topographic relief of developing structures and constructional topography formed as the result of eruptions of Saddle Mountains Basalt. The major streams joined in the central plateau and continued westward through the ancestral Columbia River Gorge of the ancestral Cascade Range.

In late Miocene time, after the Columbia River basalt eruptive episodes, stream courses were controlled on a regional basis by the major paleoslopes and locally by developing structural basins. Water gaps developed through structural ridges bounding the basins. Streams deposited thick sedimentary sequences in the basins until about the middle Pliocene, when streams began a period of degradation. The aggradational and degradational events may have been a factor in base-level changes that may have resulted from Cascadian volcanism near the present-day Columbia River Gorge.

During the Pleistocene, glaciers that impinged on the northern Columbia Plateau and catastrophic flooding that emanated from ice-margin lakes obliterated much of the secondary drainage and caused temporary or minor changes to the major stream courses on the Columbia Plateau.

INTRODUCTION

The present-day Columbia River system covers about 670,000 km² and occupies parts of seven states in the northwestern United States and part of southern British Columbia, Canada. The system extends through the intermontane basin between the Rocky Mountains and the Cascade Range. In the heart of that basin, the Columbia River and two of its major tributaries, the Snake and Yakima Rivers, flow across basaltic terrain of the Columbia Plateau (Fig. 1). On the plateau these rivers cross broad sediment-filled basins and valleys and pass through spectacular water gaps and gorges deeply entrenched into the basaltic ridges and highlands. Exposed in and overlying the

basalt sequence are extensive fluvial and lacustrine sediments, buried paleochannels and paleocanyons, and hyaloclastites and pillow basalts that mark the course of earlier streams and the presence of former lakes. These features, many kilometers from present river courses, suggest a long and complex history of drainage development on the Columbia Plateau.

Since the late 1800s geologists have studied the paleodrainage of the Columbia River system on the Columbia Plateau. Earlier workers, notably Willis (1887), Smith (1901, 1903), Bretz (1923, 1959), Warren (1941), Waters (1955), Mackin (1961), and Bond (1963), provided important observations and discussed the position and age of river courses before many of the stratigraphic details of the Columbia Plateau had been deciphered. Many of their interpretations are still valid. Recent advances in stratigraphy of the Columbia

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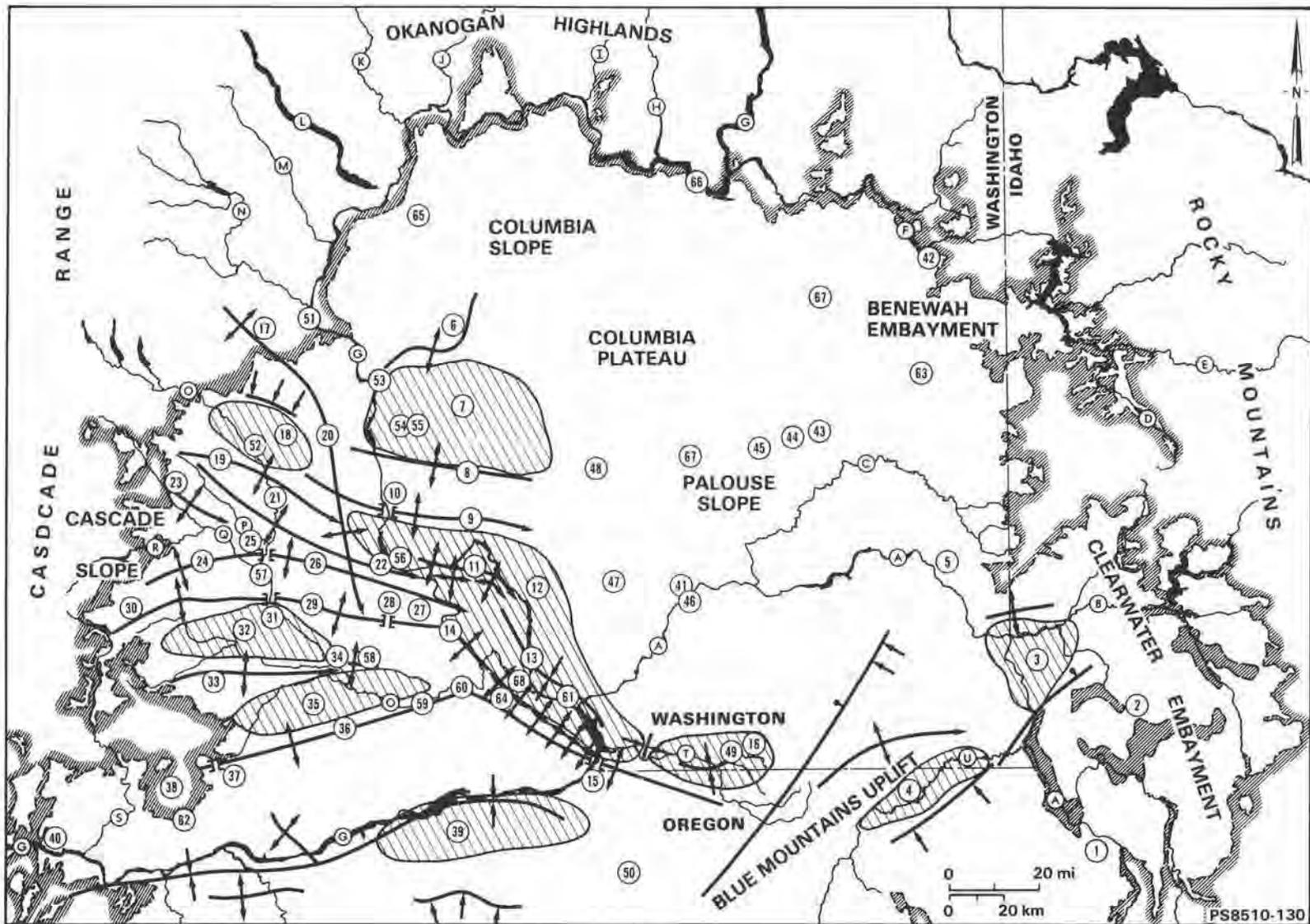


Figure 1.—Northern part of the Columbia River basalt province.

LEGEND
(FOR FIGURE 1)

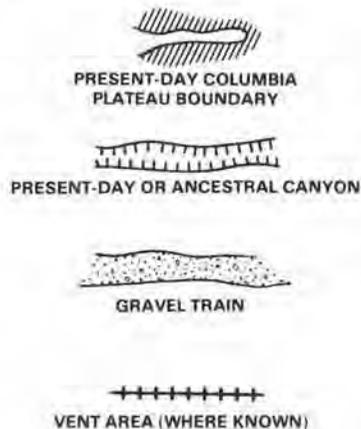
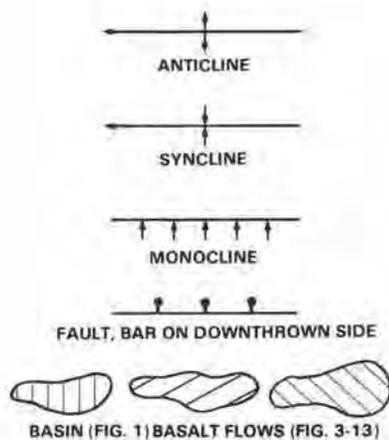
Rivers and Lakes

- A - Snake
- B - Clearwater
- C - Palouse
- D - St. Maries
- E - St. Joe
- F - Spokane
- G - Columbia
- H - Sanpoil
- I - Nespelem
- J - Okanogan
- K - Methow
- L - Lake Chelan
- M - Entiat
- N - Wenatchee
- O - Yakima
- P - Wenas Creek
- Q - Naches
- R - Tieton
- S - Klickitat
- T - Walla Walla
- U - Grande Ronde

Features

- 1. Joseph Plains
- 2. Nez Perce Plateau
- 3. Lewiston Basin
- 4. Troy Basin
- 5. Uniontown Plateau
- 6. Beezley Hills
- 7. Quincy Basin
- 8. Frenchman Hills
- 9. Saddle Mountains
- 10. Sentinel Gap
- 11. Wahluke Syncline
- 12. Pasco Syncline
- 13. Pasco Basin
- 14. Rattlesnake Mountain
- 15. Wallula Gap
- 16. Walla Walla Basin
- 17. Naneum Ridge
- 18. Kittitas Basin
- 19. Manastash Ridge
- 20. Hog Ranch Structure
- 21. Yakima Canyon
- 22. Umtanum Ridge
- 23. Cleman Mountain
- 24. Cowiche Mountains
- 25. Selah Gap
- 26. Yakima Ridge
- 27. Black Rock Valley
- 28. Sunnyside Gap
- 29. Rattlesnake Hills
- 30. Ahtanum Ridge
- 31. Union Gap
- 32. Toppenish Basin
- 33. Toppenish Ridge
- 34. Snipes Mountain
- 35. Satus Basin
- 36. Horse Heaven Hills
- 37. Satus Pass
- 38. Simcoe Mountains
- 39. Umatilla Basin
- 40. Columbia River Gorge
- 41. Devil's Canyon
- 42. Spokane
- 43. Revere
- 44. McCall
- 45. Marengo
- 46. Lower Monumental Dam
- 47. Old Maid Coulee
- 48. Warden
- 49. Walla Walla
- 50. Pendleton
- 51. Wenatchee
- 52. Ellensburg
- 53. Crescent Bar
- 54. George
- 55. Winchester Wasteway
- 56. Priest Rapids Dam
- 57. Yakima
- 58. Sunnyside
- 59. Prosser
- 60. Chandler
- 61. Pasco
- 62. Goldendale
- 63. Malden
- 64. Badger Coulee
- 65. Waterville Plateau
- 66. Columbia Valley
- 67. Cheney-Palouse Scabland Tract
- 68. Yakima Bluffs

(FOR FIGURES 1 AND 3 THROUGH 13)



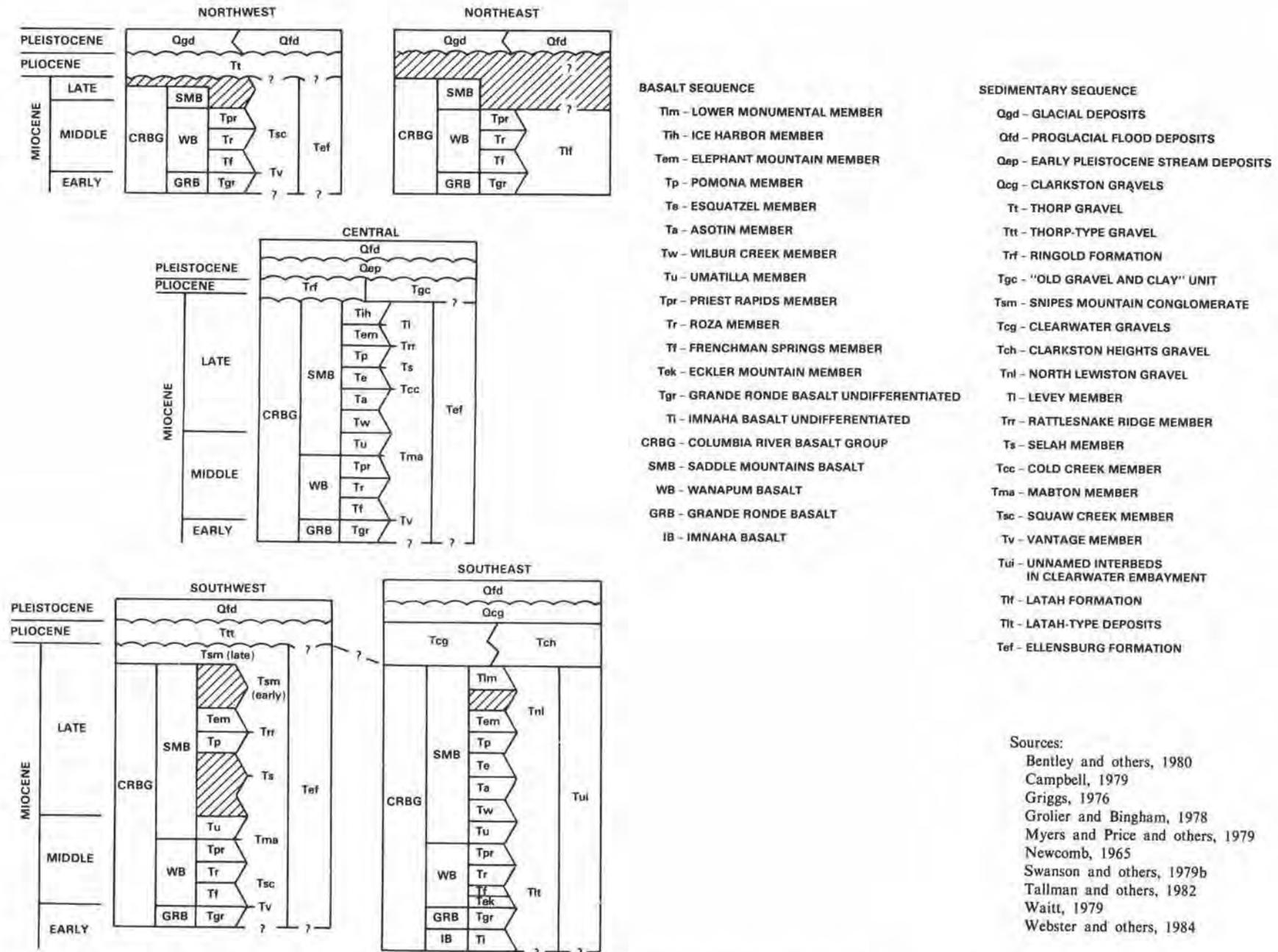


Figure 2.—Generalized stratigraphy of the Columbia Plateau in Washington State.

River Basalt Group have led to a better understanding of the location and age of ancestral stream courses within Washington State (Schmincke, 1964, 1967a; Swanson and others, 1979a,b, 1980; Myers and Price and others, 1979; Rigby and others, 1979). To date, the most thorough studies of paleostream courses on the plateau are by Schmincke (1964, 1967a) for the western and central plateau and by Swanson and Wright (1976, 1979), Swanson and others (1979b, 1980), and Camp (1981) for the eastern plateau.

The purposes of this paper are to (1) summarize the geologic evidence for ancestral stream courses, (2) describe the Neogene and Pleistocene evolution of the Columbia River drainage system, and (3) demonstrate the dynamic interaction between geologic events and paleodrainage on the Columbia Plateau of Washington State following the onset of Columbia River basalt volcanism at about 17 m.y. ago. Emphasis throughout the paper is on post-middle Miocene units in south-central Washington, where the authors have been conducting extensive field studies. The paper builds on previous observations of paleostream deposits and the inferred stream courses.

General Stratigraphic Relations

The Columbia Plateau is composed of a series of thick Miocene tholeiitic basalt flows that are, in places, interbedded with and overlain by clastic sediments of Miocene and younger age (Fig. 2). The basalt flows collectively form the Columbia River Basalt Group that consists of five formations, four of which are exposed in Washington State: the Imnaha, Grande Ronde, Wanapum, and Saddle Mountains Basalts (Swanson and others, 1979a). The Imnaha Basalt is restricted to exposures along the Snake River canyon, whereas Grande Ronde, Wanapum, and Saddle Mountains Basalts are present throughout much of the Columbia Plateau (Anderson and others, this volume). Miocene sedimentary rocks interbedded with the basalt in Washington State are designated as the Ellensburg Formation (Smith, 1901, 1903) in the western and central parts of the plateau (Waters, 1955; Mackin, 1961; Schmincke, 1964, 1967a,b; Waitt, 1979; Swanson and others, 1979b) and the Latah Formation and related sediments along the eastern margin of the plateau (Pardee and Bryan, 1926; Kirkham and Johnson, 1929; Griggs 1976). In the western part of the plateau, the Ellensburg Formation, in part, overlies the basalt sequence (Russell, 1893; Smith, 1903; Waters, 1955; Schmincke, 1967a,b).

Overlying the Columbia River Basalt Group and Miocene sedimentary rocks in topographic and structural lows of Washington are: the Miocene-Pliocene

Ringold Formation in the Quincy and Pasco basins (Merriam and Buwalda, 1917; Newcomb and others, 1972; Grolier and Bingham, 1978; Tallman and others, 1981); an "old gravel and clay" unit in the Walla Walla basin (Newcomb, 1965); the Pliocene Thorp Gravel in the Kittitas Valley (Waitt, 1979); gravels similar in lithology to and of the same age as the Thorp Gravel in the Yakima Valley (Bentley, 1977; Rigby and others, 1979; Campbell, 1983) and gravels of the Lewiston basin (Webster and others, 1984).

The Quaternary stratigraphy is dominated by Pleistocene glacial deposits along the northern margin of the plateau and glaciofluvial deposits in topographic lows of the plateau and in the Channeled Scabland of the eastern part of the plateau (Bretz, 1923, 1959; Baker, 1973; Baker and Nummedal, 1978; Rigby and others, 1979; Myers and Price and others, 1979; Waitt, 1979). Flooding from glacial Lake Bonneville down the Snake River canyon in late Pleistocene time resulted in deposition of gravel in the Lewiston Basin area (Swanson, 1984). Loess and dune sand of Pleistocene and Holocene age blanket most of the high terrain of the plateau (Russell, 1893, 1902; Fryxell and Cook, 1964; Rigby and others, 1979).

Identification of Stream Courses

Ancestral stream courses on the Columbia Plateau are delineated by identifying the distribution, texture, and composition of fluvial sequences, as well as erosional features, intracanyon flows, hyaloclastites, and pillow basalts associated with the basalt sequences. Stream course locations are most commonly based on the distribution of fluvial sediments, which occur either as narrow and well-confined channel deposits or as broad extensive sheets. Commonly found within the laterally extensive fluvial sediments are narrow trains of gravel that are inferred to mark the main channels of ancestral streams. The composition of the detrital sediments is used to determine the general source area of the major streams and their tributaries. Volcaniclastic sediments were derived principally from the southern and central Cascade Range to the west of the plateau. Sediments composed mainly of plutonic and metamorphic debris were derived from the northern Cascade Range, the Okanogan Highlands to the north of the plateau, and (or) the foothills of the Rocky Mountains to the east of the plateau. Metabasalt clasts present in interbeds and suprabasalt sediments in the Lewiston area, on the Palouse Slope, and locally in the central Columbia Plateau serve to distinguish the major streams draining the northern and northeastern Columbia Plateau from those draining the eastern plateau. Fluvial deposits composed mostly of basaltic

sediments are generally assumed to have come from streams that originated on the plateau or whose drainage area is mostly within the plateau.

In places, narrow, steep-walled stream canyons have been carved into the basalt. Ancestral canyons eroded into basalt were particularly common in the eastern plateau in Washington during the middle to late Miocene (Swanson and others, 1980; Camp, 1981); they were later filled by lava to form intracanyon flows. Traces of the intracanyon flows help define the positions of ancestral streams (Swanson and Wright, 1979; Tolan and Beeson, 1984; Anderson and Vogt, this volume). During the Pleistocene, canyons were also scoured into the basalt by catastrophic outbursts of glacial lakes. These channels generally remain only partially filled by sediments and form the Channeled Scabland (Bretz, 1923).

The character of basalt units also indicates the presence of ancestral streams or lakes. Hyaloclastites and pillow basalts that resulted from the rapid quenching of lava upon entering water (Fuller, 1931; Waters, 1960) are common to many flows in the northern and western parts of the Columbia Plateau. The areal extent of these features aids in defining the position of ancestral streams or lakes and has been particularly useful in determining the location of stream drainages during the emplacement of Grande Ronde and Wanapum Basalts. However, hyaloclastites and pillow basalts are not commonly associated with ancestral streams in the central or eastern Columbia Plateau except on the immediate margin of the eastern plateau. Apparently, lavas erupted in the eastern plateau commonly dammed the upper reaches of streams on the Palouse Slope, and subsequent lavas flowed down essentially dry stream channels (Swanson and others, 1979a; Camp, 1981).

Control of Stream Courses

Geologic features controlling the courses of major streams on the Columbia Plateau include (1) broad, gently dipping regional paleoslopes, (2) constructional topography developed by the emplacement of lava flows and volcanoclastic debris, and less commonly, by sedimentary sequences, and (3) structurally uplifted ridges and subsided synclines and basins.

Three major paleoslopes have provided the primary control for stream drainage on the Columbia Plateau in Washington State: (1) the west-dipping Palouse Slope on the eastern plateau, (2) the south-dipping Columbia Slope on the north, and (3) the east-dipping Cascade Slope on the west. (See Fig. 1.) The Palouse Slope is the most extensive of the three paleoslopes. During the emplacement of the Grande Ronde and Wanapum Basalts, the Palouse Slope was the

dominant topographic feature of the plateau. It extended from the foothills of the Rocky Mountains and gently dipped westward to the ancestral Cascade Range (Swanson and others, 1979a). The Columbia Slope, which gently dips southward into the Pasco Basin (Schmincke, 1964, 1967a), formed as the result of subsidence of the central Columbia Plateau starting at least by early to middle Miocene, but the slope apparently had little effect on stream courses until about middle Miocene. The Cascade Slope is a narrow slope formed along the eastern foothills of the ancestral Cascade Range and is mantled by volcanoclastic sediments and volcanic lahars.

Emplacement of lava flows and deposition of sedimentary detritus created constructional topography that forced re-alignment of stream courses. Basalt flows dammed streams and led to the development of streams running marginal to flow fronts during most of the eruptive episode. In the western plateau during the Miocene, volcanic eruptions and east-flowing streams in the ancestral Cascade Range built volcanic lahars and aggraded volcanoclastic sediments out onto the plateau and into the major drainage channel. In the Pleistocene, catastrophic floodwaters deposited thick sequences of gravel that clogged and blocked drainages in the Pasco Basin, forcing streams to establish new courses.

Folding on the Columbia Plateau has resulted in the shifting of stream courses away from developing structural ridges and into adjacent developing structural lows and in the entrenching of streams into structural ridges. The shifts in stream courses resulting from deformation have occurred mainly along the margins of the plateau and in the Yakima fold belt of the western plateau where folding has been most extensive. Some reaches of the Columbia and Yakima Rivers are antecedent to the uplift of the Yakima folds and have incised deep gorges and water gaps between structural basins. The eastern plateau has remained relatively undeformed except for the gentle, west-dipping Palouse Slope and the basins in southeastern Washington. Streams on the Palouse Slope are consequent.

DEVELOPMENT OF STREAM DRAINAGE

Early to Middle Miocene

Grande Ronde lava erupted from fissures in the southeastern part of the Columbia Plateau between 16.5 and 15.5 m.y. ago (McKee and others, 1977; Long and Duncan, 1982) and inundated the intermontane basin between the Rocky Mountains and ancestral Cascade Range (Smith, 1901; Waters, 1961; Swanson and

others, 1979a). Streams flowing in a centripetal drainage system into that intermontane area prior to Grande Ronde eruptions were obliterated by repeated advances of lava (Bond, 1963; Conners, 1976; Waters and others, 1981). Flows of Grande Ronde Basalt built constructional topography that formed a rather featureless west-dipping regional paleoslope (Palouse Slope) (Swanson and others, 1979b; Landon and others, 1982). However, the constructional topography had sufficient relief to cause the establishment of new river courses along the margins of basalt flows. The frequent and voluminous eruptions of Grand Ronde Basalt flows prevented the development of a well-integrated drainage system during most of early to middle Miocene time.

Exposures of Grande Ronde Basalt on the margins of the plateau and in the present-day Columbia River Gorge area exhibit colonnade-entablature jointing that in places overlies pillow basalts and hyaloclastites (Fuller, 1931; Waters, 1960; Bond, 1963; Tabor and others, 1980, 1982; Tolan, 1982). Vesicular tops of many flows have been locally scoured by stream erosion. Sedimentary interbeds are commonly observed between flows and decrease in thickness and abundance from the margins onto the center of the plateau (Griggs, 1976; Bentley and others, 1980; Tabor and others, 1980, 1982). In the central plateau, Grande Ronde Basalt is known only in deeply incised canyons and from drill holes. The basalt sequence there is typified by colonnade-entablature joints rarely associated with hyaloclastites or pillow basalts and by the relative paucity of erosional features or interbeds. Local variations of flow thickness and rare, thin sedimentary interbeds penetrated in boreholes in the Pasco Basin (Raymond and Tillson, 1968; Myers and Price and others, 1979) suggest the presence of small lakes and, possibly, small streams that formed a poorly integrated drainage system in the central Columbia Plateau during the early to middle Miocene. These observations suggest that most major stream activity was restricted to near the margins of the plateau.

