Mineral compositions *(continued)*

	174-8a				174-24			174-33a				174-36c		
	B	G	M	P	G	H	P	B	G	M	P	G	H	P
Si	5.95	5.95	6.88	–	5.86	6.39	–	5.93	5.95	6.53	–	5.93	6.53	–
Al^4	2.05	–	1.12	–	–	1.61	–	2.07	–	1.47	–	–	1.47	–
Al^6	1.60	3.96	4.34	–	4.03	1.01	–	1.76	3.97	4.89	–	4.03	0.94	–
Ti	0.20	–	0.02	–	–	0.08	–	0.24	–	0.05	–	–	0.10	–
Fe	3.32	3.55	0.43	–	3.95	2.01	–	2.74	4.41	0.11	–	3.85	1.89	–
Mg	1.72	0.31	0.37	–	0.73	2.00	–	2.27	0.83	0.13	–	1.01	2.14	–
Mn	0.02	0.22	–	–	0.37	0.03	–	0.01	0.14	–	–	0.20	0.02	–
Ca	–	2.08	–	0.37	1.18	1.87	0.38	–	0.24	–	0.34	1.02	1.79	0.36
Na	0.04	–	0.09	0.63	–	0.34	0.65	0.11	–	0.38	0.68	–	0.49	0.64
K	2.06	–	2.09	0.01	–	0.10	0.01	1.68	–	1.77	–	–	0.10	–