Sedimentary interbeds in the Grande Ronde Basalt in the northeastern part of the plateau form part of the Latah Formation (Pardee and Bryan, 1926; Griggs, 1976) and related "Latah-type" units (Kirkham and Johnson, 1929; Bond, 1963). These sediments range from claystone to fine-grained sandstone; laminated siltstone is predominant. Coarse-grained sandstone and conglomerate are uncommon interbed components (Griggs, 1976). The sedimentary interbeds are mainly composed of arkosic detritus derived from older crystalline terrain to the north and east of the present-day plateau.

Interbeds that occur along the western and northwestern plateau constitute part of the Ellens-

burg Formation (Smith, 1901, 1903; Mackin, 1961; Waters, 1955; Schmincke, 1967a,b; Waitt, 1979). The Ellensburg sedimentary rocks range from claystones to conglomerates, and siltstone and sandstone are most common. The Ellensburg rocks are arkosic in the northwestern plateau and entirely volcanoclastic farther west (Swanson, 1967; Bentley and others, 1980; Tabor and others, 1980, 1982).

Based on observations of fluvial sediments and the character of primary intraflow structures, Grande Ronde lava dammed streams flowing from the disrupted surrounding highlands, forming a drainage pattern in which shallow freshwater lakes developed behind lava flow fronts on the plateau margin. Small lakes, marshes, and, possibly, small streams were present in the central plateau. Pondered water and low-energy fluvial environments aggraded fine sediments and formed bogs and marshes, as attested by the abundance of organic debris mixed with sediments. In places, streams developed based on the presence of coarser sediments. The pondered water and streams were repeatedly inundated by lava flows, resulting in the intercalation of basalt and sediments. Subsequent to each eruption, streams were probably flowing marginal to various basalt flows in a poorly integrated ancestral Columbia River drainage system (Fig. 3). Streams in the northern and western parts of the forming plateau generally flowed between the Grande Ronde Basalt margin and the south-sloping Okanogan Highlands from northern Idaho across northern Washington, as suggested by Willis (1887). Here, the Columbia River has remained in essentially the same position along the northern margin of the Columbia Plateau since it was established during emplacement of Grande Ronde Basalt. Streams continued to flow near the plateau margin southward along the foothills of the ancestral Cascade Range, where arkosic sands transported from north of the plateau apparently mixed with, and were diluted by, the volcanoclastic and pyroclastic sediments from the ancestral Cascades (Swanson, 1967; Bentley and others, 1980). In the central plateau, small lakes and apparently small streams were probably controlled by the Palouse Slope and the constructional topography of basalt flows.

Streams that drained the Columbia Plateau during this time flowed through a broad low in the ancestral Cascade Range at the present-day site of Mount Hood (Beeson and Moran, 1979) and on to the Pacific Ocean.

Middle Miocene

The hiatus in basalt volcanism at the end of the Grande Ronde eruptive episode is represented by the Vantage Member of the Ellensburg Formation

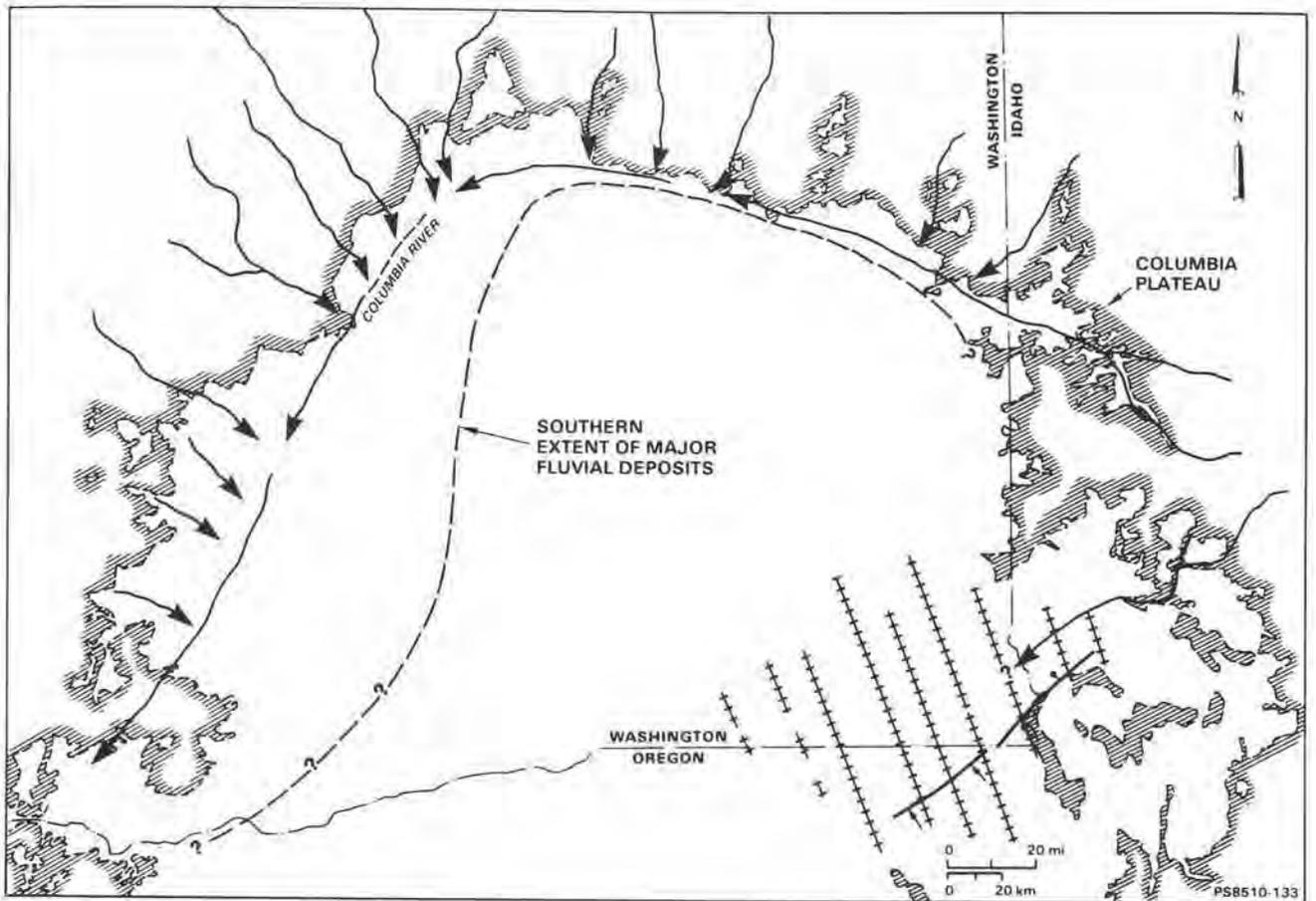


Figure 3.—Stream drainage on the northern part of the Columbia Plateau during early to middle Miocene (Grande Ronde) time.

(Mackin, 1961; Swanson and others, 1979a). West of the present-day Pasco and Quincy basins, fluvial sedimentary rocks of the Vantage Member occur between the Grande Ronde and Wanapum Basalts. South and east of these basins, a highly weathered flow top, or saprolite, caps the uppermost flow of Grande Ronde Basalt (Swanson and Wright 1976, 1981), suggesting an interlude of perhaps several hundred thousand years between the end of the Grande Ronde volcanic episode and renewed volcanism.

West of the present-day Pasco and Quincy basins, the Vantage Member is composed of arkosic and volcanoclastic fine-grained rocks (Mackin, 1961; Diery and McKee, 1969; Myers and Price and others, 1979; Bentley and others, 1980). The volcanoclastic component increases westward, and in the upper Wenatchee Valley, the Vantage Member consists entirely of volcanoclastic debris (Schmincke, 1967a). The presence of volcanoclastic sediments on the western plateau is probably due to (1) deposition by east-flowing streams of debris from active volcanic terrain of the Cascade

Range, and (or) (2) uplift of the ancestral Cascades relative to subsidence of the plateau as suggested by the offlap of Grande Ronde Basalt. The distribution of arkosic sediments suggests that the ancestral Columbia River had been diverted from near the plateau margin southeastward into the central part of the plateau (Fig. 4). The widespread distribution of sediments on the western plateau suggests that the streams probably were initially flowing over a rather broad, flat plain before becoming confined to relatively narrow courses.

The diversion of the ancestral Columbia River from its marginal course to one on the plateau near Wenatchee was originally interpreted to be a response to convergence of two lava flows (Chappell, 1936). Subsequent workers have measured numerous north- and northwest-trending basalt flow directions in the northwestern plateau, suggesting a lava source to the south or southeast, consistent with the location of dikes for Grande Ronde Basalt in the southeastern plateau (Swanson and others, 1979b). Further, mapping in the northwestern plateau has shown that feeder dikes in

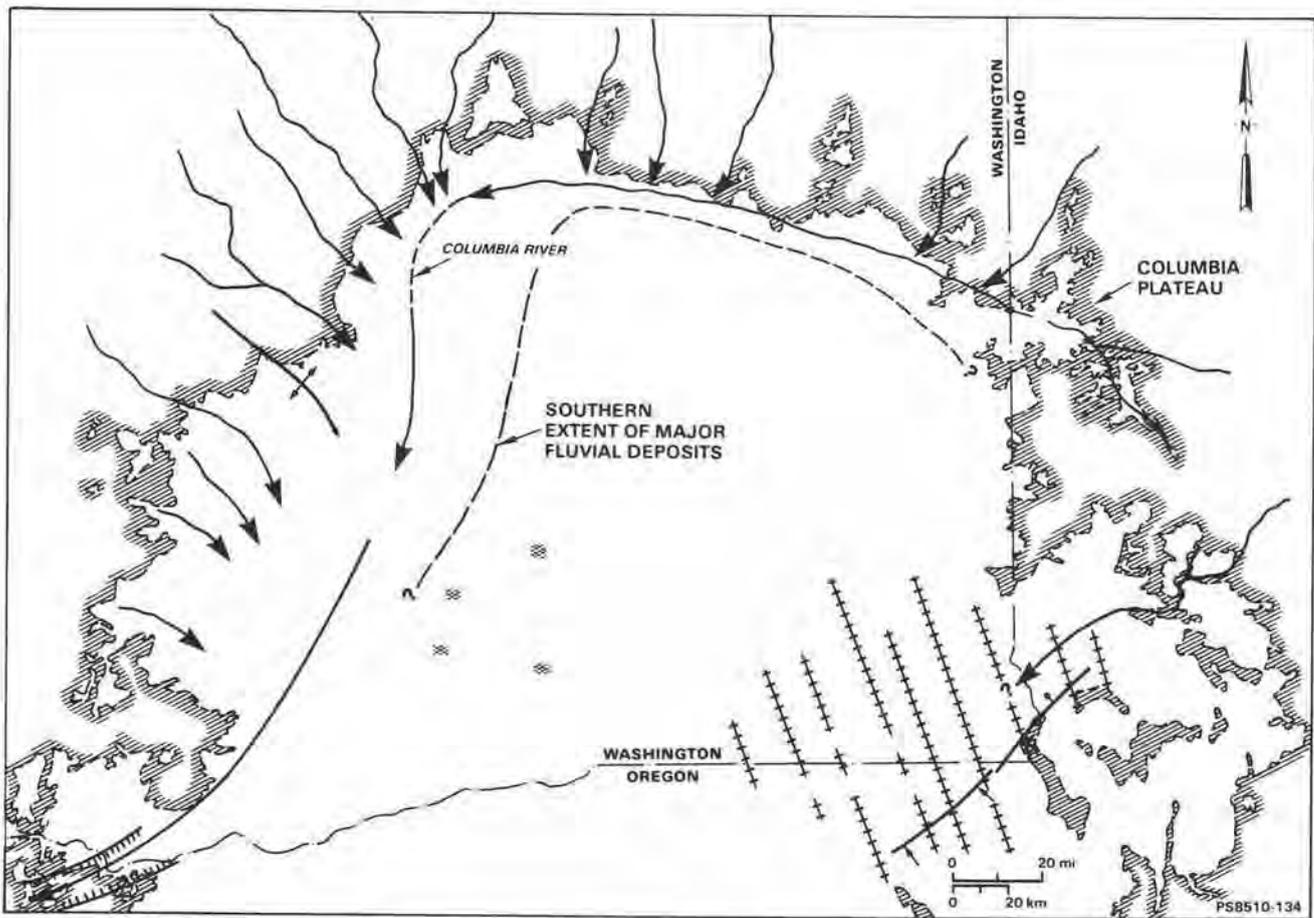


Figure 4.—Stream drainage on the northern part of the Columbia Plateau during middle Miocene (Vantage) time.

the Cascade Range and at the northwest margin of the present-day plateau are not sources for Columbia River basalts (Swanson, 1967). The east-dipping, foreset-bedded pillow basalts mapped by Chappell (1936) apparently are local features and do not represent a lava source to the west (Swanson, 1967).

Although it was rejected by Chappell (1936), another possible hypothesis for diversion of the ancestral Columbia River onto the plateau is uplift of the Wenatchee Mountains. Recent mapping southeast of Wenatchee suggests that the river was diverted as a result of uplift of the Naneum Ridge structure (Swanson and others, 1979b), which created the Wenatchee Mountains and the broad Hog Ranch structure of Mackin (1961). Evidence for structural control of the diversion is found in the thickness and composition of the Vantage Member south of Wenatchee. Field mapping there has shown that the Vantage Member thins both east and west of the course of the present-day Columbia River from Crescent Bar to the Saddle Mountains (Grolier, 1965; Grolier and Bingham, 1978;

Alto, 1955; Tabor and others, 1982). Locally, the Vantage thickens in structural lows, and it thins and pinches out on a structural high where a well-developed saprolite is present atop Grande Ronde Basalt. These observations are consistent with structurally controlled topography developing throughout the Columbia Plateau by late Grande Ronde time (Reidel, 1978; Beeson and Moran, 1979; Camp and Hooper, 1981; Landon and others, 1982).

The ancestral Columbia River apparently flowed southwest across the plateau and through the Cascade Range to the general vicinity of present-day Mount Hood (Tolan and Beeson, 1984). (See Fig. 4.) The main stream channel followed the southwest-trending Mount Hood syncline into western Oregon.

Stream drainage on the plateau was abruptly changed by renewed basalt volcanism. Wanapum Basalt erupted from fissure systems in the eastern Columbia Plateau in the middle Miocene (from about 15.5 to 13.6 m.y. ago) (McKee and others, 1977). The Frenchman Springs, Roza, and Priest Rapids Mem-

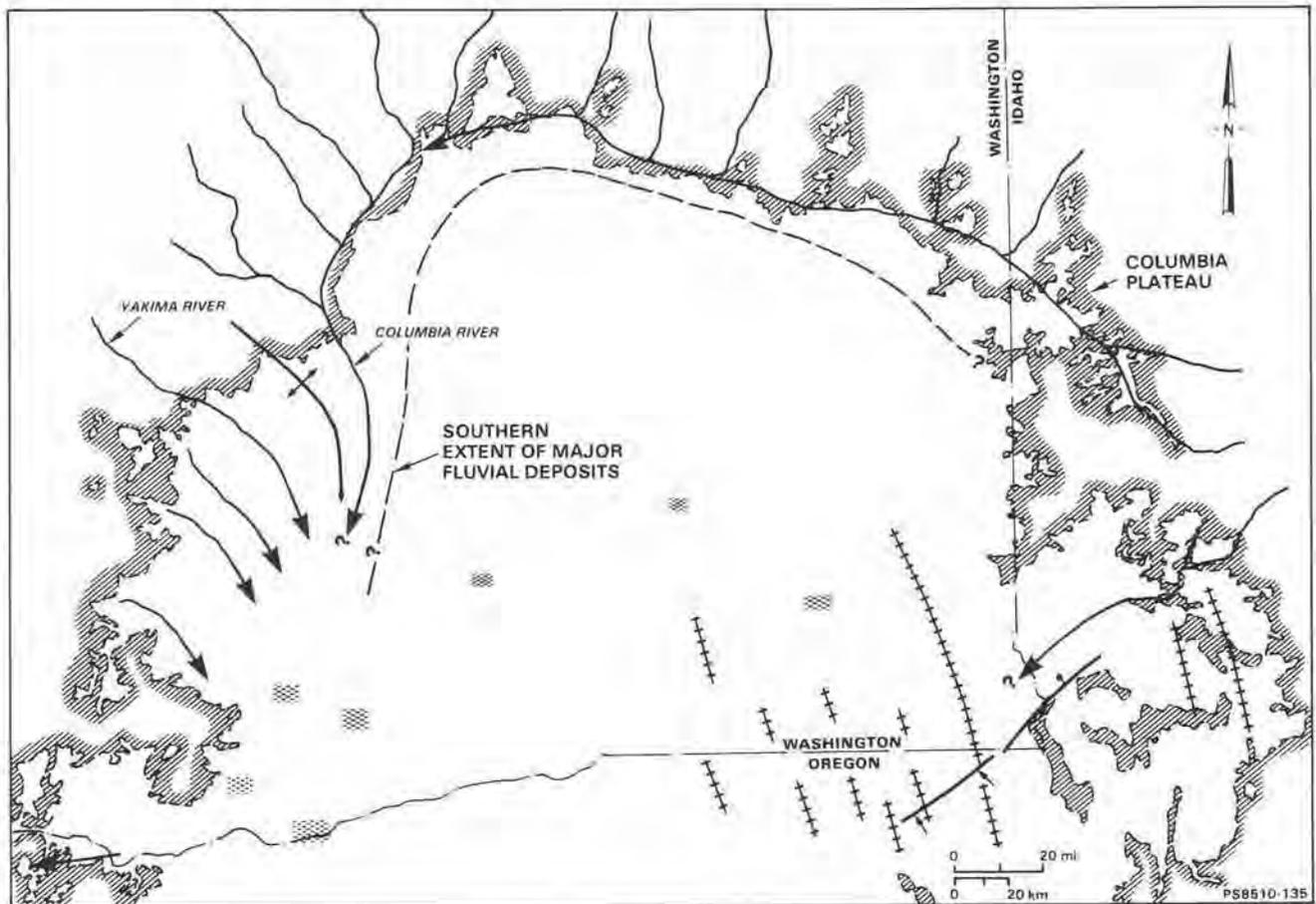


Figure 5.—Stream drainage on the northern part of the Columbia Plateau during middle Miocene (Wanapum) time.

bers of the Wanapum Basalt form much of the surface of the present plateau. Emplacement of the Wanapum flows continued to build the constructional high of the westerly dipping Palouse Slope. However, the central plateau continued to subside in the middle Miocene, as is evident from the ponding of flows in structural lows (Reidel and others, 1980; Reidel and Fecht, 1981).

Initial eruptions of Frenchman Springs lava engulfed the ancestral Columbia River and ancient lake(s) in the western plateau. The lowermost flows, termed Ginkgo flows (Mackin, 1961; Bentley, 1977), contain thick hyaloclastites and pillow basalts at their bases that are areally extensive from east of Wenatchee southwestward to the Columbia River Gorge (Mackin, 1961; Lefebvre, 1966; Diery, 1967; Bentley and others, 1980). The emplacement of Ginkgo flows apparently restricted a displaced ancestral Columbia River to a course in the western plateau (Fig. 5).

Overlying the Frenchman Springs Member in the Yakima and Vantage areas is the Squaw Creek Member of the Ellensburg Formation (Swanson and others,

1979a). The member consists of diatomite in the Vantage area and grades westward into sandstones and conglomerates composed of volcanoclastic, plutonic, and metamorphic detritus (Mackin, 1961). Mackin interprets the diatomite facies of the Squaw Creek Member as the accumulation of diatoms in a lake impounded by the uppermost flow of the Frenchman Springs Member in the Vantage area.

Eruption of the Roza and Priest Rapids Members of the Wanapum Basalt apparently did not significantly affect stream courses on the western Columbia Plateau in Washington. The western margins of the Roza and Priest Rapids flows apparently continued to restrict the south-flowing segment of ancestral Columbia River and east-flowing tributaries from the ancestral Cascade Range to the western plateau. In the northern and northeastern plateau, the Priest Rapids flows encroached on the plateau margin and inundated the St. Joe and St. Maries Rivers in the Benewah Embayment. These flows also filled older channels cut into the Latah Formation and Grande Ronde Basalt

near Spokane (Swanson and others, 1979a). Farther southwest near Malden, the basal Priest Rapids flow is pillowed over a wide area (Griggs, 1976), suggesting that the lava dammed and inundated a local drainage system whose streams flowed westward onto the eastern margin of the plateau. However, the ancestral Columbia River and its tributaries in the northeastern part of the Columbia Plateau re-established courses along the northern plateau margin.

In the southeast Columbia Plateau, the effects of emplacement of Wanapum or Grande Ronde flows on drainage (that is, the ancestral Clearwater and Salmon Rivers) from the Clearwater Embayment are not accurately known. However, the absence of interbedded sediments and erosional features associated with Grande Ronde and Wanapum flows on the Palouse Slope on the north side of the emerging Blue Mountains suggests streams draining the Clearwater Embayment may have been flowing south during middle Miocene time.

Middle to Late Miocene

During middle to late Miocene time, Columbia River Basalt Group volcanism was waning, and flows of the Saddle Mountains Basalt were areally restricted to former river canyons and valleys in the eastern plateau, the subsiding central part of the plateau, and river canyons through the ancestral Cascade Range (Myers and Price and others, 1979; Swanson and others, 1980; Waters, 1973; Tolan, 1982; Anderson and Vogt, this volume; Anderson and others, this volume). In the period of less frequent basaltic eruptions, a relatively well-integrated river system developed and deposited sediments over extensive parts of the central and western plateau (Mason, 1953; Laval, 1956; Schmincke, 1964). Deformation within the Yakima fold belt of the western plateau, which was largely masked by earlier frequent and voluminous volcanic eruptions, formed subtle topographic features of sufficient relief to control the distribution of Ellensburg sediments and Saddle Mountains flows (Reidel and others, 1980; Reidel and Fecht, 1981). The subsidence of the central plateau, which was initiated at least by early Miocene time, formed slopes gently dipping toward the structural and topographic low of the central plateau. These slopes also controlled the distribution and thickness of sediments and basalt flows.

As the central plateau was subsiding, the ancestral Columbia River began entrenching itself in about its present position along the north and northeast margin of the plateau (Fig. 6). South of Wenatchee, its course shifted from southwesterly along the fronts of Wanapum flows to more southerly as a result of contin-

ued uplift of the Hog Ranch structure. The river flowed south down the forming Columbia Slope (Schmincke 1964, 1967a) and through the structural low in the Saddle Mountains at Sentinel Gap. Laterally, the river was controlled by the Hog Ranch structure to the west and the Palouse Slope to the east. The shift appears to have preceded the eruption of the Saddle Mountains flows, based on the abundance of arkosic sediments in the Mabton Member of the Ellensburg Formation and in and near the Pasco Basin (Myers and Price and others, 1979; Reidel and Fecht, 1981).

On the western Columbia Plateau, the Yakima River and its tributaries began forming an integrated drainage system after the position of the Columbia River was established east of the Hog Ranch structure. East-flowing streams in the Yakima River drainage developed as they cut through the constructional topography of basalt flows on the northwestern plateau. Ellensburg sediments were deposited in the developing Kittitas Basin primarily by east-flowing streams carrying volcanoclastic sediments from the ancestral Cascade Range. Between Ellensburg and Yakima, the Yakima River, which had apparently previously meandered over an ancestral lowland, began entrenching a course across the rising structural ridges of the Yakima fold belt. (See Fig. 6.) The Yakima River has been interpreted to be antecedent to the fold belt (Smith, 1903; Calkins, 1905; Waters, 1955). Apparently, the entrenchment resulted from the combined effect of uplift of ridges and subsidence of basins in the western plateau.