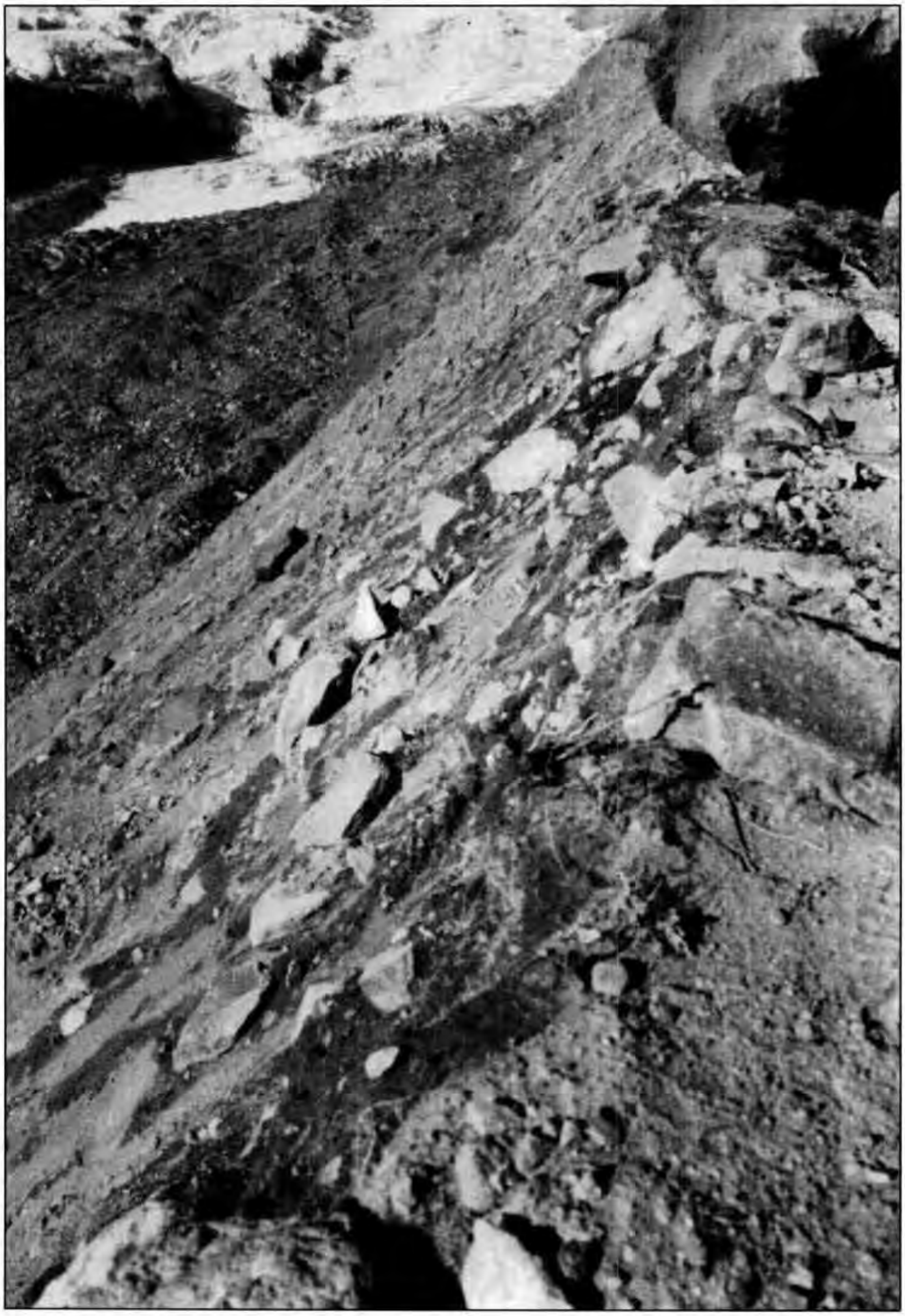
	174-45b				174-118a				OHM20			ODM22			RT48A58			
	B	G	M	P	B	G	M	P	B	G	P	B	G	P	B	G	M	P
Si	5.96	5.91	6.67	–	5.91	5.98	7.05	–	5.47	5.98	–	5.55	6.00	–	5.87	6.00	6.75	–
Al^4	2.04	–	1.33	–	2.09	–	0.95	–	2.53	–	–	2.45	–	–	2.13	–	1.25	–
Al^6	1.69	4.03	4.81	–	1.67	3.99	0.83	–	0.86	3.96	–	0.89	3.95	–	1.71	3.98	4.85	–
Ti	0.24	–	0.05	–	0.21	–	0.02	–	0.24	–	–	0.14	–	–	–	–	0.05	–
Fe	2.21	4.35	0.10	–	2.40	4.18	1.78	–	1.97	4.22	–	1.64	4.07	–	2.70	4.38	0.17	–
Mg	–	0.84	0.18	–	2.75	1.04	2.69	–	2.66	0.97	–	3.08	1.29	–	2.25	0.41	0.22	–
Mn	–	0.06	–	–	0.01	0.19	0.02	–	0.01	0.44	–	0.01	0.05	–	0.01	0.11	–	–
Ca	–	0.89	–	0.40	–	0.61	1.83	0.25	–	0.44	0.28	–	0.66	0.38	–	1.12	–	0.27
Na	0.07	–	0.40	0.59	0.03	–	0.33	0.76	0.07	–	0.70	0.07	–	0.62	0.08	–	0.14	0.74
K	1.71	–	1.73	–	1.91	–	0.03	–	1.64	–	–	1.65	–	–	1.88	–	1.59	–

B. Aluminum-in-hornblende (Johnson and Rutherford, 1989)

Sample number	Pluton	Latitude (degrees and minutes)	Longitude (degrees and minutes)	Al (total)	Pres. (kb)	Ref.
118-21a	Marble Creek	48 33.4	121 16.5	2.73	8.1	7
119-139b	Marble Creek	48 31.8	121 15.7	2.77	8.3	1
164-20	Jordan Lale	48 29.5	121 21.4	0.77	<2	2
164-35	Chaval	48 24.9	121 23.6	1.85	4.4	2
164-46b	Hidden Lake Peak	48 28.7	121 10.8	1.78	4.1	2
169-93b	Eldorado	48 31.3	121 2.1	1.76	4.0	3
169-219	Eldorado	48 31.3	121 7.1	1.81	3.7	3
169-B1	Eldorado	48 30.1	120 3.8	1.84	4.3	3
169-E4C	Eldorado	48 34.4	121 11.5	1.99	5.0	3
169-HL5	Hidden Lake Peak	48 30.6	121.11.6	1.70	3.7	2
174-49	Bench Lake	48 21.3	121 12.3	2.53	7.3	5
174-122a	Jordan Lake	48 27.4	121 18.1	0.99	<2	5
174-130	Jordan Lake	48 29.6	121 19.5	0.91	<2	5
OHM 66	Chaval	48 19.6	121 19.0	2.18	5.8	6