The remaining discussion of events in middle to late Miocene time will be primarily directed to the Yakima Valley and Pasco Basin areas, where the emplacement of flows and deformation of the plateau had a pronounced influence on drainage development. The discussion is based on evidence from the intervals between eruptions of Saddle Mountains Basalt. Intervals are used in this paper for convenience rather than formally and informally named interbeds or members of the Ellensburg Formation, primarily because many basalt flows are not as areally extensive as the sedimentary deposits.

Post-Umatilla Interval

Stream drainage on the Columbia Plateau in southern Washington after emplacement of the Umatilla Member is reconstructed from (1) the remnants of the canyon-filling Wilbur Creek Member, (2) the position of the established courses of the Columbia and Yakima Rivers in the northern plateau, and (3) the distribution of Ellensburg sediments along the margin of the Umatilla Member in the Yakima Valley.

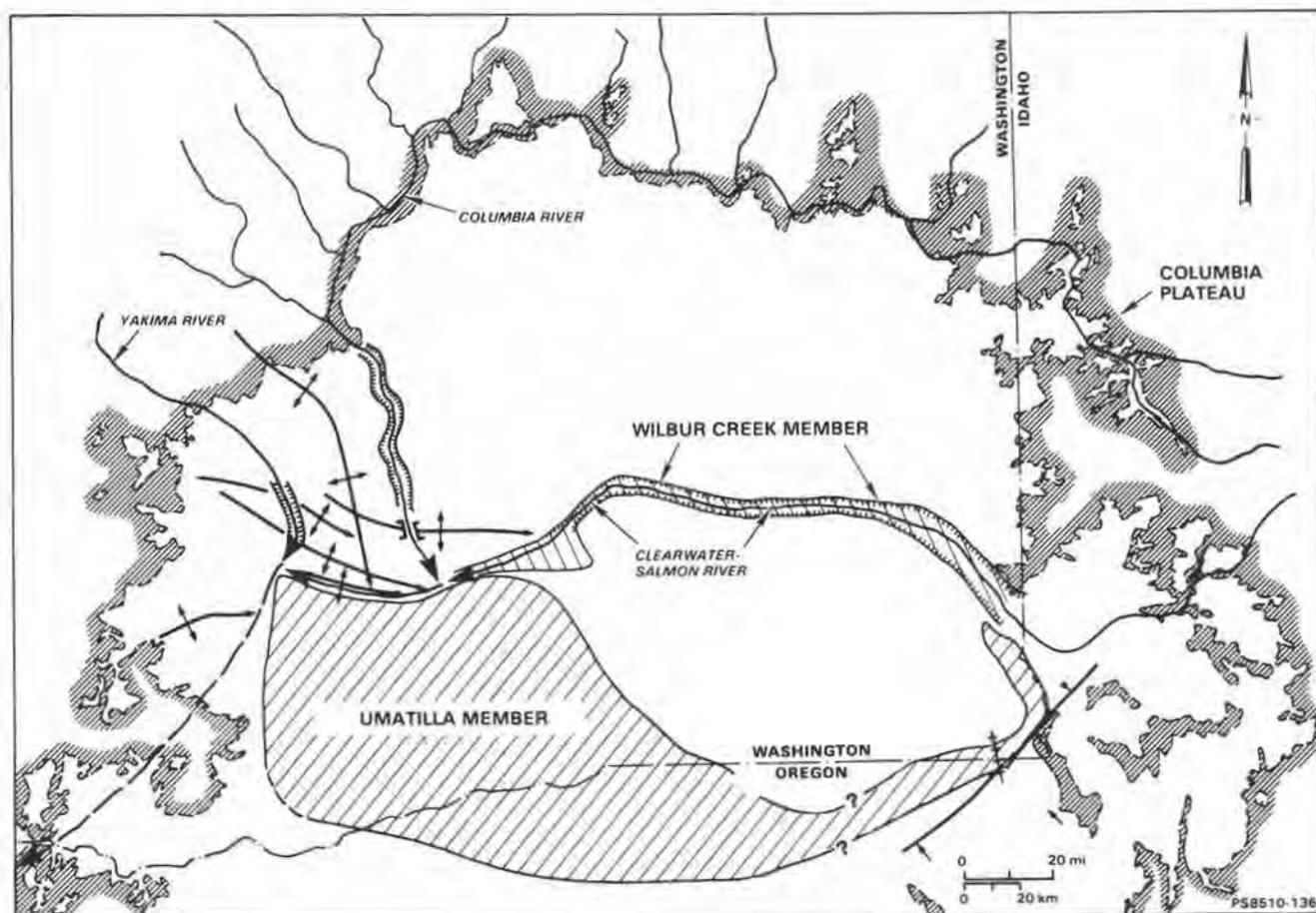


Figure 6.—Stream drainage on the northern part of the Columbia Plateau during middle to late Miocene (post-Umatilla interval) time.

The Wilbur Creek Member eruption was relatively small in volume. Wilbur Creek lava was erupted from fissures in the North Fork area of the Clearwater River in Idaho and flowed down canyons of the ancestral Clearwater drainage into the Lewiston Basin (Camp, 1967, 1981). Stream gravels are present beneath the Wilbur Creek Member in the Clearwater Embayment (Camp, 1981). (See Fig. 6.) Wilbur Creek lava flowed northwest out of the Lewiston Basin and onto the Uniontown Plateau around the rising Blue Mountains, then west down the Palouse Slope, filling the ancestral Clearwater-Salmon River canyon (Swanson and others, 1979a, 1980). At Warden, the intracanyon remnants of the Wilbur Creek Member trend southwest across the northern Pasco Basin and west to Yakima (Bentley, 1977; Myers and Price and others, 1979; Reidel and others, 1980). In the Pasco Basin, this ancestral river course was controlled by the structural low of the developing Wahluke syncline and by the rising Saddle Mountains to the north and approximately 45-m-thick flow front of the Umatilla

Member to the south (Reidel and others, 1980; Reidel and Fecht, 1981). The Columbia River, flowing south through the Saddle Mountains at Sentinel Gap, apparently joined the ancestral Clearwater-Salmon River in an area southeast of Priest Rapids Dam. (See Figs. 1 and 6.) The course of the combined rivers between the Pasco Basin and Yakima continued to be controlled on the south by the flow front of the Umatilla Member and to the north by the rising Yakima Ridge.

In the Yakima area the combined Columbia/Clearwater-Salmon River joined the ancestral Yakima River to flow southwest, possibly through the Grayback Mountain paleocanyon of Anderson and Vogt (this volume) to the Columbia River Gorge area where the streams occupied the Bridal Veil channel of Tolan and Beeson (1984). The southwesterly stream course was generally controlled by the western margin of the Umatilla Member to the east and the syncline of the northeast-southwest-trending Yakima folds of this area.

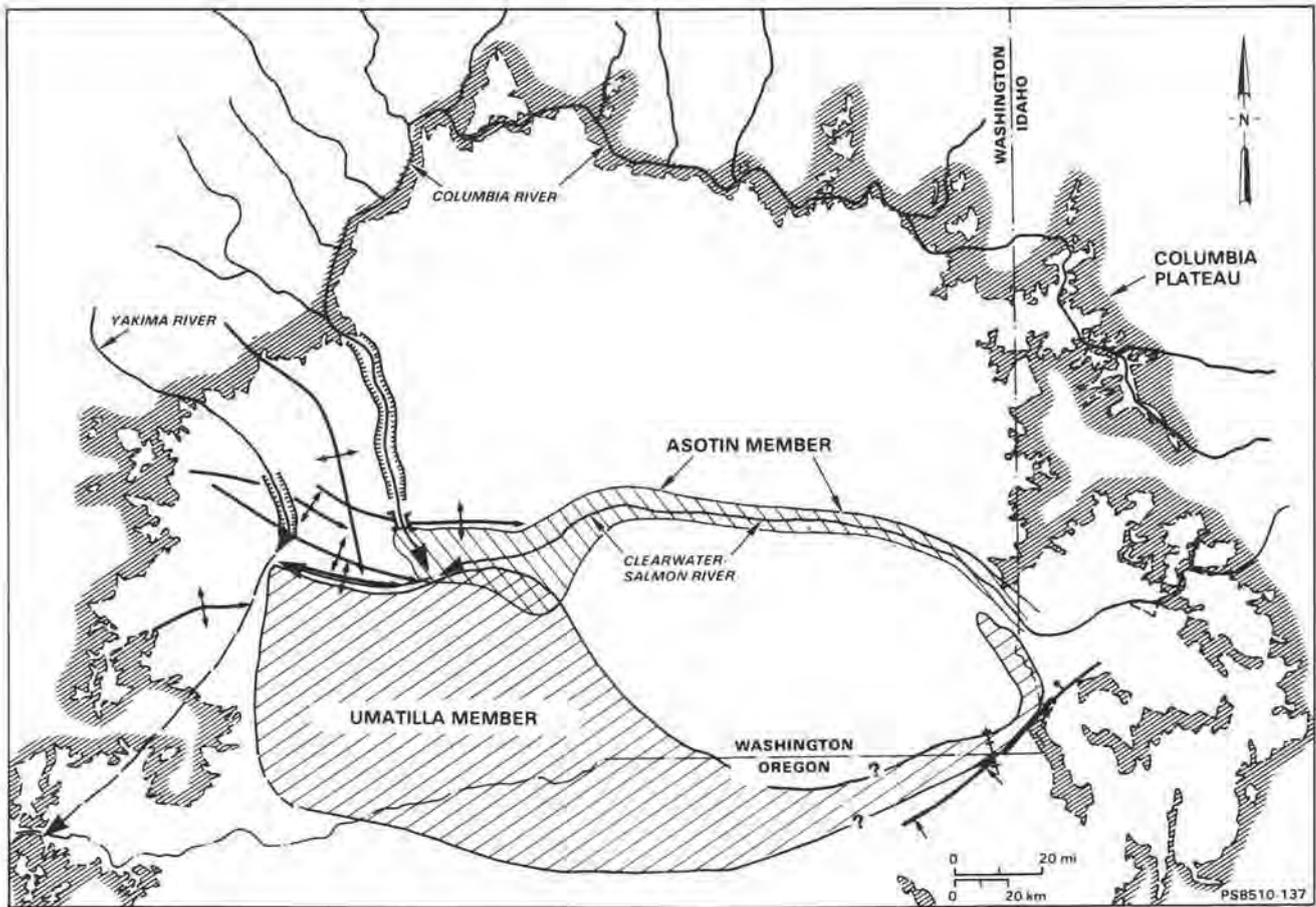


Figure 7.—Stream drainage on the northern part of the Columbia Plateau during middle to late Miocene (post-Wilbur Creek interval) time.

Post-Wilbur Creek Interval

The position of major stream courses during the post-Wilbur Creek interval (Fig. 7) is very similar to that of the post-Umatilla interval. (See Fig. 6.) The similarity is due to (1) lack of a sufficient volume of lava in the Wilbur Creek eruption to disrupt drainage development, (2) continuation of structural and constructional controls, and (3) the short period between the two intervals.

Stream courses during the post-Wilbur Creek interval are mainly defined by the canyon-filling and minor valley-filling flows of the Asotin Member (Swanson and others, 1979a, 1980; Myers and Price and others, 1979; Reidel and others, 1980). The Asotin Member erupted from a source area in the Clearwater drainage (Camp, 1981) and flowed down the ancestral Clearwater-Salmon River canyon from the Lewiston Basin across the Palouse Slope to the Pasco Basin. (See Fig. 7.) In the Lewiston and Pasco Basins, the Asotin lava overflowed the river channel and spread

laterally, partially filling the basins (Camp, 1981; Reidel and others, 1980; Reidel and Fecht, 1981). In the Pasco Basin the thickest section (about 67 m) of Asotin basalt is interpreted to correspond with the former position of the ancestral Clearwater-Salmon River, although no extensive sedimentary deposits are known to be associated with the post-Wilbur Creek interval.

The courses of the Columbia, Clearwater-Salmon, and Yakima Rivers from the central plateau into the present-day Columbia River Gorge area remained essentially unchanged from the post-Umatilla interval. (See Fig. 7.)

Post-Asotin Interval

Following the eruption of the Asotin Member, the ancestral Clearwater-Salmon River re-established a course on the Palouse Slope about 25 km south (Fig. 8) of the river canyons formed during the post-Umatilla and post-Wilbur Creek intervals. The river incised

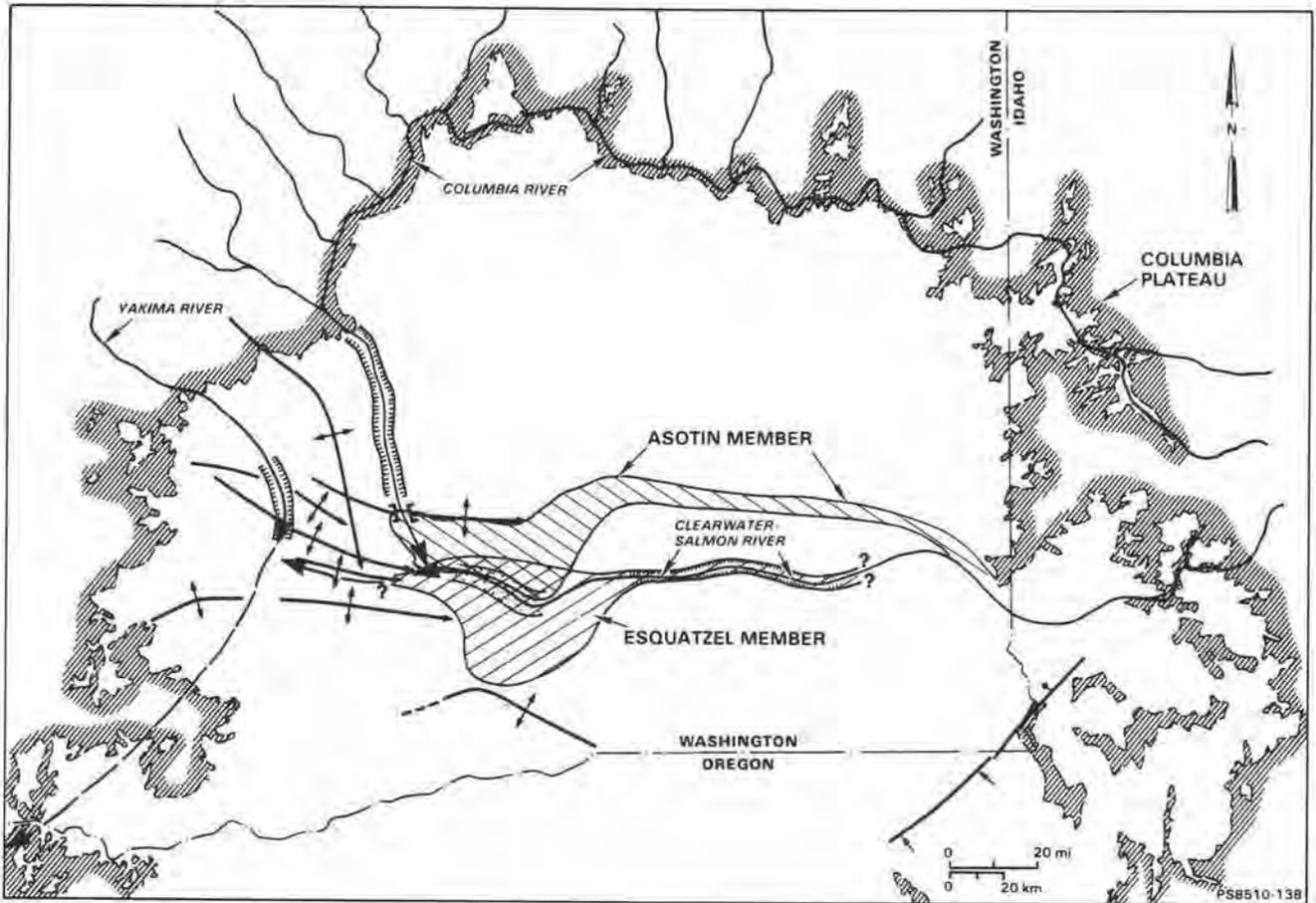


Figure 8.—Stream drainage on the northern part of the Columbia Plateau during middle to late Miocene (post-Asotin interval) time.

a canyon into the Palouse Slope that is now preserved by the Esquatzel intracanyon flow (Swanson and others, 1980). The new course was near the present-day Snake River (Swanson and others, 1980).

The course of the ancestral Clearwater-Salmon River, as determined by the distribution of the Esquatzel Member, loses definition where Esquatzel lavas spread laterally and ponded in the subsiding Pasco Basin. However, the course can be defined by clastic sediments underlying the Esquatzel Member. These form part of the Cold Creek interbed of the Ellensburg Formation and are divided into two distinct textural facies: (1) a fine-grained, tuffaceous and arkosic sandstone and mudstone facies and (2) a coarse-grained sandstone and conglomerate facies (Myers and Price and others, 1979; Reidel and Fecht, 1981). The coarse-grained facies follows an arcuate trend across the basin and is subparallel to the flow front of the Asotin Member. The coarse-grained facies is interpreted to represent main-channel deposits of the ancestral Clearwater-Salmon River; the fine-grained facies

mainly represents overbank and early post-Asotin fluvial deposits. The course of the ancestral Clearwater-Salmon River in the Pasco Basin was controlled by the topographic and structural low along the southern margin of the Asotin flow front.

At the western margin of the Pasco Basin, the ancestral Clearwater-Salmon and Columbia Rivers joined in a poorly defined area, probably near the east end of the present-day Yakima Ridge, to flow west out of the basin. The course of the combined rivers to the west is defined by remnants of the Esquatzel Member, which at least partially filled the eastern part of Black Rock Valley between the rising Yakima ridge and Rattlesnake Hills (Myers and Price and others, 1979).

Esquatzel flows have not been mapped west of about Long. 120° W., but because the Yakima Ridge and Rattlesnake Hills were emerging, the combined course of the Columbia/Clearwater-Salmon Rivers was probably confined between these basaltic ridges to near Yakima. At Yakima, these combined rivers joined with

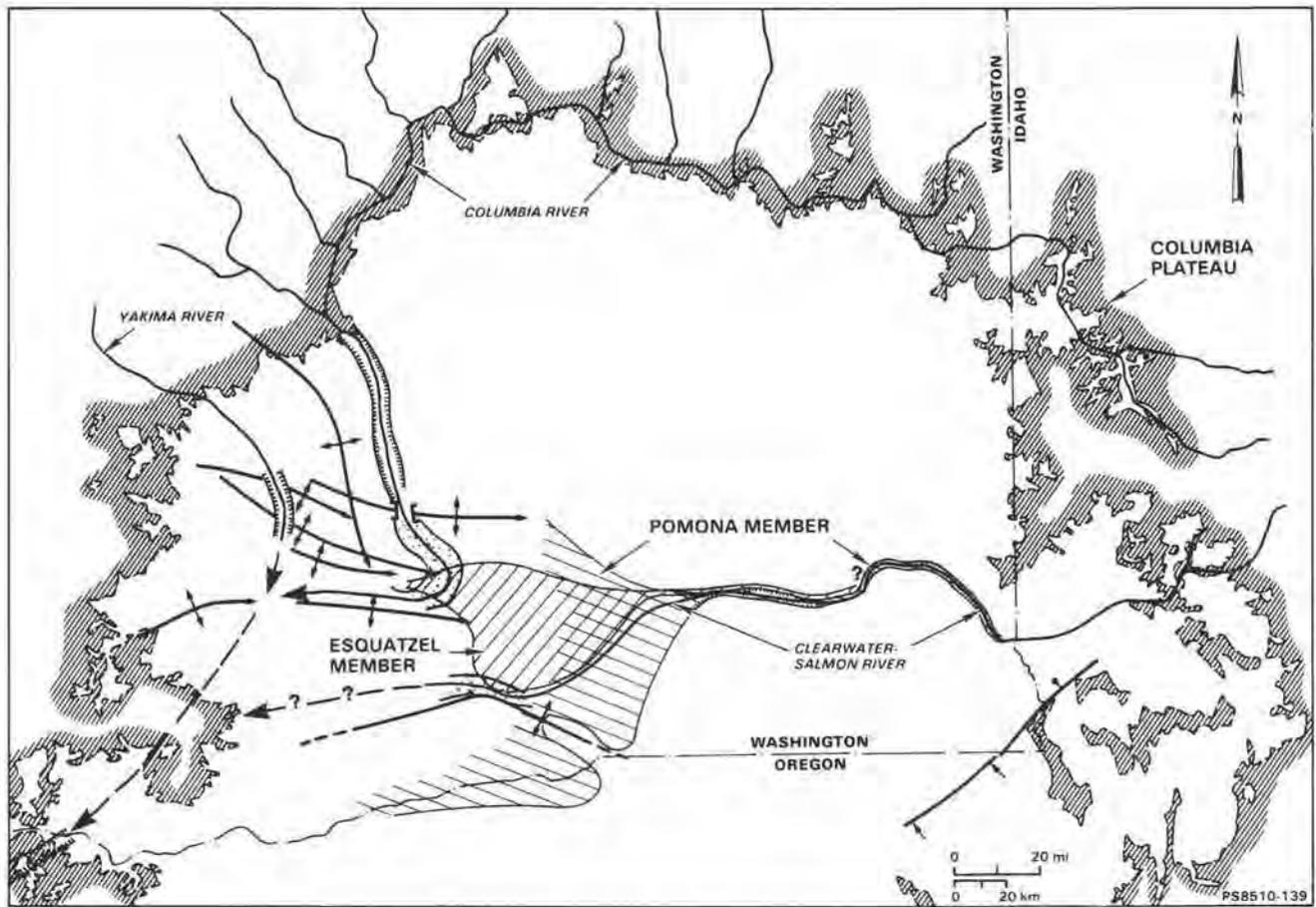


Figure 9.—Stream drainage on the northern part of the Columbia Plateau during middle to late Miocene (post-Esquatzel interval) time. Only the eastern part of the Pomona Member is shown.

the ancestral Yakima River to flow southwest to the present-day Columbia River Gorge area in a course similar to that of earlier intervals. (See Fig. 8.)

Post-Esquatzel Interval

The course of the ancestral Clearwater-Salmon River during the post-Esquatzel interval is defined by the canyon-filling Pomona Member (Swanson and others, 1979a, 1980). The position of the river was essentially coincident with its position during the post-Asotin interval (Fig. 9). In the Pasco Basin, Pomona lavas spread laterally and inundated the basin. Beneath the Pomona Member there, extensive sedimentary deposits form the Selah Member of the Ellensburg Formation (Myers and Price and others, 1979). This member is composed of volcanoclastic and arkosic sediments that range in size from clay to gravel. The Selah gravels are known in three areas of the Pasco Basin and help define the courses of main through-flowing streams. In the northwest part of the basin, a

gravel train can be defined from Sentinel Gap in the Saddle Mountains, southeast around the east end of Umtanum Ridge, and then west over the east end of Yakima Ridge. In this area, the Selah gravels are mainly composed of basaltic, plutonic, and metamorphic clasts. In the southwestern part of the Pasco Basin, Selah gravels are exposed on the north flank of the Horse Heaven Hills and lie on the Umatilla Member and beneath a thick (>60 m) Pomona Member (Bond and others, 1978). These gravels are mainly composed of basaltic clasts. The small nonbasaltic fraction consists of a variety of metamorphic and plutonic clasts, including a few metabasalt clasts. In the southeastern Pasco Basin, an outcrop of Selah gravels is exposed immediately northeast of Wallula Gap (Gardner and others, 1981). These gravels also lie between the Umatilla and Pomona Members and are composed only of basaltic clasts.

The differences in the compositions of the three Selah gravel deposits suggest deposition by three separate streams. The distribution of the northern gravel

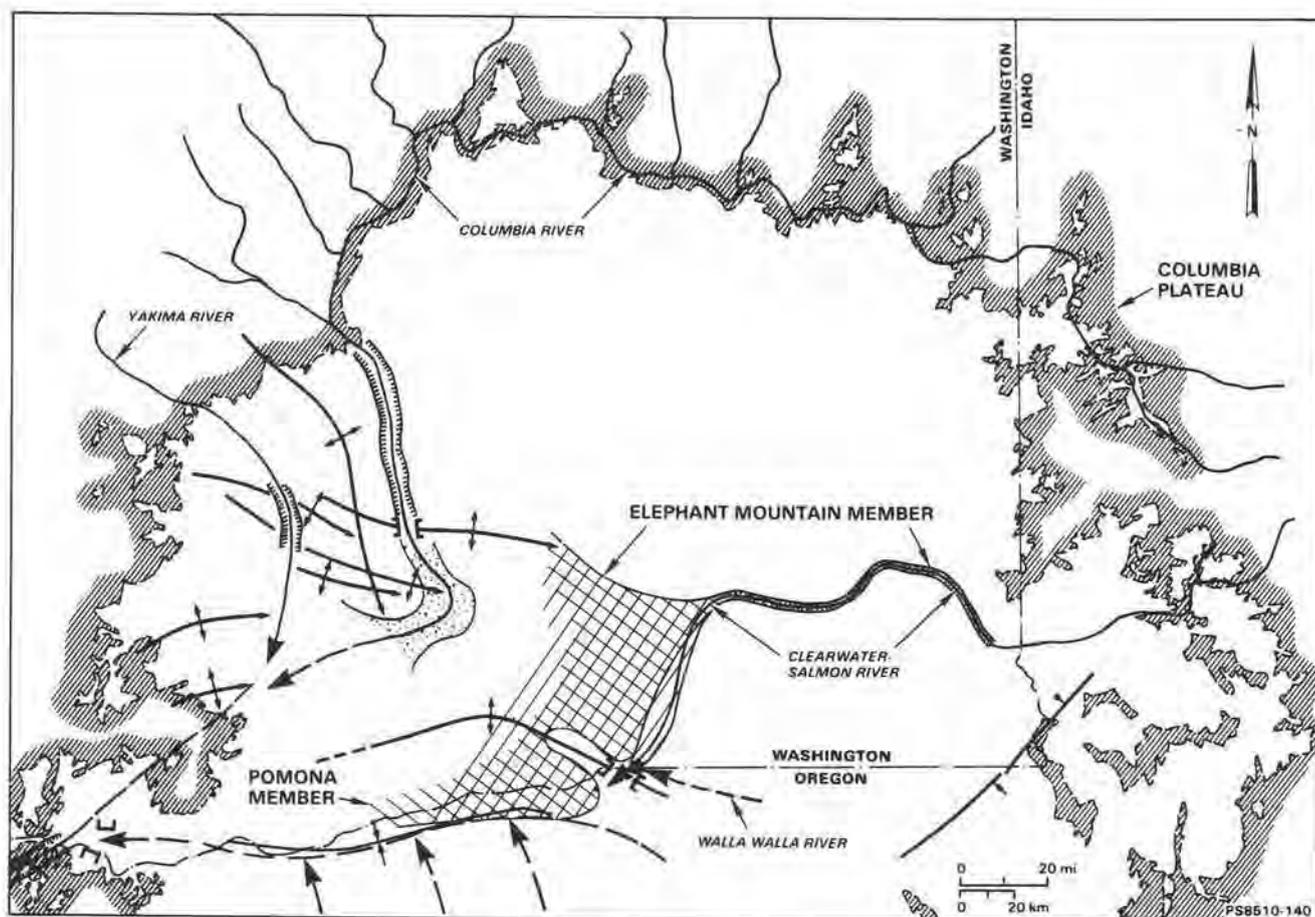


Figure 10.—Stream drainage on the northern part of the Columbia Plateau during middle to late Miocene (post-Pomona interval) time. Only the eastern part of the Elephant Mountain Member is shown.

train corresponds closely to former courses of the ancestral Columbia River during middle to late Miocene time. (See Figs. 7-9.) The composition of these northern gravels also resembles that of other deposits of the ancestral Columbia River, which, during this interval, flowed in essentially the same course as during the post-Asotin interval. This course was probably controlled by the northern margin of the Esquatzel Member.

Gravel clast lithologies in the southwestern Pasco Basin suggest deposition by the ancestral Clearwater-Salmon River. That river is believed to have flowed from the Devil's Canyon area southwest across the southern Pasco Basin, through the Chandler area, and west into the lower Yakima Valley. In the Pasco Basin, the ancestral Clearwater-Salmon River was forced south to a position apparently controlled by the southern flow front of the Esquatzel Member and, in the Chandler area, by the rising Horse Heaven Hills.

The basaltic Selah gravels of the southeastern Pasco Basin are suggestive of a stream that originated

on the Columbia Plateau. We speculate that these gravels represent the ancestral Walla Walla River, which flowed west along the front of the Horse Heaven Hills or along the series of rising brachyanticlines that now form the southeastern part of the Rattlesnake Hills.

Post-Pomona Interval

The ancestral Columbia and Clearwater-Salmon Rivers that joined in the Pasco Basin in the earlier intervals of middle to late Miocene time appear to have separated during the post-Pomona interval. The course of the ancestral Columbia River through the Pasco Basin during this interval is well defined by a conglomerate preserved in the Rattlesnake Ridge Member of the Ellensburg Formation (Fig. 10). The conglomerate is present from Sentinel Gap southeast around the east end of Umtanum Ridge and southwest across the east end of Yakima Ridge. It is flanked by fining-upward cycles of sandstone and mudstone that

thin and grade southeast to mainly sandy mudstone and mudstone on the margin of the Pasco Basin.

Southwest of the Pasco Basin, the course of the ancestral Columbia River shifted widely across the Rattlesnake Hills, as indicated by the widespread occurrences of conglomerate mapped in the Rattlesnake Ridge interbed by Schmincke (1967a). The ancestral Columbia River joined the Yakima River in the present upper Yakima Valley between about Union Gap and Toppenish Ridge and then flowed across the eastern part of the Yakima Indian Reservation (Bentley and others, 1980). The combined course of the rivers apparently passed southwest in the paleocanyon described by Anderson and Vogt (this volume) and into the Bridal Veil channel of Tolan and Benson (1984). These paleocanyons have been partially filled by Pomona lava.

The main course of the ancestral Clearwater-Salmon River during the post-Pomona interval has been only tentatively defined in the central Columbia Plateau. Based on examination of borehole samples and surface exposures of the Rattlesnake Ridge interbed in the Pasco Basin, no evidence has been found to establish the ancestral Clearwater-Salmon River from the Palouse Slope westward across the Pasco Basin. Gravel clasts of the ancestral Columbia River west of the Pasco Basin were also examined to determine if the ancestral Clearwater-Salmon River may have joined the ancestral Columbia in the central plateau of Washington State. Key rock types of the ancestral Clearwater-Salmon River sediments (for example, metabasalts) are absent, suggesting that the ancestral Clearwater-Salmon River did not flow across the central plateau. We believe that the course of the ancestral Clearwater-Salmon River may have been diverted south of the Pasco Basin through Horse Heaven Hills, probably at Wallula Gap.

Evidence for the course of the Clearwater-Salmon River through Wallula Gap is, in part, the occurrence of an unnamed conglomerate unit exposed near the crest of the Horse Heaven Hills on the west side of the gap. The unit is a mixture of basaltic, metamorphic (including metabasalts), and plutonic clasts in proportions similar to those of main channel gravel deposits in the ancestral Clearwater-Salmon River drainage on the eastern Columbia Plateau. At Wallula Gap, this unnamed interbed of the Ellensburg Formation partially fills a canyon eroded into the Umatilla Member and is capped by the Ice Harbor Member (basalt of Martindale) of the Saddle Mountains Basalt (ARHCO, 1976). The age of the gravels is not well constrained. Field mapping in the Wallula Gap area on the north flank of the Horse Heaven Hills has not revealed the exotic gravels that occur in interbeds

beneath the Ice Harbor, Elephant Mountain, or Pomona Members at other localities (Gardner and others, 1981). However, gravels in the same stratigraphic position and of similar composition occur in an exposure 10 km northeast of Wallula Gap along the present-day Walla Walla River. The Selah Member immediately northeast of Wallula Gap is texturally and compositionally different than the unnamed conglomerate. This suggests that the conglomerate is probably post-Selah in age.

If the gravels at Wallula Gap are, in part, of post-Pomona-interval age, then the west-flowing ancestral Clearwater-Salmon River was diverted to a southwesterly course along or very near the flow front of the Pomona Member in the southeastern Pasco Basin. (See Fig. 10.) Diversion of the river across the Horse Heaven Hills at Wallula Gap may have been in response to emplacement of the Pomona Member. The course through the Horse Heaven Hills was apparently controlled by the southern flow front of the Pomona Member on the west side of the gap and by a structural low in the Horse Heaven Hills at the gap (Gardner and others, 1981).

Southwest of Wallula Gap, the ancestral Clearwater-Salmon River probably flowed southwest into the incipient Umatilla Basin, where the river was captured by a river system draining north-central Oregon. In the Umatilla Basin, the major river was probably directed in the forming syncline south of the southern margin of the Pomona Member. The position of the course is based on the occurrence of fine sediments in the Rattlesnake Ridge interbed in the Umatilla Basin (Shannon and Wilson, 1972, 1975). In that basin, no coarse clastics associated with the river system of the post-Pomona interval are known to include rock types diagnostic of the ancestral Clearwater-Salmon River. The lack of these rock types may, however, be due to erosion and reworking of fluvial sediments in the area over the past 10.5 to 12 m.y. The major river of the north-central Oregon system is assumed to have flowed west, parallel to the axis of the forming Umatilla-Dalles syncline and through the Columbia Hills at the present-day Columbia River water gap. The water gap is interpreted to have been the site of an active stream through the Columbia Hills since at least the beginning of middle Miocene time. This interpretation is based on the absence of erosional features, intracanyon flows, and coarse clastics of late Miocene age elsewhere across the Columbia Hills. The ancestral river joined the ancestral Columbia-Yakima River between the Columbia Hills water gap and Hood River, Oregon, to flow through the ancestral Cascade Range.

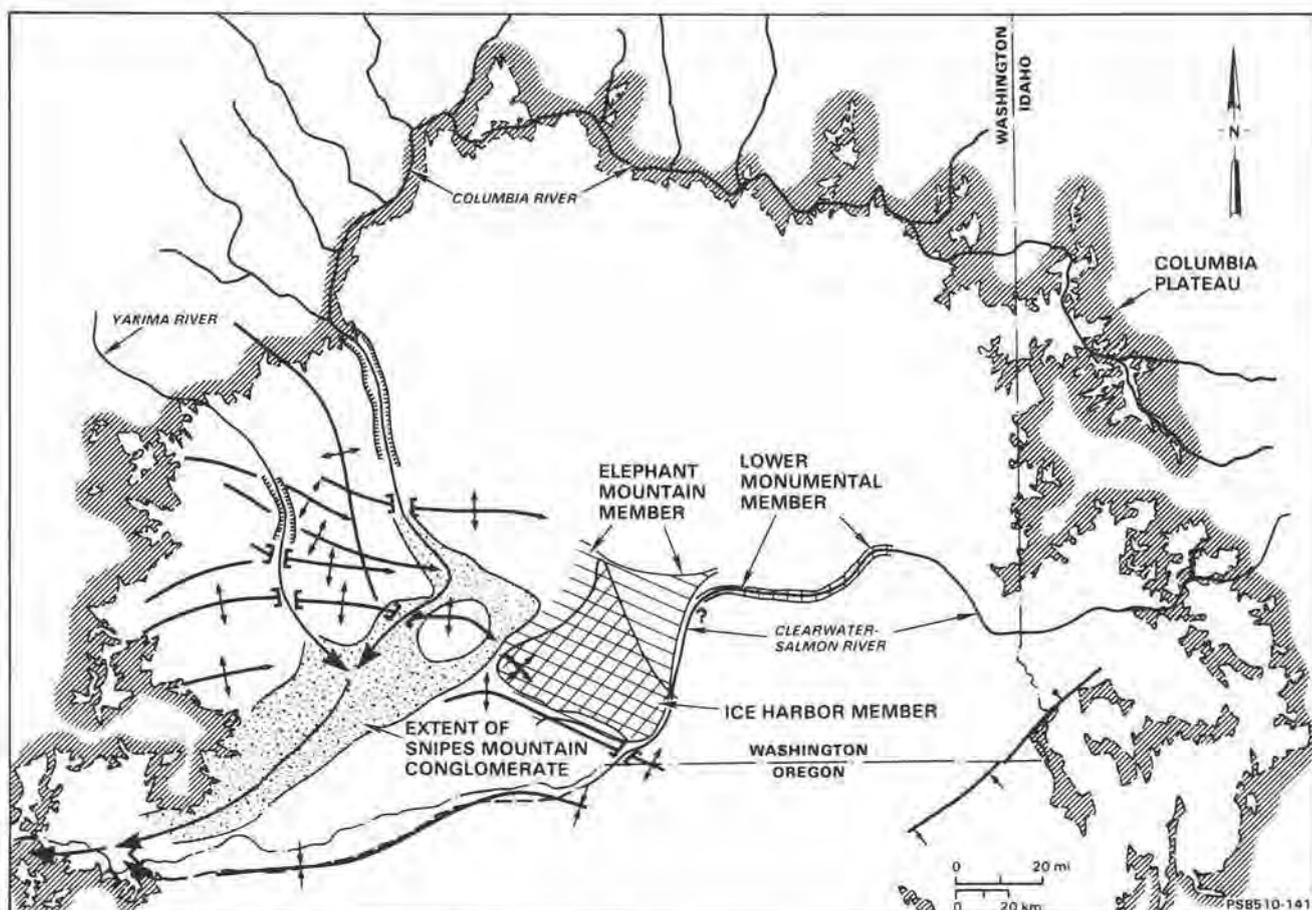


Figure 11.—Stream drainage on the northern part of the Columbia Plateau during middle to late Miocene (post-Elephant Mountain interval, early Snipes Mountain) time. Only the central part of the Elephant Mountain Member is shown.

Post-Elephant Mountain Interval

Following the Elephant Mountain volcanic episode when flows inundated much of the central Columbia Plateau, the major ancestral rivers re-established courses in about the same positions as in the post-Pomona interval.

The presence of the ancestral Columbia River in the post-Elephant Mountain interval is defined by the widespread occurrence of stream gravels that overlie the Elephant Mountain Member from the Sentinel Gap area southwest to the Columbia River Gorge. The gravels include the quartzite gravels of Warren (1941), Snipes Mountain Conglomerate of Schmincke (1967a), and "gravels of the ancestral Columbia River" of Campbell (1979) and Rigby and others (1979). These gravels are composed of quartzitic, plutonic, basaltic, and other metamorphic clasts. Russell (1893) and Waring (1913) first described the occurrence of quartzitic pebbles in the southwestern Columbia Plateau. Warren (1941) suggested that the gravels

marked the former course of the Columbia River from Sentinel Gap southwest through Satus Pass and into the Columbia River Gorge. Waters (1955) later demonstrated that Satus Pass resulted from local stream capture and not from a course incised by an ancestral Columbia River. The relatively broad course of the ancestral Columbia River south and east of Satus Pass in the Goldendale area was delineated by Schmincke (1967a) and Rigby and others (1979), based on the widespread occurrence of quartzitic gravels.

The general course of the ancestral Columbia River during the post-Elephant Mountain interval can be defined by the distribution of these quartzitic gravels from Sentinel Gap southeast around the east end of Umtanum Ridge and southwest in a broad path over the Rattlesnake Hills to Snipes Mountain (Fig. 11). The course of the river over the central Columbia Plateau was not controlled by flow margins as in previous intervals. Near the center of the path of quartzitic gravels, an erosional canyon has been cut through a structural low across the Rattlesnake Hills.

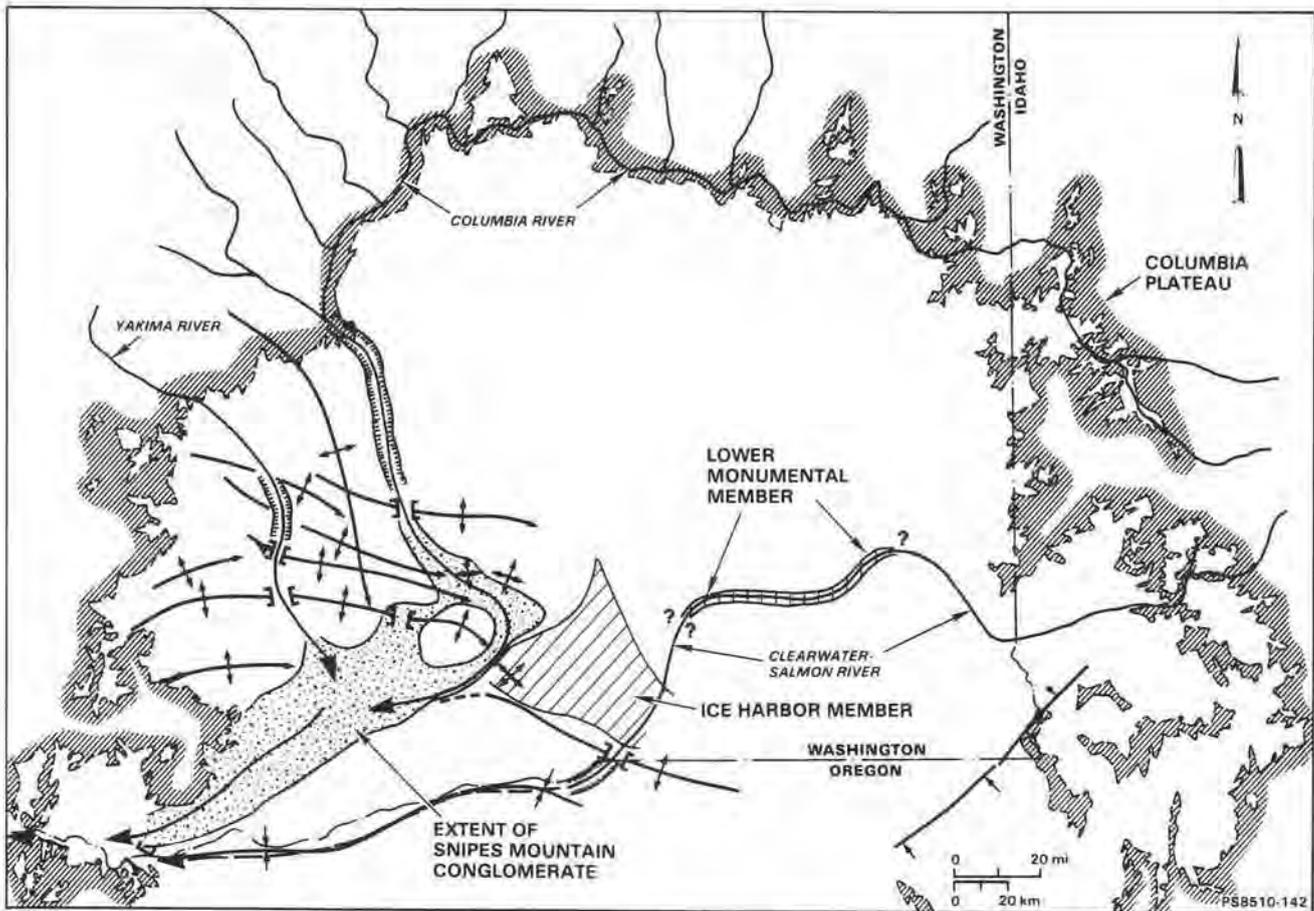


Figure 12.—Stream drainage on the northern part of the Columbia Plateau during late Miocene to middle Pliocene (late Snipes Mountain, basal Ringold) time.

We speculate that the ancestral Columbia River may have cut the Sunnyside Gap during uplift of the hills after emplacement of the Elephant Mountain Member. (See Fig. 11.)

During the post-Elephant Mountain interval, the ancestral Yakima River was flowing south from Yakima through Union Gap in Ahtanum Ridge and southeast into the upper Yakima Valley (Toppenish Basin). (See Fig. 11.) Sediments deposited by the ancestral Yakima River there are mainly composed of volcanic debris. Immediately south of Snipes Mountain, the gravel train contains a mixture of volcanic, plutonic, and metamorphic rock clasts, which represents the combined rivers flowing southwest toward Goldendale and into the present-day Columbia River Gorge area along a course described by Warren (1941). This is consistent with the southwest-trending current directions measured by Schmincke (1964) and the mixed lithologies of the gravels.

The west-flowing ancestral Clearwater-Salmon River on the Palouse Slope flowed southwest along or very near the front of the Elephant Mountain flows to

Wallula Gap. (See Fig. 11.) The course through the Horse Heaven Hills was apparently controlled by the flow front of the Elephant Mountain Member on the west side of the gap and by a structural low in those hills at the gap (Gardner and others, 1981). South of the Horse Heaven Hills, a major stream flowed in a course similar to that of the post-Pomona interval.

Late Miocene to Middle Pliocene

The ancestral Clearwater-Salmon River apparently continued to flow through Wallula Gap during the late Miocene to middle Pliocene (Fig. 12), although sediments indicative of its course have not been found on the Palouse Slope. Gravel clasts indicative of the ancestral river are rarely found in the Snipes Mountain Conglomerate. The absence of these clasts in the conglomerate suggests that the course of the river was probably maintained across the southeastern Pasco Basin through Wallula Gap. The course was probably controlled by the constructional topography built by the Ice Harbor Member (Myers and Price and others, 1979; Gardner and others, 1981).

The course of the ancestral Columbia River was restricted to the structural lows of the Pasco Basin and lower Yakima Valley and was controlled by the structural topography of rising ridges and constructional topography of the Ice Harbor Member. The southwesterly course of this river across the Yakima Ridge and Rattlesnake Hills was apparently diverted east by combined uplift of these ridges and subsidence of the Pasco Basin. In contrast to its position in the post-Elephant Mountain interval (Fig. 11), the new course of the river was around the east end of Umtanum Ridge and into the central Pasco Basin. (See Fig. 12.) Extensive gravel deposits on the Elephant Mountain Member have been penetrated in drill holes in the central Pasco Basin (Fecht and Lillie, 1982) and mark the course of the river there. These gravels are more than 30 m thick (Brown and Brown, 1961; Routson and Fecht, 1979) and are lithologically similar to gravels found along the former course of the ancestral Columbia River across the Yakima Ridge and the Rattlesnake Hills. These gravels form the basal unit of the Ringold Formation (Routson and Fecht, 1979; Tallman and others, 1979; Myers and Price and others, 1979).

In the central Pasco Basin, the basal Ringold gravels essentially terminate north of the Ice Harbor Member margin. (See Fig. 12.) In the southern Pasco Basin, the basal Ringold gravels are only locally present and attain a maximum thickness of about 2 m (Brown, 1979). Brown and Brown (1961) interpreted the gravels as, in part, equivalent to the Levey interbed of the Ellensburg Formation that lies beneath the Ice Harbor Member. The Levey consists of fine-grained sedimentary rocks, arkosic sandstones and mudstones. However, the Levey does not coarsen toward the margin of the Ice Harbor flows, as might be expected if interbed and basal Ringold were equivalent. We believe that basal Ringold gravels are not a lateral equivalent of the Levey interbed, but that they were deposited after emplacement of the Ice Harbor Member. The basal Ringold gravels must be younger than the Ice Harbor Member, which has been dated at 8.5 m.y. (McKee and others, 1977) and which marked the end of Columbia River basalt volcanism in the central plateau.

The course of the ancestral Columbia River south from the central Pasco Basin is defined by a series of thin remnant gravel deposits exposed on the southeast slope of Rattlesnake Mountain and on the north flank of the Horse Heaven Hills near Prosser. These gravels overlie the Elephant Mountain Member and are lithologically similar to the quartzitic gravels on nearby ridges and in the central Pasco Basin. The distribution and extent and similarity of composition of these

gravel deposits suggest that the ancestral Columbia River flowed from the central Pasco Basin southwest into the lower Yakima Valley (Brown, 1966; Fecht and others, 1982). (See Fig. 12.)

In the lower Yakima Valley, the ancestral Columbia River still joined the ancestral Yakima River between Sunnyside and the east end of Toppenish Ridge. (See Fig. 12.) The combined course of the Columbia-Yakima Rivers to the Columbia River Gorge area was also similar to that of the previous interval. (See Fig. 11.)

The cause of the shift of the Columbia River to a course through the Horse Heaven Hills at Wallula Gap has long been debated. Warren (1941) thought that uplift of the Horse Heaven Hills defeated the former course of the Columbia and Yakima Rivers and that the courses shifted eastward to a lower path across the ridge at Wallula Gap. Waters (1955) argued that the Columbia River was displaced eastward by accelerated deposition of volcanoclastic sediment from east-flowing streams draining the Cascade Range. Schmincke (1964) observed that quartzite-bearing gravels stratigraphically overlie the volcanoclastic sediments and, therefore, volcanic debris could not have diverted the river eastward. Waters (1955) noted that olivine-rich Simcoe basalt flows near the Simcoe Mountains (Fig. 1) had filled former stream valleys in the Columbia River basalts and the Ellensburg Formation. The widespread presence of the Snipes Mountain Conglomerate over the Horse Heaven Hills and the absence of a canyon that was typical of earlier channels suggest that channel incision by the ancestral Columbia-Yakima River across the hills was interrupted. Whether the Simcoe lavas repeatedly blocked the Columbia-Yakima River drainage across the Horse Heaven Hills and forced a more easterly course needs further study. However, we believe that continued subsidence of the central Columbia Plateau centered in the Pasco Basin and uplift of the Yakima folds, specifically the Horse Heaven Hills, were important factors in the diversion of the Columbia River. Field mapping and borehole studies have demonstrated continued subsidence of the central Columbia Plateau from at least the early Miocene to about the middle Pliocene (Routson and Fecht, 1979; Myers and Price and others, 1979; Reidel and others, 1980; Reidel and Fecht, 1981; Tallman and others, 1981; Puget Sound Power and Light, 1982). Once the ancestral Columbia River overcame the constructional topography of the Ice Harbor Member, it apparently flowed southeast in the trough of the Pasco syncline of Newcomb and others (1972). South of Pasco, the river was captured by the ancestral Clearwater-Salmon River.

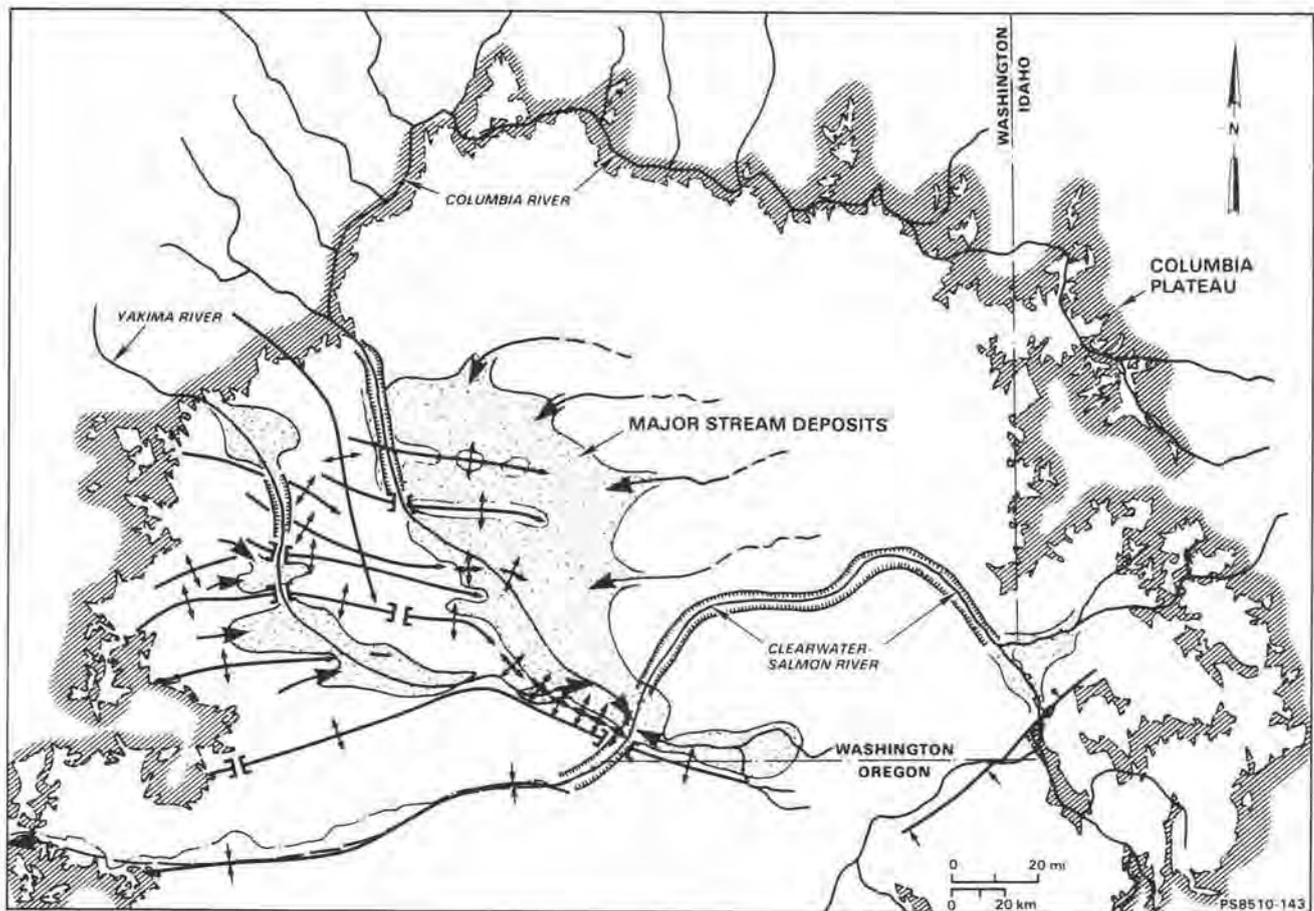


Figure 13.—Stream drainage on the northern part of the Columbia Plateau during late Miocene to middle Pliocene (post-Snipes Mountain) time.

The Yakima River was also diverted from a southwesterly course across the Horse Heaven Hills to a southeasterly course down the lower Yakima Valley and into the Pasco Basin through Badger Coulee (Fig. 13). The diversion of the Yakima River occurred at about the same time as that of the Columbia River, as indicated by the absence of sediments consisting of distinctive volcanoclastic material where the former Columbia-Yakima River crossed the Horse Heaven Hills.

Following the capture of the Columbia and Yakima Rivers, the rivers on the plateau continued to aggrade in the developing synclinal troughs and basins and to scour ever-deepening gorges through rising basaltic ridges. Deposition in structural basins continued on the plateau until about the middle Pliocene. The continued aggradation of the sedimentary sequences is due to a high relative base level that resulted from (1) the absence of Columbia River basalt volcanism forming constructional topography on the plateau (except in the area of the Clearwater Embayment), (2) continued subsidence and folding of

the plateau, (3) uplift of the Cascade Range, and (4) possibly, volcanism in the Columbia River Gorge area that repeatedly choked the river with hyaloclastic debris and lava flows.

In the western Columbia Plateau, the Yakima River and its tributaries deposited the Pliocene Thorp Gravel of Waitt (1979) that unconformably overlies the Ellensburg Formation in the Kittitas Basin. South into the Topenish and Satus Basins, the Yakima River and its tributaries deposited gravels that are present along the basin margin. Here, the Ellensburg Formation is unconformably overlain by these gravel deposits in fluvial terraces and fans. These gravels may be coeval with the Thorp Gravel (Bentley, 1977; Rigby and others, 1979; Campbell, 1983).

In the eastern Columbia Plateau, the ancestral Clearwater and Salmon Rivers deposited the North Lewiston Gravel, Clearwater Gravels, Clarkston Heights Gravel, and perhaps the Clarkston Gravels in the Lewiston Basin (Kuhns, 1980; Waggoner, 1981). On the Palouse Slope, the ancestral Clearwater-Salmon River deposited gravel within a canyon cut

into Wanapum Basalt. The gravels are capped by the Lower Monumental Member of the Saddle Mountains Basalt. (See Fig. 12.) The Walla Walla River and its tributaries deposited an "old gravel and clay" unit containing clasts derived from the Columbia River basalt (Newcomb, 1965).

In the central plateau, the most extensive fluvial deposit of late Miocene to middle Pliocene age is the Ringold Formation (Merriam and Buwalda, 1917). The Columbia River and its tributaries from the Palouse Slope deposited mainly interbedded fine sands, silts, and clays, as much as 165 m thick in the Quincy Basin (Grolier and Bingham, 1978, 1979). In the Pasco Basin the Columbia, ancestral Clearwater-Salmon, and Yakima Rivers deposited nearly 360 m of Ringold sediments—gravel, sand, silt, and clay that can be informally subdivided into four or more units based on texture (Newcomb and others, 1972; Brown, 1979; Tallman and others, 1979; Puget Sound Power and Light, 1982). The upper Ringold unit, exposed at the White Bluffs in the northeast Pasco Basin, is composed of mainly interbedded fine sand and silt that contain microtine rodent fossils 3.7 to 4.8 m.y. old (Repenning, in press). This age range falls in a reversed magnetic polarity event believed to correspond to the Gilbert Magnetic Polarity Epoch (3.4-5.12 m.y. ago) (LaBreque and others, 1977; Mankinen and Dalrymple, 1979). Parts of a Hemphillian (>4.8 m.y.) rhinoceros were identified by Gustafson (1978) in underlying middle Ringold conglomerate. The basal Ringold gravel, previously discussed, is interpreted to be of post-Ice Harbor age (8.5 m.y.). Thus, the Ringold Formation ranges in age from about 3.4 to 8.5 m.y.

Other deposits of the late Miocene to middle Pliocene are, in part, coeval with the Ringold, based on ages and stratigraphic position (Tallman and others, 1982). Fission-track ages from two tephras in the Kittitas Valley are 3.64 ± 0.74 and 3.70 ± 0.20 m.y. (Waite, 1979). Bentley (1977) and Campbell (1983) have interpreted the old gravel deposits in the Yakima Valley to be equivalent to the Thorp Gravel of the Kittitas Basin. The ages of the gravels in the Lewiston Basin and on the Palouse Slope are, in part, constrained by the Lower Monumental Member, which is radiometrically dated at 6.0 m.y. (McKee and others, 1977). Finally, Newcomb (1965) considers the "old gravel and clay" unit in the Walla Walla Basin to be, in part, equivalent to the Ringold Formation, based on similar lithologies. Although the gravel exposures Newcomb interpreted to be equivalent underlie Ice Harbor basalt, the "old gravel and clay" unit and the Ringold Formation are likely, in part, equivalent based on stratigraphic position.

Middle to Late Pliocene

The deposition of Ringold sediments in the Pasco Basin ended rather abruptly, as suggested by the major incision of the Ringold Formation by the Columbia River system, the preservation of a largely unmarred Ringold surface in the eastern and northern basin, and the absence of deposition on the Ringold surface (Newcomb, 1965). The incision is dated between 2.0 ± 0.22 and 3.4 m.y.; the younger age is based on a single thermoluminescence date from the calcrete that caps the Ringold surface on the White Bluffs.

The Thorp and Thorp-like gravels of the western plateau also were deeply incised; this is indicated by remnants of terraces and fans high above the Pleistocene and present-day Kittitas and Yakima Valley floors. Pliocene gravel deposits in the Lewiston Basin also are above the present-day valley floor. We suspect that the incision of Ringold, Thorp, Thorp-like sediments, and possibly the incised gravel deposits in the Lewiston Basin is approximately coeval because of the similar magnitude of the incision.

The cause of the base level change that precipitated these major incisions has been speculated on by earlier workers. Flint (1938b) and Newcomb (1958) suggested that waters of the Columbia River were impounded or perhaps periodically impounded behind the rising Horse Heaven Hills and eventually spilled over and incised a course through the structural and topographic low at Wallula Gap. However, geologic mapping in the gap area by Gardner and others (1981) indicates a relatively slow and continuous rate of uplift of the Horse Heaven Hills since at least middle Miocene time, consistent with uplift rates in the Rattlesnake Hills and Saddle Mountains (Reidel and Focht, 1981). In addition, fluvial sands and silts in the upper Ringold unit of the Pasco Basin that are intercalated with lacustrine facies rule out a totally lacustrine (impounded) origin for the upper unit. We believe that the Columbia, Clearwater-Salmon, and Yakima Rivers were flowing through Wallula Gap essentially continuously (with minor ponding) since river capture in the late Miocene. Therefore, the cause(s) of the major regional base level change must be related to something other than just uplift of Wallula Gap.

Recent work by Tolan and Beeson (1984) in the Columbia River Gorge has demonstrated that the source of the ancestral Columbia River, which is defined by the distribution of the canyon-filling Pomona Member and sedimentary lithology of the Troutdale Formation, was continually being inundated by large volumes of high-alumina basalts and volcanic debris from the Cascades; this aggraded the Columbia River canyon and forced the river to a more northerly

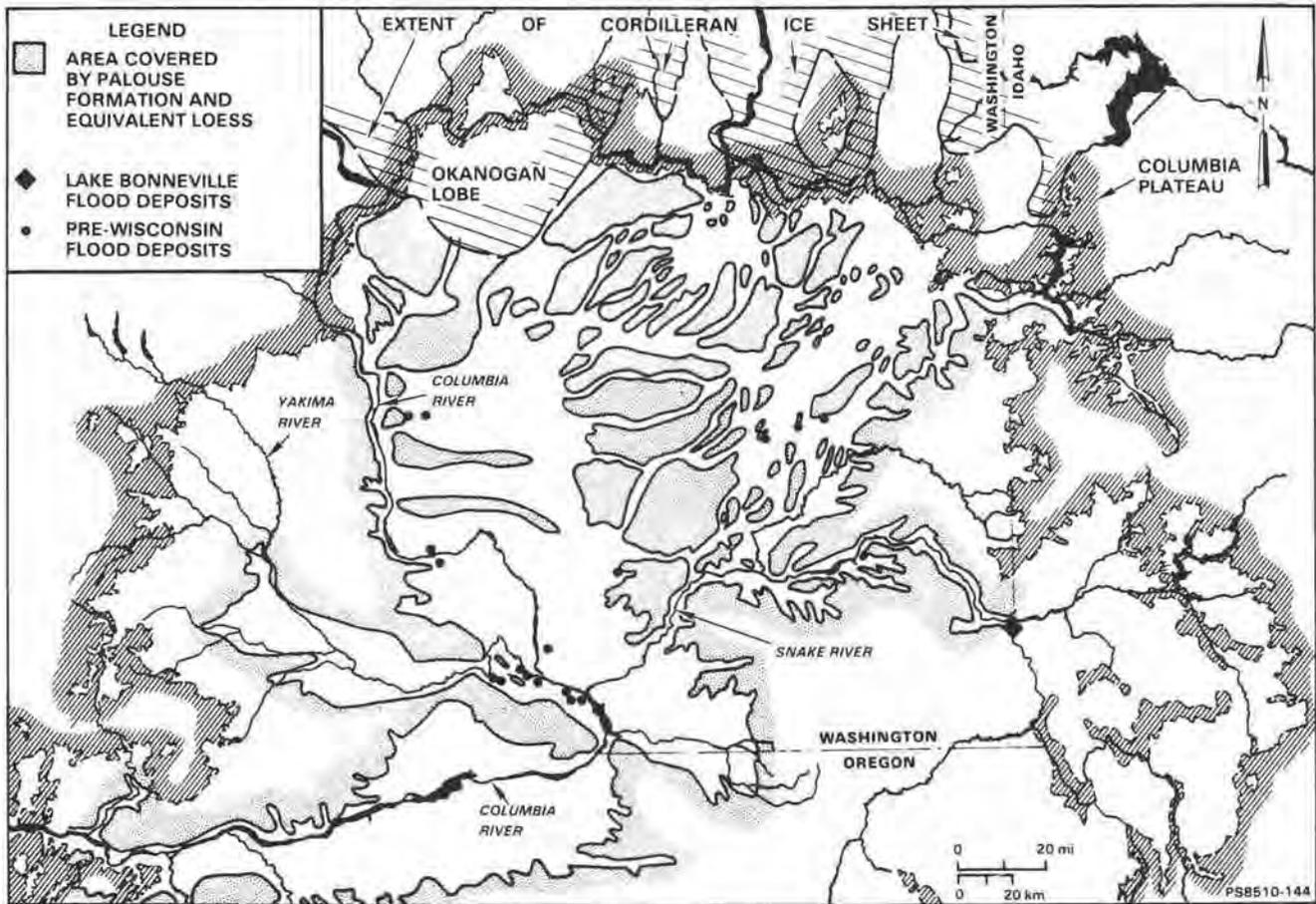


Figure 14.—Stream drainage on the northern part of the Columbia Plateau during the Pleistocene (after Bretz, 1928; Baker and Nummedal, 1978; Richmond and others, 1965; observations by the authors).

course through the Cascade Range. The new course was established in about its present-day position, near where the basalts lap out against old rocks of the Skamania Volcanic Series of Trimble (1963). The Columbia River eroded a new channel through the Columbia River Basalt Group and rapidly incised into the weathered and altered Skamania volcanics rocks (Tolan and Beeson, 1984). Erosion of the new channel is estimated to have started about 2 m.y. ago, at the same time as the onset of Cascadian uplift (Tolan and Beeson, 1984). The entrenchment of the Columbia River in the Skamania rocks may have caused a lowering of the base level and initiated headward erosion into the Columbia Plateau. Although not proven, the combined volcanism and channel erosion in the Columbia River Gorge are factors that may help to explain the regional depositional and erosional events on the Columbia Plateau of Washington State in middle to late Pliocene time.

By the late Pliocene, the Columbia and Yakima Rivers were established very near their present positions. Also, in the late Pliocene or early Pleistocene,

the Snake River was captured by a tributary of the Salmon River and became part of the Columbia River system (Wheeler and Cook, 1954; Webster and others, 1984). This capture completed the latest shift in the paleodrainage of major streams in the Columbia system.

Pleistocene

During the Pleistocene epoch, continental glaciers repeatedly advanced and retreated from British Columbia and northern Washington and occasionally impinged on the northern Columbia Plateau (Fig. 14). The glaciers temporarily plugged drainages, diverting the Columbia River to more southern courses and impounding meltwaters in glacial lakes. This plugging and diversion had minimal and temporary effects on the drainage system in the Kittitas Valley, Waterville Plateau, and Columbia valley. However, failure of ice dams impounding glacial lakes, particularly glacial Lake Missoula, released tremendous volumes of meltwater that catastrophically flooded major parts of the Columbia Plateau of Washington State. The major

floods obliterated much of the secondary drainage system of the Palouse Slope and caused minor changes in the courses of the Columbia and Yakima Rivers in the Pasco Basin.

In the Kittitas Basin, till and outwash accumulated during three Cascadian alpine glacial events: the Lookout Mountain Ranch Drift (older than the Kittitas Drift), Kittitas Drift (about 130,000-140,000 yr old) and the Lakedale Drift (about 10,000-20,000 yr old) (Porter, 1976; Waitt, 1979). Glaciation in the upper Kittitas Valley had little effect on the course of the Yakima River on the Columbia Plateau during the Pleistocene.

In the north-central Columbia Plateau, the Okanogan Lobe of the Cordilleran Ice Sheet (Fig. 14) encroached onto the Waterville Plateau (Salisbury, 1901; Waters, 1933; Flint, 1935). This lobe periodically impounded the Columbia River in the Columbia Valley to form temporary lakes (glacial Lake Columbia) and caused the diversion of the river to a more southerly course through a series of coulees south and east of the Waterville Plateau (Bretz, 1923; Hanson, 1970). The diversion of the river was temporary, and upon northward retreat of the Okanogan Lobe, the Columbia River re-established its course along the northern margin of the plateau.

The most significant change to the drainage system on the plateau during the Pleistocene resulted from catastrophic floods from glacial lakes. Failure of ice dams released large volumes of meltwater, which flowed across the Palouse Slope and into the central Columbia Plateau. Fluvial erosion associated with the flooding created a spectacular landscape of anastomosing channels, wide cataracts, and deep plunge pools of the Channeled Scabland (Bretz, 1923). The flooding also created large gravel bars along the channels and deposited thick sequences of glaciofluvial deposits in the basins and major valleys of the Columbia Plateau (Fig. 14), with the exception of the Kittitas Valley.

At least four major catastrophic floods have been recorded on the Columbia Plateau—two pre-Wisconsin and two Wisconsin events. Three of these events may have included one or more floods. The flooding events have been analyzed using paleomagnetic measurements from fine-grained flood deposits or capping loess horizons, U-Th data from overlying calcic horizons, and tephra data.

Flood deposits at two localities have reversed magnetic polarity. At the Marengo site (Patton and Baker, 1978), in the Cheney-Palouse scabland tract, flood gravels are overlain by a loess that has reversed magnetic polarity (Van Alstine, 1982). These gravels are believed to predate the last reversal in magnetic polarity (Matuyama Epoch), which ended about 730,000

years ago (Mankinen and Dalrymple, 1979). Along the Yakima Bluffs in the lower reaches of the Yakima River about 9.5 km upstream of the confluence with the Columbia River (Fig. 1) is a rhythmic series of turbidite-like deposits resembling late Wisconsin Touchet Beds (Flint, 1938a) that are underlain by a Pleistocene fluvial sequence. Both the turbidite-like beds and the upper fine unit of the older alluvial sequence have reversed magnetic polarity, indicating an age greater than 730,000 yr. The Marengo and Yakima Bluffs sites are the only early flood sites that have an established magnetostratigraphy. Other old flood deposits at Revere, McCall, and Old Maid Coulee in the Cheney-Palouse scabland tract are tentatively included in the early flooding event, based on the correlations of Patton and Baker (1978). The old flood deposits at George and Winchester Wasteway in the western Quincy Basin are also included in this early flooding event, based primarily on the degree of weathering of the clasts and the petrocalcic paleosol that caps the flood gravels at both localities. The distribution of the old flood gravels suggests that the early floods initiated the formation of the Channeled Scabland but apparently had little effect on the course of major streams during glacial periods.

Evidence of the second major flooding event includes a series of gravel deposits that are commonly capped by a calcic paleosol. Two paleosol samples have been radiometrically (U-Th) dated at 200,000 +250,-700 and 220,000 +380,-70 yr old (Tallman and others, 1978). Silty sand lenses interbedded in the gravel characteristically have a normal magnetic polarity; their age is thus less than 730,000 yr. We interpret the second flooding event to be between about 200,000 and 730,000 yr old. The best exposures of deposits from the second flooding event are in the southern Pasco Basin, where the gravels form a terrace that unconformably overlies the Ringold Formation. Gravels of this flooding event are also exposed in the northern and eastern parts of the Pasco Basin. (See Fig. 14.)

The first or second major Pleistocene flooding events had the most pronounced effects on major stream courses in the Pasco Basin. The Yakima River apparently abandoned its course through Badger Coulee in the southern Pasco Basin and shifted to its present course after the coulee was plugged with a 30-m-thick sequence of flood gravels. Bunker (1980) maintained those gravels represented the combined course of the Columbia and Yakima Rivers through the coulee. However, we believe the gravels are flood deposits because of their large-scale foreset beds and abundant subangular basalt clasts mixed with a variety of exotic lithologies; these features have been recently

exposed in gravel pits. Also during one of these early flooding events, the Columbia River channel between Gable Mountain and Gable Butte, east of Umtanum Ridge, appears to have been diverted around the east end of Gable Mountain to near its present-day course. The pre-flood river channel is plugged by a large flood bar south of Gable Mountain and Gable Butte.

A breach along the northern shoreline of Lake Bonneville at Red Rock Pass, near Preston, Idaho, resulted in flooding down the Snake River and onto the Columbia Plateau (Malde, 1968; Swanson, 1984). This was the third major flooding event on the plateau. The event, which occurred about 14,000 to 15,000 yr ago (Scott, 1983), must have resulted in the flooding of river courses to the Pacific Ocean. However, sedimentary deposits formed during the glacial Lake Bonneville flood are only known along the southeastern margin of the Columbia Plateau. In the Lewiston Basin, flood debris formed several bars marginal to the present-day Snake River. Glacial Lake Bonneville apparently had little, if any, impact on stream courses in the Pasco Basin or elsewhere on the Columbia Plateau.

The last major flooding event is the glacial Lake Missoula flood of Bretz (1923) that inundated much of the Columbia Plateau in Washington State. (See Fig. 14.) The flooding resulted in the last scouring of the Channeled Scabland and widespread deposition of glaciofluvial sediments on the plateau. (See Bretz, 1923; Bretz and others, 1956; Baker, 1973; Baker and Nummedal, 1978; Waitt, 1980.) The last major flooding occurred at about 12,000 yr ago, based on the presence of Mount St. Helens Set "S" ash in the upper part of these flood sediments (Mullineaux and others, 1978).

The catastrophic floods (except from Lake Bonneville) referred to here may represent one great flood in each event, as Bretz (1923) proposed for the late Wisconsin flooding, or multiple floods, as demonstrated by Waitt (1980) for late Wisconsin flooding. However, flood events and associated glaciation resulted in only temporary or minor changes to major streams on the plateau. Catastrophic flooding (except from Lake Bonneville) created the Channeled Scabland and destroyed the pre-Pleistocene drainage network on the Palouse Slope. The Pleistocene events resulted in the last major sculpturing of the plateau landscape and essentially the final positioning of major stream courses; only minor shifts occurred during the Holocene.

SUMMARY

The evolution of the Columbia River drainage system on the Columbia Plateau of Washington in the

last 17 m.y. reflects the geologic history of the plateau. We have updated an interpretation of the evolution of the river system and defined the geomorphic and structural features that have controlled the position of ancestral streams. The sequence of geologic events and the resulting drainage system for various time intervals in the last 17 m.y. are summarized below.

Early to Middle Miocene

Grande Ronde Basalt was repeatedly erupted from fissures in the eastern Columbia Plateau. The flood basalts flowed westward down the Palouse Slope and inundated the intermontane basin between the Rocky Mountains and the ancestral Cascade Range. Evidence of major streams is found mainly near the margins of the plateau, where those streams flowed between the constructional topography of basalt flows and the surrounding highlands of pre-Columbia River basalt age. Major streams were flowing westward around the margins of the plateau before coalescing to flow through the ancestral Cascade Range south of the present-day Columbia River Gorge. Because of repeated volcanism, integrated drainage did not have a chance to develop in the central plateau. Thus, drainage there consisted of poorly integrated small streams and shallow lakes, marshes, and bogs.

Middle Miocene

The beginning of the middle Miocene was marked by a short period of volcanic quiescence over much of the Columbia Plateau, represented by the Vantage Member interbed of the Ellensburg Formation. During this interval, continued uplift of the Naneum Ridge diverted the ancestral Columbia River to the east near the north-central part of the vast basaltic plain. The diversion of the river led to the development of an ancestral Yakima River in the northwest plateau. Renewed Columbia River basalt volcanism (the Wanapum Basalt) again inundated the Columbia Plateau. The drainage pattern during this volcanic episode was similar to that of early Miocene time; streams were confined to the perimeter of the plateau, and local streams, lakes, marshes, and bogs occupied the central plateau area.

Middle to Late Miocene

Columbia River basalt volcanism continued to wane during the middle to late Miocene; periodic eruptions produced the Saddle Mountains Basalt. Subsidence of the central plateau and continued deformation of parts of the plateau formed topographic features that were not being continuously covered by basalt flows. Saddle Mountains Basalt was erupted

from fissures in the eastern plateau, primarily in the Clearwater Embayment, and flowed down the ancestral Clearwater-Salmon River canyons—the intra-canyon flows define the drainage channels of these streams. These flows, in conjunction with gravel trains in the Ellensburg Formation indicate the position of the paleodrainage system in the southern part of the Columbia Plateau of Washington. The ancestral Columbia, Clearwater-Salmon, and Yakima Rivers were controlled by structural relief developing on the plateau and by constructional topography formed with emplacement of flows in the central plateau. The ancestral Columbia River flowing south down the Columbia Slope and the ancestral Clearwater-Salmon flowing west down the Palouse Slope joined in the northern Pasco Basin to flow southwest to the present-day Columbia River Gorge area. The ancestral Yakima River joined the combined Columbia and Clearwater-Salmon Rivers in the developing Yakima Valley. Later in the Saddle Mountains volcanic episode, the ancestral Clearwater-Salmon River was diverted into the southern Pasco Basin and through the Horse Heaven Hills at Wallula Gap.

Late Miocene to Middle Pliocene

Columbia River basalt volcanism had essentially ceased by late Miocene time, but continued tectonic deformation of parts of the basaltic plain resulted in the development of prominent structural ridges and lows. The major streams flowed in the structural basins and synclinal valleys and in water gaps through structural ridges. Near the beginning of this time interval, the ancestral Columbia and Yakima Rivers were captured by the ancestral Clearwater-Salmon River to establish the entire northern part of the Columbia River system through the Horse Heaven Hills at Wallula Gap. The capture may have been the result of plugging of the combined ancestral Columbia and Yakima river channel by the eruption of the Simcoe Volcanics in the Goldendale area, and (or) continued subsidence of the Pasco Basin and accumulation of a sufficient thickness of sediments to cover the constructional topography of the Ice Harbor Member. Following the river capture, the ancestral rivers aggraded sediments in the basins of the Columbia Plateau until the middle Pliocene.

Middle to Late Pliocene

In about middle Pliocene time, aggradation of sediments in the structural lows of the central Columbia Plateau ended rather abruptly, and the rivers began downcutting. The slow rate of deformation of the Horse Heaven Hills apparently allowed the

Columbia River system to maintain its channel through Wallula Gap during uplift of the ridge. However, that deformation explains only a part of a major regional base level change observed throughout the plateau. Tolan and Beeson (1984) proposed that the ancestral channel of the Columbia River in the Cascade Range was inundated by basalts and volcanic debris from the Cascade Range. We believe this was a factor in a higher base level and regional aggradation of sediments in the Columbia Plateau. The Columbia River then established a new channel in its present-day position about 2 m.y. ago. Downcutting in the Columbia River Gorge resulted in a lower base level and consequent incision into sedimentary deposits in the plateau. At about 2 m.y. ago, the Snake River that flowed south of the Columbia Plateau was captured by the Columbia River system, thus completing the modern Columbia River system.

Pleistocene

Glaciation on the northern edge of the Columbia Plateau and catastrophic flooding that resulted from the dumping of glacially dammed lakes caused only temporary or minor changes to the major rivers on the plateau. At least four flooding events resulted in the last major sculpturing of the plateau and essentially the final positioning of major stream courses. Only minor local readjustments in the major streams in topographic lows have occurred during the Holocene.

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REFERENCES CITED

- Alto, B. R., 1955, Geology of a part of the Boylston quadrangle and adjacent areas in central Washington: University of Washington Master of Science thesis, 38 p.
- Anderson, J. L.; Beeson, M. H.; Bentley, R. D.; Fecht, K. R.; Hooper, P. R.; Niem, A. R.; Reidel, S. P.; Swanson, D. R.; Tolan, T. L.; Wright, T. L., 1987, Distribution maps of stratigraphic units of the Columbia River Basalt Group. In Schuster, J. E., editor, 1987, Selected papers on the geology of Washington: Washington Division of Geology and Earth Resources Bulletin 77, (this volume).

- Anderson, J. L.; Vogt, B. F., 1987, Intracanyon flows of the Columbia River Basalt Group in the southwest part of the Columbia Plateau and adjacent Cascade Range, Oregon and Washington. In Schuster, J. E., editor, 1987, Selected papers on the geology of Washington: Washington Division of Geology and Earth Resources Bulletin 77, (this volume).
- ARHCO, 1976, Preliminary feasibility study on storage of radioactive wastes in Columbia River basalts: Atlantic Richfield Hanford Company, report ARH-ST-137, 2 v.
- Baker, V. R., 1973, Paleohydrology and sedimentology of Lake Missoula flooding in eastern Washington: Geological Society of America Special Paper 144, 79 p.
- Baker, V. R.; Nummedal, Dag, editors, 1978, The Channeled Scabland—A guide to the geomorphology of the Columbia Basin, Washington: U.S. National Aeronautics and Space Administration, 186 p.
- Beeson, M. H.; Moran, M. R., 1979, Columbia River Basalt Group stratigraphy in western Oregon: Oregon Geology, v. 41, no. 1, p. 11-14.
- Bentley, R. D., 1977, Stratigraphy of the Yakima basalts and structural evolution of the Yakima Ridges in the western Columbia Plateau. In Brown, E. H.; Ellis, R. C., editors, 1977, Geological excursions in the Pacific Northwest, Geological Society of America Annual Meeting: Western Washington University, p. 339-389.
- Bentley, R. D.; Anderson, J. L.; Campbell, N. P.; Swanson, D. A., 1980, Stratigraphy and structure of the Yakima Reservation with emphasis on the Columbia River Basalt Group: U.S. Geological Survey Open-File Report 80-200, 86 p., 1 pl.
- Bond, J. G., 1963, Geology of the Clearwater Embayment: Idaho Bureau of Mines and Geology Pamphlet 128, 83 p.
- Bond, J. G.; Kauffman, J. D.; Miller, D. A.; Barrash, W., 1978, Geology of the southwestern Pasco Basin: Rockwell Hanford Operations RHO-BWI-C-25, 217 p., 43 pl.
- Bretz, J. H., 1923, The channeled scablands of the Columbia Plateau: Journal of Geology, v. 31, no. 8, p. 617-649.
- Bretz, J. H., 1928, The channeled scablands of eastern Washington: The Geographical Review, v. 18, no. 3, p. 446-477.
- Bretz, J. H., 1959, Washington's channeled scabland: Washington Division of Mines and Geology Bulletin 45, 57 p., 4 pl.
- Bretz, J. H.; Smith, H. T. U.; Neff, G. E., 1956, Channeled scablands of Washington—New data and interpretations: Geological Society of America Bulletin, v. 67, no. 8, p. 957-1049.
- Brown, R. E., 1966, The relationships between the Ellensburg and Ringold Formations in the Pasco Basin: Battelle Pacific Northwest Laboratory BNWL-SA-621, 12 p.
- Brown, R. E.; Brown, D. J., 1961, The Ringold Formation and its relationships to other formations: General Electric Company HW-SA-2319, 17 p.
- Bunker, R. C., 1980, Catastrophic flooding in the Badger Coulee area, south-central Washington: University of Texas at Austin Master of Arts thesis, 183 p.
- Caggiano, J. A.; Duncan, D. W., editors, 1983, Preliminary interpretation of the tectonic stability of the reference repository location, Cold Creek syncline, Hanford Site: Rockwell Hanford Operations RHO-BWI-ST-19P, 1 v., 5 pl.
- Calkins, F. C., 1905, Geology and water resources of a portion of east-central Washington: U.S. Geological Survey Water-Supply Paper 118, 96 p.
- Camp, V. E., 1976, Petrochemical stratigraphy and structure of the Columbia River basalt, Lewiston area, Idaho-Washington: Washington State University Doctor of Philosophy thesis, 201 p., 1 pl.
- Camp, V. E., 1981, Geologic studies of the Columbia Plateau—Part II. Upper Miocene basalts distribution, reflecting source locations, tectonism and drainage history in the Clearwater Embayment, Idaho: Geological Society of America Bulletin, Part I, v. 92, no. 9, p. 669-678.
- Camp, V. E.; Hooper, P. R., 1981, Geologic studies of the Columbia Plateau—Part II. Late Cenozoic evolution of the southeast part of the Columbia River basalt province: Geological Society of America Bulletin, Part I, v. 92, no. 9, p. 659-668.
- Campbell, N. P., 1979, Geologic map of the Toppenish Ridge, and Geologic map of the north flank of Saddle Mountains and Smyrna Bench. In Rigby, J. G.; Othberg, K. L.; Campbell, N. P.; Hanson, L.; Kiver, E.; Stradling, D.; Webster, G., 1979, Reconnaissance surficial geologic mapping of the late Cenozoic sediments of the Columbia Basin, Washington: Washington Division of Geology and Earth Resources Open File Report 79-3, 97 p., 7 pl.
- Campbell, N. P., 1983, Correlation of late Cenozoic gravel deposits along the Yakima River drainage from Ellensburg to Richland, Washington: Northwest Science, v. 57, no. 3, p. 179-193.
- Chappell, W. M., 1936, The effect of Miocene lavas on the course of the Columbia River in central Washington: Journal of Geology, v. 44, p. 379-386.
- Connors, J. A., 1976, Quaternary history of northern Idaho and adjacent areas: University of Idaho Doctor of Philosophy thesis, 117 p., 1 pl.
- Diery, H. D.; McKee, Bates, 1969, Stratigraphy of the Yakima Basalt in the type area: Northwest Science, v. 43, no. 2, p. 47-64.
- Fecht, K. R.; Lillie, J. T., 1982, A catalog of borehole lithologic logs from the 600 area, Hanford site: Rockwell Hanford Operations RHO-LD-158, 249 p.
- Fecht, K. R.; Reidel, S. P.; Tallman, A. M., 1982, Evolution of the Columbia River system in the central Columbia Plateau of Washington from Miocene to the present [abstract]: Geological Society of America Abstracts with Programs, v. 14, no. 4, p. 163.
- Flint, R. F., 1935, Glacial features of the southern Okanogan region: Geological Society of America Bulletin, v. 46, no. 2, p. 169-193.
- Flint, R. F., 1938a, Summary of late-Cenozoic geology of southeastern Washington: American Journal of Science, 5th Series, v. 35, no. 207, p. 223-230.
- Fryxell, Roald; Cook, E. F., 1964, A field guide to the loess deposits and channeled scablands of the Palouse area, eastern Washington: Washington State University Laboratory of Anthropology Report of Investigations 27, 32 p.
- Fuller, R. E., 1931, The aqueous chilling of basaltic lava on the Columbia Plateau: American Journal of Science, 5th Series, v. 21, no. 5, p. 281-300.
- Gardner, J. N.; Snow, M. G.; Fecht, K. R., 1981, Geology of the Wallula Gap area, Washington: Rockwell Hanford Operations RHO-BWI-LD-9, 88 p.
- Griggs, A. B., 1976, The Columbia River Basalt Group in the Spokane quadrangle, Washington, Idaho, and Montana: U.S. Geological Survey Bulletin 1413, 39 p.
- Grolier, M. J., 1965, Geology of part of the Big Bend area in the Columbia Plateau, Washington: The Johns Hopkins University Doctor of Philosophy thesis, 265 p., 1 pl.
- Grolier, M. J.; Bingham, J. W., 1971, Geologic map and sections of parts of Grant, Adams, and Franklin Counties, Washington: U.S. Geological Survey Miscellaneous Geologic Investigations Map I-589, 6 sheets, scale 1:62,500.

- Grolier, M. J.; Bingham, J. W., 1978, Geology of part of Grant, Adams, and Franklin Counties, east-central Washington: Washington Division of Geology and Earth Resources Bulletin 71, 91 p.
- Gustafson, E. P., 1978, The vertebrate faunas of the Pliocene Ringold Formation, south-central Washington: University of Oregon Museum of Natural History Bulletin 23, 62 p.
- Hanson, L. G., 1970, The origin and development of Moses Coulee and other scabland features of the Waterville Plateau, Washington: University of Washington Doctor of Philosophy thesis, 139 p.
- Kirkham, V. R. D.; Johnson, M. M., 1929, The Latah formation in Idaho: *Journal of Geology*, v. 37, no. 5, p. 483-504.
- Kuhns, M. J. P., 1980, Late Cenozoic deposits of the lower Clearwater valley, Idaho and Washington: Washington State University Master of Science thesis, 71 p., 1 pl.
- LaBreque, J. L.; Kent, D. V.; Cande, S. C., 1977, Revised magnetic polarity time scale for Late Cretaceous and Cenozoic time: *Geology*, v. 5, p. 330-335.
- Landon, R. D.; Long, P. E.; Reidel, S. P., 1982, Stratigraphy and flow distribution of Grande Ronde Basalt—Evidence for early tectonic deformation: Rockwell Hanford Operations RHO-BWI-SA-184, 11 p.
- Laval, W. N., 1956, Stratigraphy and structural geology of portions of south-central Washington: University of Washington Doctor of Philosophy thesis, 223 p. 11 pl.
- Lefebvre, R. H., 1966, Variations of flood basalts on the Columbia River Plateau, central Washington: Northwestern University Doctor of Philosophy thesis, 211 p.
- Long, P. E.; Duncan, R. A., 1982, $^{40}\text{Ar}/^{39}\text{Ar}$ ages of Columbia River basalts from deep boreholes in south-central Washington [abstract]: *Eos [American Geophysical Union Transactions]*, v. 64, no. 9, p. 90.
- Mackin, J. H., 1961, A stratigraphic section in the Yakima Basalt and Ellensburg Formation in south-central Washington: Washington Division of Mines and Geology Report of Investigations 19, 45 p. 9 pl.
- Malde, H. E., 1968, The catastrophic late Pleistocene Bonneville Flood in the Snake River Plain, Idaho: U.S. Geological Survey Professional Paper 596, 52 p.
- Mankinen, E. A.; Dalrymple, G. B., 1979, Revised geomagnetic polarity time scale for the interval 0-5 m.y.B.P.: *Journal of Geophysical Research*, v. 84, no. 2, p. 615-626.
- Mason, G. W., 1953, Interbasalt sediments of south-central Washington: Washington State College Master of Science thesis, 116 p., 3 pl.
- McKee, E. H.; Swanson, D. A.; Wright, T. L., 1977, Duration and volume of Columbia River basalt volcanism, Washington, Oregon, and Idaho [abstract]: *Geological Society of America Abstracts with Programs*, v. 9, no. 4, p. 463-464.
- Merriam, J. C.; Buwalda, J. P., 1917, Age of strata referred to the Ellensburg Formation in the White Bluffs of the Columbia River: University of California Publications in Geological Sciences, v. 10, no. 15, p. 255-266.
- Mullineaux, D. R.; Wilcox, W. R.; Ebaugh, W. R.; Fryxell, Roald; Rubin, M., 1978, Age of the last major scabland flood of Columbia Plateau in eastern Washington: *Quaternary Research*, v. 10, no. 2, p. 171-180.
- Myers, C. W.; Price, S. M., and others, 1979, Geologic studies of the Columbia Plateau—A status report: Rockwell Hanford Operations RHO-BWI-ST-4, 541 p., 53 pl.
- Newcomb, R. C., 1958, Ringold Formation of Pleistocene age in type locality, the White Bluffs, Washington: *American Journal of Science*, v. 256, no. 5, p. 328-340.
- Newcomb, R. C., 1965, Geology and ground-water resources of the Walla Walla River basin, Washington-Oregon: Washington Division of Water Resources Water-Supply Bulletin 21, 151 p., 4 pl.
- Newcomb, R. C.; Strand, J. R.; Frank, F. J., 1972(1973), Geology and ground-water characteristics of the Hanford Reservation of the U.S. Atomic Energy Commission, Washington: U.S. Geological Survey Professional Paper 717, 78 p.
- Pardee, J. T.; Bryan, Kirk, 1926, Geology of the Latah Formation in relation to the lavas of the Columbia Plateau near Spokane, Washington: U.S. Geological Survey Professional Paper 140-A, p. 1-16.
- Patton, P. C.; Baker, V. R., 1978, New evidence for pre-Wisconsin flooding in the channeled scabland of eastern Washington: *Geology*, v. 6, no. 9, p. 567-571.
- Porter, S. C., 1976, Pleistocene glaciation in the southern part of the north Cascade Range, Washington: *Geological Society of America Bulletin*, v. 87, no. 1, p. 61-75.
- Puget Sound Power and Light, 1982, Preliminary safety analysis report for the Skagit/Hanford nuclear project: Bellevue Washington.
- Raymond, J. R.; Tillson, D. D., 1968, Evaluation of a thick basalt sequence in south-central Washington—Geophysical and hydrological exploration of the Rattlesnake Hills deep stratigraphic test well: Battelle Pacific Northwest Laboratory BNWL-776, 120 p.
- Reidel, S. P., 1978, Stratigraphy and petrogenesis of the Grande Ronde basalt in the lower Salmon and adjacent Snake River Canyons: Rockwell Hanford Operations RHO-BW-SA-62, 415 p.
- Reidel, S. P.; Fecht, K. R., 1981, Wanapum and Saddle Mountains Basalts in the Cold Creek Syncline Area. In Myers, C. W.; Price, S. M., editors, 1981, Subsurface geology of the Cold Creek syncline: Rockwell Hanford Operations RHO-BWI-ST-14, p. 3-1 - 3-45.
- Reidel, S. P.; Ledgerwood, R. K.; Myers, C. W.; Jones, M. G.; Landon, R. D., 1980, Rate of deformation in the Pasco Basin during the Miocene as determined by distribution of Columbia River basalt flows [abstract]: *Geological Society of America Abstracts with Programs*, v. 12, no. 3, p. 149.
- Repenning, C. A., in press, Biochronology of microtine rodents in the United States. In Woodburne, M. O., editor, *Cenozoic mammals—Their record, biostratigraphy, and biochronology*: University of California at Berkeley Press.
- Richmond, G. M.; Fryxell, Roald; Neff, G. E.; Weis, P. L., 1965, The Cordilleran ice sheet of the northern Rocky Mountains, and related Quaternary history of the Columbia Plateau. In Wright, H. E.; Frey, D. G., editors, 1965, *The Quaternary of the United States*: Princeton University Press, p. 231-242.
- Rigby, J. G.; Othberg, K.; Campbell, N. P.; Hanson, L.; Kiver, E.; Stradling, D.; Webster, G., 1979, Reconnaissance surficial geologic mapping of the late Cenozoic sediments of the Columbia Basin, Washington: Washington Division of Geology and Earth Resources Open File Report 79-3, 94 p., 7 pl.
- Routson, R. C.; Fecht, K. R., 1979, Soil (sediment) properties of twelve Hanford wells: Rockwell Hanford Operations RHO-LD-82, 52 p.
- Russell, I. C., 1893, A geologic reconnaissance in central Washington: U.S. Geological Survey Bulletin 108, 108 p.
- Russell, I. C., 1901, Geology and water resources of Nez Perce County, Idaho [Parts I and II]: U.S. Geological Survey Water-Supply Papers 53 and 54, 141 p.
- Salisbury, R. D., 1901, Glacial work in the western mountains: *Journal of Geology*, v. 9, p. 721-730.

- Schmincke, H. -U., 1964, Petrology, paleocurrents, and stratigraphy of the Ellensburg Formation and interbedded Yakima Basalt flows, south-central Washington: The Johns Hopkins University Doctor of Philosophy thesis, 426 p.
- Schmincke, H. -U., 1967a, Flow directions in Columbia River basalt flows and paleocurrents of interbedded sedimentary rocks, south-central Washington: *Geologisches Rundschau*, v. 56, no. 3, p. 992-1019.
- Schmincke, H. -U., 1967b, Stratigraphy and petrology of four upper Yakima basalt flows in south-central Washington: *Geological Society of America Bulletin*, v. 78, no. 11, p. 1385-1422.
- Scott, W. E., 1983, The last two cycles of Lake Bonneville, [abstract]: *Geological Society of America Abstracts with Programs*, v. 15, no. 5, p.300.
- Shannon and Wilson, 1972, Geologic and subsurface investigation—Proposed mid-Columbia thermal Power Plant Sites, Morrow, Gilliam, and Umatilla Counties, Oregon: for Portland General Electric Company, Portland, Oregon.
- Shannon and Wilson, 1975, Geotechnical investigations for central plant facilities, Pebble Springs site, Boardman Nuclear Project, Gilliam County, Oregon: for Portland General Electric Company, Portland, Oregon.
- Smith, G. O., 1901, Geology and water resources of a portion of Yakima County, Washington: U.S. Geological Survey Water-Supply Paper 55, 68 p.
- Smith, G. O., 1903, Ellensburg, Washington: U.S. Geological Survey Geologic Atlas of the United States, Folio 86, 7 p., 3 maps.
- Swanson, D. A., 1967, Yakima basalt of the Tieton River area, south-central Washington: *Geological Society of America Bulletin*, v. 78, no. 9, p. 1077-1109.
- Swanson, D. A.; Bentley, R. D.; Byerly, G. R.; Gardner, J. N.; Wright, T. L.: 1979b, Preliminary reconnaissance geologic maps of the Columbia River Basalt Group in parts of eastern Washington and northern Idaho: U.S. Geological Survey Open-File Report 79-534, 25 p., 9 pl.
- Swanson, D. A.; Wright, T. L.: 1976, Guide to field trip between Pasco and Pullman, Washington, emphasizing stratigraphy, vent areas, and intracanyon-flows of Yakima Basalt; *Geological Society of America, Cordilleran Section, 72nd Annual Section Meeting, Pullman, Field guide no. 1: Washington State University [Department of Geology]*, 33 p.
- Swanson, D. A.; Wright, T. L.: 1979, Paleogeography of southeast Washington during the middle and late Miocene based on the distribution of intracanyon basalt flows [abstract]. In *Armstrong, J. M.; Cole, M. R.; Ter Best, Harry, Jr., editors, 1979, Cenozoic paleogeography of the western United States, Pacific Coast Paleogeography Symposium 3: Society of Economic Paleontologists and Mineralogists Pacific Section*, p. 331.
- Swanson, D. A.; Wright, T. L.; Camp, V. E.; Gardner, J. N.; Helz, R. T.; Price, S. A.; Reidel, S. P.; Ross, M. E., 1980, Reconnaissance geologic map of the Columbia River Basalt Group, Pullman and Walla Walla quadrangles, southeast Washington and adjacent Idaho: U.S. Geological Survey Miscellaneous Geologic Investigations Map I-1139, 2 sheets, scale 1:250,000.
- Swanson, D. A.; Wright, T. L.; Hooper, P. R.; Bentley, R. D., 1979a, Revisions in stratigraphic nomenclature of the Columbia River Basalt Group: U.S. Geological Survey Bulletin 1457-G, 59 p., 1 pl.
- Swanson, E. C., 1984, Bonneville flood deposits along the Snake River near Lewiston, Idaho: Washington State University Master of Science thesis, 74 p.
- Tabor, R. W.; Frizzell, V. A., Jr.; Whetten, J. T.; Swanson, D. A.; Byerly, G. R.; Booth, D. B.; Hetherington, M. J.; Waitt, R. B., 1980, Preliminary geologic map of the Chelan 1:100,000 quadrangle, Washington: U.S. Geological Survey Open-File Report 80-841, 46 p. 1 pl., scale 1:100,000.
- Tabor, R. W.; Waitt, R. B.; Frizzell, V. A., Jr.; Swanson, D. A.; Byerly, R.; Bentley, R. D., 1982, Geologic map of the Wenatchee 1:100,000 quadrangle, central Washington: U.S. Geological Survey Miscellaneous Investigations Map I-1311, 26 p., 1 pl., scale 1:100,000.
- Tallman, A. M.; Fecht, K. R.; Lillie, J. T.; Campbell, N. P.; Hanson, L. G.; Kiver, E. P.; Stradling, D. F.; Webster, G. D., 1982, Post-Columbia River basalt stratigraphy of the Columbia Plateau [abstract]: *Geological Society of America Abstracts with Programs*, v. 14, no. 4, p. 239.
- Tallman, A. M.; Fecht, K. R.; Marratt, M. C.; Last, G. V.; and others, 1979, Geology of the separations areas, Hanford Site, south-central Washington: Rockwell Hanford Operations RHO-ST-23, 134 p.
- Tallman, A. M.; Lillie, J. T.; Caggiano, J. A., 1978, The late Cenozoic geology of the western Pasco Basin. In *Basalt Waste Isolation Program Annual Report—Fiscal Year 1978: Rockwell Hanford Operations RHO-BWI-78-100*, p. 78-81.
- Tallman, A. M.; Lillie, J. T.; Fecht, K. R., 1981, Suprabasalt sediments of the Cold Creek syncline area. In *Myers, C. W.; Price, S. M., editors, 1981, Subsurface geology of the Cold Creek syncline: Rockwell Hanford Operations RHO-BWI-ST-14*, p. 2-1 - 2-28.
- Tolan, T. L., 1982, The stratigraphic relationships of the Columbia River Basalt Group in the Lower Columbia River Gorge of Oregon and Washington: Portland State University Master of Science thesis, 151 p., 1 pl.
- Tolan, T. L.; Beeson, M. H., 1984, Intracanyon flows of the Columbia River Basalt Group in the Lower Columbia River Gorge and their relationship to the Troutdale Formation: *Geological Society of America Bulletin*, v. 95, no. 4, p. 463-477.
- Trimble, D. E., 1963, Geology of Portland, Oregon, and adjacent areas: U.S. Geological Survey Bulletin 1119, 119 p., 1 pl.
- Van Alstine, D. R., 1982, Paleomagnetic investigation of pre-Missoula gravels, Pasco Basin and vicinity, Washington: Sierra Geophysics, Inc., SGI-R-82-055, for Golder Associates, 1 v.
- Waggoner, G. L., 1981, Sedimentary analysis of gravel deposits in the vicinity of Clarkston, Washington: Washington State University Master of Science thesis, 107 p., 1 pl.
- Waitt, R. B., 1979, Late Cenozoic deposits, landforms, stratigraphy and tectonism in Kittitas Valley, Washington: U.S. Geological Survey Professional Paper 1127, 18 p.
- Waitt, R. B., 1980, About forty last-glacial Lake Missoula jokulhlaups through southern Washington: *Journal of Geology*, v. 88, no. 6, p. 653-679.
- Waring, G. A., 1913, Geology and water resources of a portion of south-central Washington: U.S. Geological Survey Water-Supply Paper 316, 46 p.
- Warren, C. R., 1941, Course of the Columbia River in south-central Washington: *American Journal of Science*, v. 239, no. 3, p. 209-232.
- Waters, A. C., 1933, Terraces and coulees along the Columbia River near Lake Chelan, Washington: *Geological Society of America Bulletin*, v. 44, no. 4, p. 783-820.
- Waters, A. C., 1955, Geomorphology of south-central Washington, illustrated by the Yakima East quadrangle: *Geological Society of America Bulletin*, v. 66, no. 6, p. 663-684.
- Waters, A. C., 1960, Determining direction of flow in basalts: *American Journal of Science (Bradley Volume)*, v. 258-A, p. 350-366.
- Waters, A. C., 1961, Stratigraphic and lithologic variations in the Columbia River Basalt: *American Journal of Science*, v. 259, no. 8, p. 583-611.

- Waters, A. C., 1973, The Columbia River Gorge-Basalt stratigraphy, ancient lava dams and landslide dams. *In* Beaulieu, J. D., editor, 1973, Geologic field trips in northern Oregon and southern Washington: Oregon State Department of Geology and Mineral Industries Bulletin 77, p. 133-162.
- Waters, A. C.; Myers, C. W.; Brown, D. J.; Ledgerwood, R. K., 1981, Columbia Plateau with special emphasis on the basalt stratigraphy of the Pasco Basin. *In* Subbarao, K. V.; Sukhewala, R. N., editors, 1981, Deccan volcanism and related basalt provinces in other parts of the world: Geological Society of India Memoir 3, p. 19-44.
- Webster, G. D.; Kuhns M. J. P.; Waggoner C. L., 1984, Late Cenozoic gravels in Hells Canyon and the Lewiston Basin, Washington and Idaho. *In* Bonnicksen, B.; Breckenridge, R. M., editors, 1984, Cenozoic geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 669-683.
- Wheeler, H. E.; Cook E. F., 1954, Structural and stratigraphic evidence of the Snake River capture, Idaho-Oregon: *Journal of Geology*, v. 62, p. 525-536.
- Willis, Bailey, 1887, Changes in river courses in Washington Territory due to glaciation: U.S. Geological Survey Bulletin 40, 10 p.

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INTRACANYON FLOWS OF THE COLUMBIA RIVER BASALT GROUP IN THE SOUTHWEST PART OF THE COLUMBIA PLATEAU AND ADJACENT CASCADE RANGE, OREGON AND WASHINGTON

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ABSTRACT

The youngest members of the Miocene Columbia River Basalt Group to reach western Oregon and Washington were largely confined to paleogeographic low areas along rivers draining the Columbia Plateau from about 12 to 14 million years ago. Two of these, the Priest Rapids Member of the Wanapum Basalt and the Pomona Member of the Saddle Mountains Basalt, were intracanyon flows where they crossed the axis of the present Cascade Range. A second flow of the Saddle Mountains Basalt, the Elephant Mountain Member, was also an intracanyon flow, but it has been observed only in the eastern foothills of the present Cascades and is not known to have crossed the range. The two Saddle Mountains members are present in the same canyon and are separated by sandstone and conglomerate of the Rattlesnake Ridge Member of the Ellensburg Formation. The occurrence of both members together demonstrates the presence of a drainage well to the north of the present Columbia River that existed for at least 1.5 million years (12-10.5 million years ago). The Rattlesnake Ridge Member interbed contains a significant component of sediment that originated from sources other than the Columbia Plateau and Cascade Range, indicating that the canyon was part of a regional drainage system that was a probable ancestor to the present Columbia River. The Priest Rapids Member filled a less evolved drainage system south of the Saddle Mountains members. These Columbia River basalt flows were confined to canyons because sufficient time transpired between eruptions to allow erosional downcutting. Canyons were cut where the ancestral Columbia River extended across areas of active uplift or where headward erosion occurred. The overall distribution of the Pomona Member reveals uplifted areas where the river canyon was deep enough to completely confine the lava flow and subsided areas where it was shallow enough to allow overflow into adjacent areas. Areas of uplift and subsidence do not appear to coincide entirely with areas of later uplift and subsidence because of changes in the size and shape of evolving folds. Uplift west of Shellrock Mountain, between Mount Hood and Mount Adams, appears to have been influenced to a greater extent by general upwarping of the Cascade Range than has uplift to the east. Virtually all uplift east of Shellrock Mountain can be attributed to Columbia Plateau/Yakima Fold Belt-style deformation. Subsidence attributable to faults of the "Hood River graben" does not appear to exceed subsidence within the individual basins separating Yakima Ridge uplifts nearby.

INTRODUCTION

The Columbia River Basalt Group of the Pacific Northwest is a major subareal accumulation of tholeiitic flood basalt that erupted 16.5 to 6 m.y. ago from north- to northwest-trending fissures in southeastern Washington, northeast Oregon, and western Idaho (Swanson and others, 1979b). These vents erupted successive flows that eventually covered an area of approximately 164,000 km² (T. L. Tolan, personal commun.; Fig. 1). (See related papers by Anderson and others and by Hooper and Swanson, this volume.)

All the Columbia River basalt units thus far identified in western Oregon and Washington are formations and members of the Yakima Basalt Subgroup, the most voluminous part of the Columbia River Basalt Group. Figure 2 shows the units known to have crossed the axis of the present day Cascade Range. Of these, the Grande Ronde Basalt and the Frenchman Springs Member of the Wanapum Basalt flowed through an east-west-trending lowland approximately 75 km wide that was bounded on the north and south by mountains (Waters, 1973; Anderson, 1978; Beeson



Figure 1.—Regional geologic map of the Columbia River Basalt Group including formations of the Yakima Basalt Subgroup (stippled). I, Grande Ronde Basalt; II, Wanapum Basalt; III, Saddle Mountains Basalt. Modified from Waters (1961) and Swanson and others (1979a).

and Moran, 1979). (See also Fecht and others, this volume.)

During a period of approximately 2.5 m.y. duration (from approximately 16.5 to 14 m.y. ago), more than 30 flows, which were erupted from vents within the Columbia Plateau, spread across the lowland and covered it to a general depth of approximately 0.55 to 0.60 km (Beeson and others, 1982), with local depths of greater than 1.2 km along the northern margin. These flows were but a partial representation of an impressive number of eruptions taking place in the central and eastern plateau. It is probable that only the most voluminous eruptions reached the vicinity of the Cascade Range and that many less voluminous flows simply lapped out closer to the source vents. Some flows were probably blocked by broad low-amplitude arching and resultant tilting along the margins of the Cascade Range. Subsidence within the Columbia Plateau itself could also have been a factor in causing off-lap of units. Local thinning and exclusion of flows at anticlinal uplifts has been demonstrated within the central Columbia Plateau (Reidel, 1984) and within the Cascade Range (Vogt, 1981). Similarly, strike-slip faults and associated folds have caused thinning and exclusion of Columbia River basalt flows in the southwest Columbia Plateau (Anderson and Tolan, 1986).

Most of the section exposed in the Cascade Range consists of conformable layered flow sequences that form impressive cliffs. Such successions of flows represent well over 90 percent of the overall original volume of Columbia River basalt in the area. However, each hiatus in eruptive activity provided an opportunity for renewed stream erosion so that, as eruptions became less frequent in the later history of the Columbia River Basalt Group, drainage development became more pronounced. For this reason, the latest flows to cross the axis of the Cascade Range were either partially or wholly confined to canyons of rivers antecedent to the range.

The oldest known canyon of an ancestral Columbia River system was located in the central and southern part of the lowland that extended across the Cascade Range in Miocene time. This canyon was filled by the basal flow of the Frenchman Springs Member, the Ginkgo flow (Beeson, 1980, unpub. data; Hoffman, 1981; Beeson and others, 1985). The Frenchman Springs Member was erupted following what is thought to have been a brief lull in flood basalt volcanism. It was during this eruptive hiatus that the Vantage Member of the Ellensburg Formation was deposited in many parts of the Columbia Plateau. These deposits in part mark the positions of channels of the Columbia River as it re-established its course across the basalt plateau and eroded a canyon.

Peripheral erosion of the basalt appears to have been active during Grande Ronde Basalt time along the northern edge of the lowland (Tolan, 1982; Waters, 1973) where Columbia River Basalt Group rocks lap out against older underlying units. The central southern margin, by contrast, was a site of net deposition of volcanoclastic sediment that is now preserved as interbeds as much as 100 m thick in the basalt sequence (Anderson, 1978). Interbeds of similar thickness and composition are absent along the northern perimeter, which could imply that sediment was being efficiently removed by streams, as suggested by Tolan (1982), or that sediment was not equally available. Three relatively thin interbeds containing micaceous sand occur at Dog Mountain, which is located in the eastern Columbia River gorge in the deepest part of the Grande Ronde section. These deposits, in contrast to those along the southern margin, represent sediments of an ancestral Columbia River, deposited during brief intereruptive periods. Anderson (1978) has suggested that interbeds on the south, which include lahars, are a reflection of Cascadian volcanism contemporaneous with Columbia River Basalt Group activity. Volcanism during and after Frenchman Springs time has been subsequently demonstrated by Beeson (in Priest and

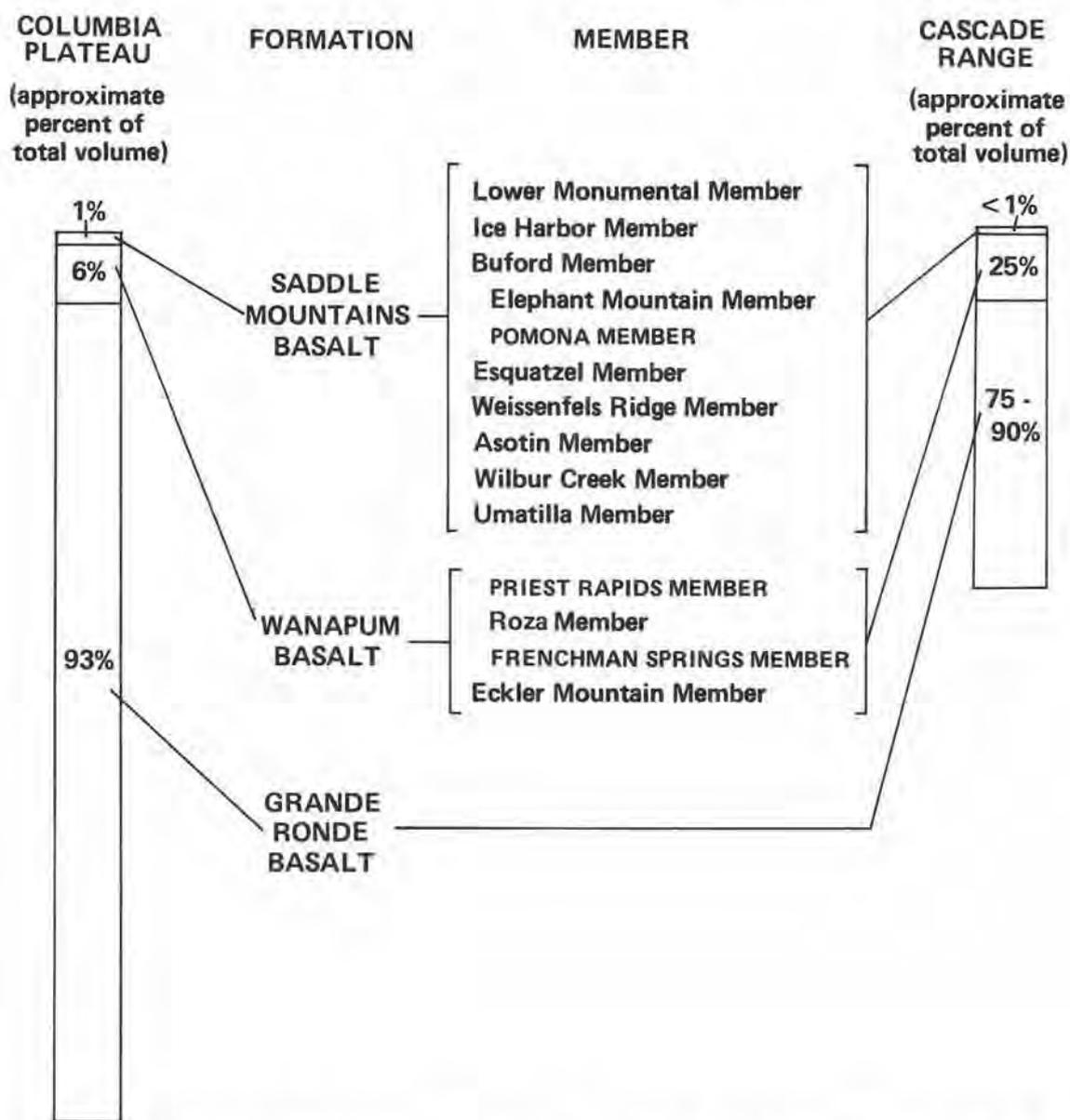


Figure 2.—Stratigraphy and volumetric relations of the Yakima Basalt Subgroup (Columbia River Basalt Group). Members that crossed the present Cascade Range in addition to the Grande Ronde Basalt are named in capital letters; indented names are those of all members that occur in the area discussed.

others, 1982). Andesitic lava flows and debris flows were observed to be interstratified with Columbia River basalt units near Mount Hood, suggesting that a volcanic center was present in the area.

The Wanapum Basalt marks an important transition from a time when eruptions were frequent and/or voluminous enough to overwhelm structural barriers and spread out over the broad lowland extending through the ancestral Cascades and a time when eruptions were less frequent and flows became more

restricted in extent in the lowland or were excluded entirely.

The purpose of this paper is to summarize and interpret paleodrainage data for the northeastern part of the paleotopographic lowland extending across the Cascade Range, starting with the Priest Rapids Member (13.5 m.y.) and ending with the early history of the Simcoe volcanic field (4.8-2.95 m.y. ago). Preliminary inferences can then be made about magnitude, location, and timing of deformation.

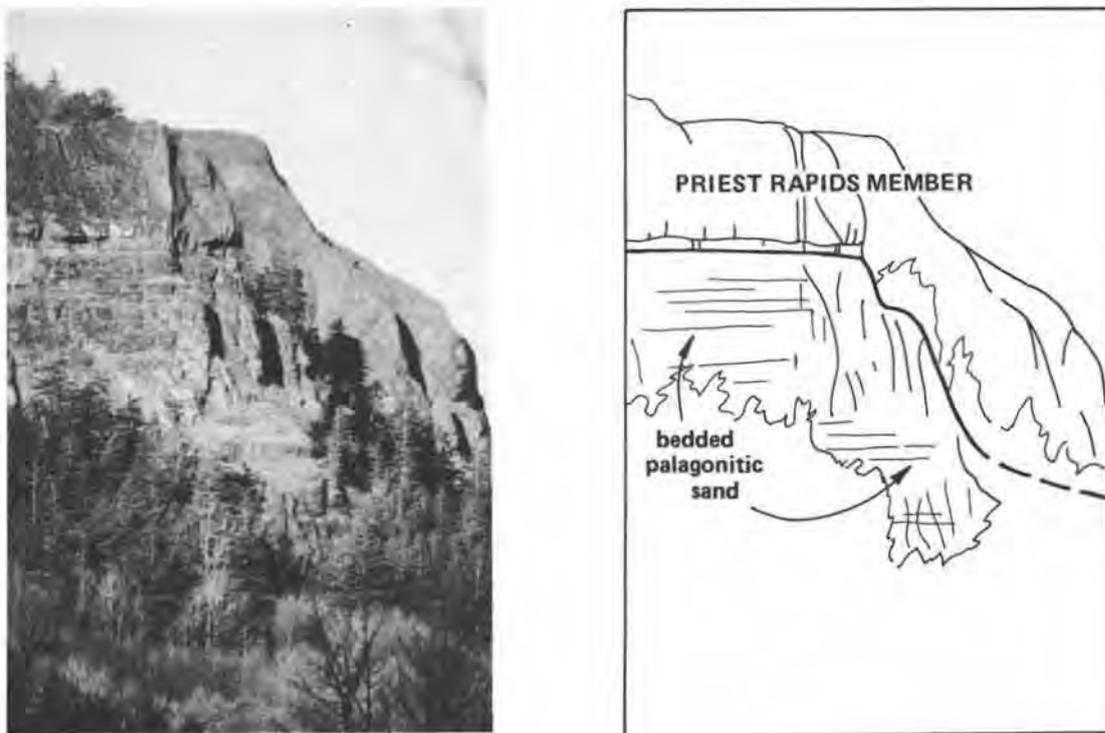


Figure 3.—Crown Point, Oregon. View to the west from the Columbia River Gorge Scenic Highway of Priest Rapids intracanyon flow and underlying palagonitic sand deposit. Location of Crown Point is shown on Figure 4.

PRIEST RAPIDS MEMBER INTRACANYON FLOW

Waters (1973) first suggested that rocks at Crown Point in the Columbia River Gorge of Oregon were an intracanyon flow of the Columbia River Basalt Group. He further speculated that this flow filled a channel of the ancestral Columbia River (Fig. 3). Research by Tolan and Beeson in the Crown Point area (Tolan, 1982; Tolan and Beeson, 1984) and by the authors working farther to the east and to the south confirms the presence of this ancestral drainage, extending from north of Mosier, Oregon, across the axis of the Cascade Range to Crown Point, a distance of approximately 90 km (Fig. 4).

The channel of this river was cut into basalt of the older Frenchman Springs Member and the formation underlying it, the Grande Ronde Basalt (Fig. 5). The Priest Rapids Member filled and thereby preserved this channel. Constructional volcanism in post-Columbia River Basalt Group time deeply buried the paleodrainage so that exposures are limited to areas where later deep stream erosion has occurred or where structural relief is significant.

Exposures studied by the authors are in the Bull Run watershed, along the West Fork of Hood River and the lower Hood River valley, and southwest of Mosier, Oregon (Fig. 5). The canyon defined by these exposures is 1 to 2 km wide and locally more than 200 m deep. The maximum depth is poorly known because the base of the canyon is not exposed over most of the area. Observed minimum depths are 225 m at Crown Point (Waters, 1973), 174 m in the Bull Run watershed (Vogt, 1981), 150 m at the West Fork of Hood River, 155 m in the lower Hood River Valley, 80 m southwest of Mosier, and 70 m northeast of Bingen, Washington. The general increase in minimum depth across the area from east to west may reflect a real increase in canyon depth, but this cannot be confirmed at present because of the lack of canyon bottom exposures.

The Priest Rapids intracanyon flow is of the Rosalia chemical type, the older of two variants in the Priest Rapids Member. The younger variant, the Lolo chemical type, is present near The Dalles to the east but has not been found by the authors in the Cascade Range. Beeson (Priest and others, 1982) has identified cuttings from the drill holes at Old Maid Flat, just west of Mount Hood, as Lolo chemical type. However,

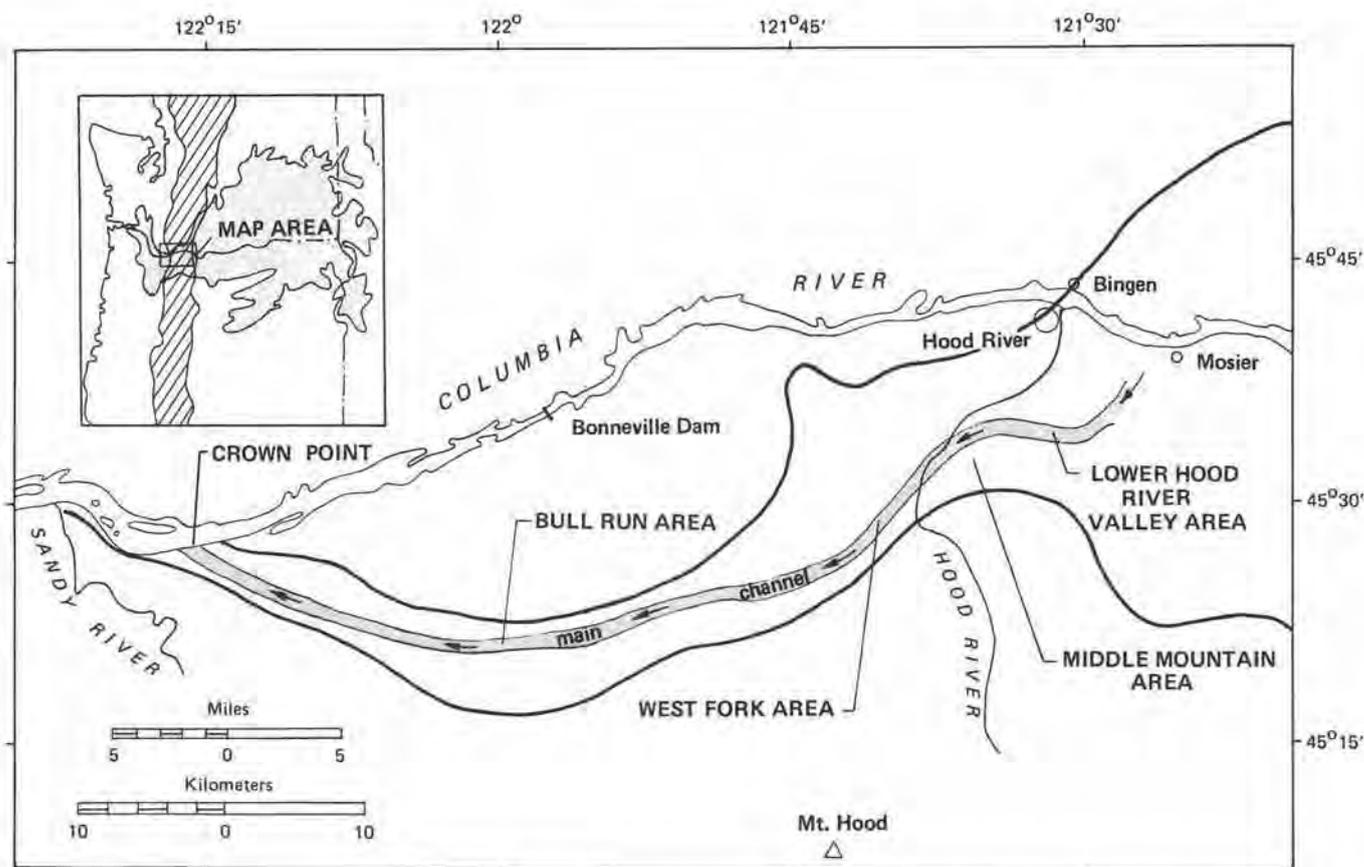


Figure 4.—Position of canyon filled by the Priest Rapids Member about 14.5 m.y. ago with respect to the present Columbia River. Heavy lines north and south of stippled main channel indicate area of overflow. Inset map shows distribution of Columbia River Basalt Group (stippled) with respect to Cascade Range (hachured).

no surface exposures are known. Compositions of both chemical types are listed in Table 1, and the differences between them are discussed in Swanson and others (1979b). The general Priest Rapids chemical type differs from underlying and adjacent Columbia River Group flows in having higher concentrations of MgO , TiO_2 , FeO , and P_2O_5 and a lower Al_2O_3 content (Table 1). In addition to distinctive chemistry, the Priest Rapids intracanyon flow has a number of other features that are useful in distinguishing it from other flows in the field, including reversed paleomagnetic polarity. This characteristic is particularly helpful where buttress unconformities separate this unit from lithologically similar, normally polarized flows at paleocanyon walls.

The Priest Rapids Member in the Cascade Range has significantly different intraflow structures than it has in the Columbia Plateau. Where it occurs as an intracanyon flow, its texture is more quenched, finer grained, and glassier than at the type locality (Mackin, 1961). The jointing of the intracanyon flow is domi-

nantly hackly, in contrast to the broad platy columns of the conformable Priest Rapids Member in the Columbia Plateau (Mackin, 1961, p. 24). The intracanyon flow has a well-developed entablature with closely spaced irregular joints that constitutes more than 95 percent of the overall flow thickness. A basal colonnade composed of vertical prismatic columns is commonly present (Fig. 3). Because of its relatively thick entablature, the Priest Rapids Member outwardly resembles many of the Grande Ronde Basalt flows in the area. However, contact relationships, flow facies, chemistry, and lithology do not resemble those of the Grande Ronde Basalt.

Lithologically, the non-intracanyon exposures of the Priest Rapids Member resemble certain flows of the older underlying Frenchman Springs Member that contain sparse plagioclase phenocrysts. Both can be relatively coarse grained, and both contain plagioclase phenocrysts of similar size. However, the differences discussed above in intracanyon exposures make identification of the flow relatively straightforward.

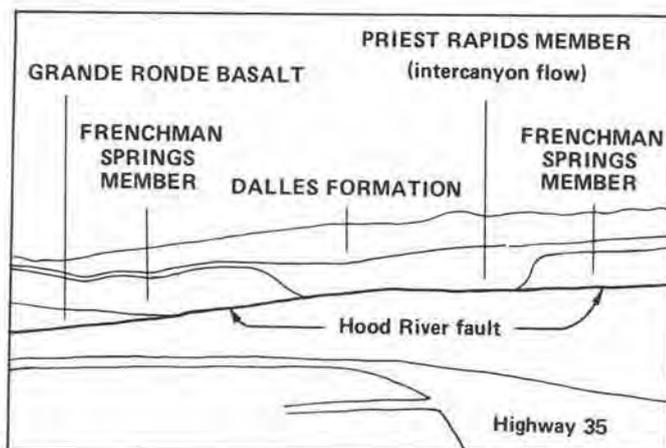


Figure 5.—Ridge on the east side of the lower Hood River valley where a Columbia River Basalt Group intracanyon flow and adjacent conformable overflow deposits are exposed. Channel is approximately 2 km wide at this point. View to the east from Oregon Highway 35 (Davis Road intersection). Area of photo is shown in Figure 4.

One of the most unusual aspects of the intracanyon flow is the presence of large volumes of sand-sized palagonitic detritus (Fig. 6) and the apparent invasion of this sand by flow lobes. Crown Point (Tolan, 1982), the two Hood River localities, and the exposures in the Mosier area all show this relationship.

Figure 7 shows several flow lobes of Priest Rapids Member interbedded with palagonitic sand in the West Fork area. Pillows are almost totally absent in these exposures and, where present, display little or no vesiculation. Where pillows occur, they are commonly surrounded with sediment (Fig. 8). The upper surface

Table 1.—Chemical analyses of some Miocene basalts in the Columbia Plateau. Average concentrations in weight percent. XRF whole-rock analyses performed by P. R. Hooper, Dept. of Geology, Washington State University. All analyses are normalized on a volatile-free basis with oxidation state of iron set at the arbitrary ratio of $\text{Fe}_2\text{O}_3/\text{FeO}=0.87$.

| Oxides | Olivine basalt of Simcoe Mountains | Elephant Mountain Member | Pomona Member | Priest Rapids Member* | Priest Rapids Member** | Frenchman Springs Member | Grande Ronde Basalt, High MgO | Grande Ronde Basalt Low MgO |
|--------------------------------|---|--------------------------------|------------------|-----------------------------|------------------------------|--------------------------------|--|--------------------------------------|
| SiO ₂ | 51.72 | 50.79 | 51.60 | 50.29 | 49.93 | 51.67 | 53.00 | 55.46 |
| Al ₂ O ₃ | 17.21 | 13.82 | 15.39 | 14.01 | 14.37 | 14.21 | 15.11 | 14.90 |
| FeO*** | 10.54 | 15.15 | 10.77 | 15.13 | 12.08 | 14.46 | 11.71 | 12.14 |
| MgO | 5.50 | 4.05 | 6.81 | 4.12 | 5.41 | 4.00 | 4.79 | 3.59 |
| CaO | 8.78 | 8.45 | 10.47 | 8.33 | 8.72 | 7.93 | 8.41 | 7.24 |
| Na ₂ O | 2.19 | 2.27 | 2.27 | 2.31 | 2.49 | 2.80 | 2.63 | 2.24 |
| K ₂ O | 0.86 | 1.29 | 0.65 | 1.33 | 1.07 | 1.38 | 1.16 | 1.79 |
| TiO ₂ | 2.69 | 3.45 | 1.63 | 3.52 | 3.07 | 2.99 | 1.85 | 2.01 |
| P ₂ O ₅ | 0.35 | 0.50 | 0.24 | 0.67 | 0.64 | 0.54 | 0.29 | 0.32 |
| MnO | 0.15 | 0.22 | 0.19 | 0.25 | 0.22 | 0.22 | 0.21 | 0.20 |
| No. of Analyses | (1) | (1) | (5) | (10) | (1) | (8) | (15) | (13) |

* Lolo chemical type

** Rosalia chemical type

*** Total iron as FeO



Figure 6.—Contact (at arrow) between canyon wall (left) and palagonitic sand of the Priest Rapids Member in the Bull Run watershed, Oregon.

of individual flow lobes is not vesicular (Fig. 9), indicating that the lava did not accumulate subaerially. A similar relationship exists above the Hood River fault on the east side of the Hood River valley, where alternating layers of bedded palagonitic sand and lava lobes occur. At both Hood River localities, sediment was apparently concentrated along the south side of the channel, as shown in Figures 4 and 7. A palagonitic sand deposit intruded by apophyses of lava is also present in road cuts along State Highway 14 in Washington, due north of Mosier, Oregon. All these localities clearly show that the sand deposits were invaded by lobes of lava under enough confining pressure to prevent vesiculation.

The palagonitic sand deposits differ in some important respects from deposits elsewhere in the Columbia Plateau referred to as hyaloclastites. Hyaloclastic deposits associated with flows other than the Priest Rapids Member occur along the foothills of the Washington Cascade Range, particularly west of Yakima and southwest of Ellensburg. These deposits consist of chaotic, glassy, basaltic debris mixed with palagonite that is poorly sorted and in few places shows evidence of bedding. Bedding where present is very irregular. Basalt fragments range in size from small pebbles to large boulders. These deposits were apparently formed when lava flowing toward the Cascades displaced water ahead of it with resultant phre-

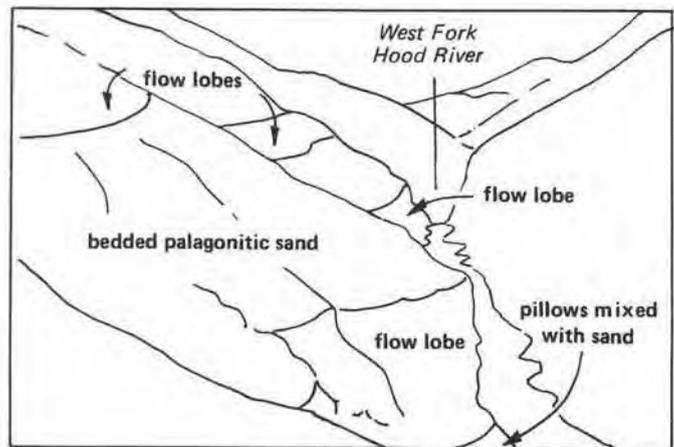


Figure 7.—Flow lobes and interbedded palagonitic sand of the Priest Rapids Member along the West Fork of Hood River. Two major lobes are shown. View to the north.



Figure 8.—Pillows surrounded by palagonitic sand at the base of the lowest flow lobe shown in Figure 7. Very little vesiculation is evident.

atic brecciation along the flow front. Debris formed in this manner was then pushed ahead of the flow and ultimately was carried onto its upper surface to produce the chaotic deposits observed. Vesiculation, as in the case of the intracanyon exposures, was suppressed by the confining pressure produced by as much as 150 m of hyaloclastic debris above the lava flow. Such deposits were produced contemporaneously with lava-flow advance, unlike the deposits associated with the Priest Rapids intracanyon flow, where palagonitic sand was deposited prior to the arrival of the lava flow.

Tolan and Beeson (1984) have suggested that the Priest Rapids flow, while advancing across the western part of the Columbia Plateau, entered numerous lakes located on the poorly drained basalt surface and that water from lakes and ponds was displaced toward the west along the axis of a southeast-trending topographic low. This scenario resembles that of the flows impinging upon the foothills of the Cascade Range except that, in the case of the Priest Rapids flow, there was an outlet channel leading to the ocean. This channel,



Figure 9.—Flow lobe of the Priest Rapids Member overlain by palagonitic sand in road cut along Columbia River (Washington Highway 14) across the river from Mosier, Oregon (Fig. 4). Lower photo is a close-up of the lower right side of the upper photo.

in the view of Tolán and Beeson, provided an egress for the accumulating water and debris which would have flooded the canyon downstream. Swanson and others (1979a, 1981), in reconnaissance mapping of the southwestern Columbia Plateau, delineated the distribution

of the various Columbia River Basalt Group units, including the Priest Rapids Member. This distribution suggests that the Priest Rapids Member filled a lowland trending west to southwest, centered near The Dalles, Oregon, with relatively uniform width of 35 to 40 km. Outside this low, approximately 50 km northeast of The Dalles, the Priest Rapids flow spread out over a much broader part of the plateau. The overall length of this topographic low was more than 70 km, and it is evident from mapping by the authors that the contemporaneous paleodrainage exited along the northern margin of the elongate depression.

The canyon begins at least 20 km to the east and starts along the northern edge of the low. This physiography supports the contention by Tolan and Beeson (1984) that water ponded ahead of the advancing flow in the lowland would have provided the necessary hydraulic mechanism for transport and redeposition of palagonitic debris downstream in the canyon. The reader is referred to their paper for a more complete discussion of this interpretation. The point we stress is that the differences between hyaloclastites of the Cascade foothills and hyaloclastic deposits of the Priest Rapids intracanyon flow are the result of entrapment of water in the first case versus free but restricted egress of water in the latter case. The intracanyon palagonitic sand was deposited by water moving well ahead of the lava flow. This accounts for both the lack of chaotic bedding in these deposits and the lack of clasts larger than sand and pebble size. The absence of palagonitic debris piled atop the Priest Rapids reflects the absence of a physiographic barrier without an outlet, in contrast to the Cascade foothills to the north.

Deposition of the palagonitic debris prior to the arrival of the lava flow within the canyon is further indicated by the presence of such sediment beneath the lava flow in the Crown Point area (Fig. 3) and in the Bull Run watershed and by the intrusive contact relations observed between lava lobes and sediment. Sediment was apparently pushed aside by the advancing flow or was overridden by it. This pushing aside resulted in the invasion (shallow intrusion) of relatively denser basaltic lava into wet sediment.

INTRACANYON FLOWS OF THE POMONA AND ELEPHANT MOUNTAIN MEMBERS

By about 12 m.y. ago, the ancestral Columbia River had re-established a channel through the late Miocene Cascade Range (Fig. 10) along the north edge of the Priest Rapids flow. This channel was inundated, as was the Columbia paleodrainage in

Priest Rapids time, and thereby was preserved by lava of the Pomona Member of the Saddle Mountains Basalt. The drainage originally extended entirely across the Columbia Plateau. The Pomona Member reached the Columbia River via a tributary stream extending eastward across the Pasco basin area into Idaho (Camp, 1981). It spread out in the central Columbia Plateau and covered large areas as an essentially conformable unit.

Mapping by Anderson and Bentley (Swanson and others, 1979b) has shown that the edge of the conformable Pomona Member in the central Columbia Plateau (Fig. 1) extends across the Horse Heaven Hills uplift of the anticlinal Yakima Fold Belt, approximately 8 km east of Satus Pass, Washington. The easternmost intracanyon exposure of this member occurs at Grayback Mountain at the Klickitat River 38 km to the west of Satus Pass. The area between the edge of the conformable Pomona and this first intracanyon exposure is largely covered by Pliocene and Pleistocene lavas of the Simcoe volcanic field, which obscure relations with underlying strata. The Klickitat River canyon is the only exposure in the area that is deep enough to reveal the intracanyon flow. The point at which the lava flow began to be completely confined to the river channel is therefore poorly constrained at present and is presumed to lie at some point north of Satus Pass near the mapped edge of the conformable Pomona Member.

Previous hypotheses concerning the path of flow of the Pomona Member across the southwestern part of the Columbia Plateau have suggested that it followed a low area that was approximately aligned with The Dalles-Umatilla syncline (Swanson and others, 1979b). This conclusion was based on the presence of the Pomona Member in the Arlington area of Oregon and Washington and in the Mosier syncline, 85 km to the west (Fig. 10). A connection between the two localities was hypothesized, pending field verification. The results of regional mapping in the area by Anderson (Swanson and others, 1981) have ruled out such a connection. Similarly, detailed mapping of the next Synclinal basin to the north, the Klickitat Valley, has ruled out the presence of a Saddle Mountains intracanyon flow in that area (Anderson, 1987; Sylvester, 1978; Swanson and others, 1979a).

Excellent exposures of the Pomona Member at several localities along the Columbia River downstream from Mosier, Oregon, together with the exposures near Grayback Mountain at the Klickitat River, confirm a paleodrainage extending from an area north of the Horse Heaven Hills uplift to the Mosier syncline and then west across the axis of the present Cascade Range. Notable exposures are present in the

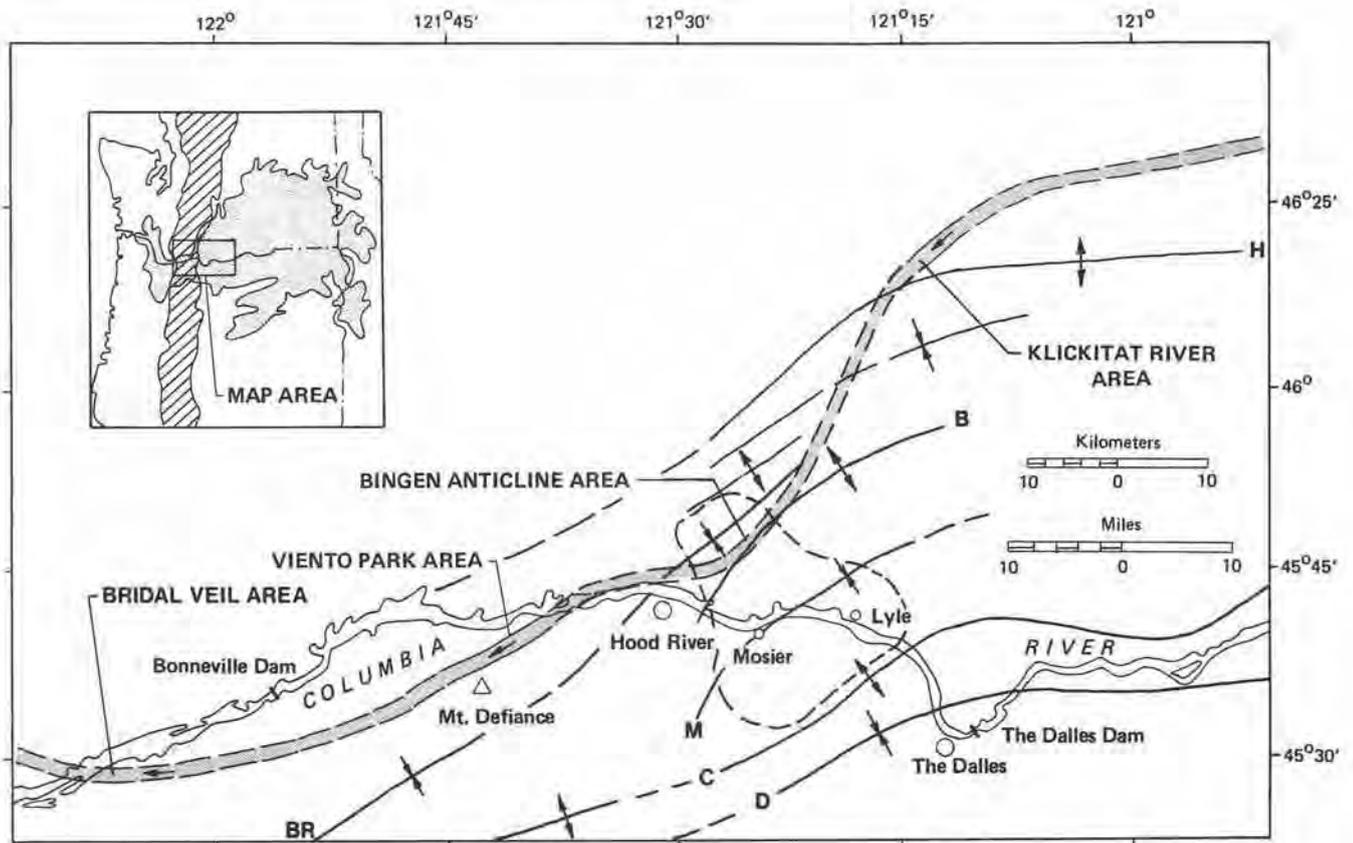


Figure 10.—Approximate position of the main Columbia River paleocanyon (stippled) filled by the Pomona Member about 12 m.y. ago with respect to the present Columbia River and of folds of the Yakima Fold Belt. Dashed outline indicates area of Pomona Member overflow on either side of main channel. Inset map shows distribution of Columbia River Basalt Group (stippled) with respect to the Cascade Range (hachured). BR, Bull Run syncline; B, Bingen anticline; M, Mosier syncline; C, Columbia Hills anticline; D, The Dalles—Umatilla syncline; H, Horse Heaven Hills anticline.

area west of the Grayback Mountain area (1) at the crest of the Bingen anticline, 10 km north of Mosier, (2) at Mitchell Point near Viento Park (Anderson, 1980), and (3) in the Bridal Veil area to the west (Tolan and Beeson, 1984; Tolan, 1982) (Fig. 10). Other exposures are present high on the walls of the south side of the Columbia River Gorge between Mitchell Point and the Bridal Veil area. All exposures are located north of the Priest Rapids intracanyon flow.

When compared to the Priest Rapids Member, the Pomona Member is lithologically more distinct. In the Pomona flow, the phenocrysts of plagioclase are commonly single crystals that appear to be either elongated or equant in hand sample; crystals are from 0.25 to 1.0 cm long. Olivine usually occurs as glomerocrysts. Equant crystals commonly appear to be zoned, with dark cores and, in some specimens, appear to be exsolved along their perimeters. These characteristics set the Pomona Member apart from underlying phyric

flows such as the Frenchman Springs Member, which lacks olivine and has much larger plagioclase crystals that occur as glomerocrysts 1.5 to 5.0 cm in diameter. The abundance of plagioclase phenocrysts in the Pomona Member is highly varied. Most intracanyon exposures appear to be more abundantly phyric than conformable exposures. However, exceptions to this pattern are notable locally in both the Klickitat River and the Mitchell Point areas. This has led the authors to conclude that the Pomona Member is variably phyric due to crystal stratification within the flow. Such variably phyric character, where phenocryst abundance increases or decreases vertically or horizontally within a single flow, is not uncommon among other phyric flows of the Columbia River Group, particularly flows of the Wanapum Basalt. Such variation is not strictly a function of crystal settling, because crystals are not exclusively concentrated near the bases of flows. Where the Pomona Member is abundantly phy-