C. NaM_4-in-actinolite (Brown, 1977)

Sample number	Unit	Latitude (degrees and minutes)	Longitude (degrees and minutes)	Al^4	NaM_4	Pres. (kb)	Ref.
118-15e	Cascade River Schist	48 32.4	121 19.6	0.61	0.15	3–4	7



Upvalley view of Nisqually Glacier (upper left) on Mount Rainier from the crest of the left lateral moraine. The rocky rubble in the upper left covers much of the glacier's surface. During the Little Ice Age (about A.D. 1250 to the mid-1800s), the surface of Nisqually Glacier reached to at least the top of the moraine, and the glacier's terminus extended at least 1 km farther down the valley. Photo by Timothy J. Walsh, 1991.

Cenozoic Unconformity-Bounded Sequences of Central and Eastern Washington

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ABSTRACT

Remnants of four interregional Cenozoic unconformity-bounded sequences are preserved in, and extend beyond, central and eastern Washington. The Eocene (~55–36 Ma) Challis sequence consists of arkosic and volcanic rocks. The name Kittitas sequence is proposed for the 36–20 Ma intermediate to felsic volcanic rocks best exposed in the southern Cascade Range. The most voluminous lithostratigraphic unit of the Walpapi sequence (18–2 Ma) is the Columbia River Basalt Group. The post-2 Ma sequence is informally referred to as the High Cascade, but it also contains alpine and lowland glacial deposits. Each of these four sequences rests on pre-Cenozoic basement rocks somewhere in Washington.

At least eight smaller regional sequences occur within the Challis. The locally varied preservation of these regional Challis sequences has fostered models of local, commonly aulacogenic, depositional basins. However, in northeastern Washington and adjacent British Columbia some of the same sequences occur in more than one graben or half-graben (Pend Oreille River valley, First Thought Mountain, Republic, Toroda Creek, the American part of the Okanogan Valley, and the British Columbia part of the Okanagan Valley). Some of these sequences persist westward and are overlain by younger regional Challis sequences at Island Mountain, the Chiwaukum graben, Roslyn basin, Manastash Ridge, and the Rimrock Lake inlier. Many of the regional sequences also occur along the eastern edge of the Puget Lowland near Kelso, Centralia, Carbonado, Seattle, and Bellingham.

This sequence stratigraphy argues against the popular model of sub-regional depositional basins, isolated Tertiary volcanic centers, and a continuously active Cascade magmatic arc since 36 Ma. The thermal maturity of sedimentary rocks and their contained coals appears to be at least partially sequence-dependent. Further, the presence of regional and interregional unconformities indicates that the volume, extent, and rate of extrusion of the Columbia River basalts have been routinely underestimated.

The Challis sequence overlaps the Proterozoic and Paleozoic sequences indigenous to North America, as well as the Intermontane, Insular, and the North West Cascades superterranes. The Challis sequence does not seem to overlie the Coast Range superterrane, the docking of which probably caused the sub-Kittitas unconformity. The sub-Walpapi unconformity probably records the initial extension of the Basin and Range Province and, possibly, initial crustal uplift and erosion over the mantle plume that is now the Yellowstone hotspot. If the Hoh assemblage is subducted beneath the metasedimentary rocks of the core of the Olympic Mountains, this subduction may coincide with an intra-Walpapi unconformity above the Columbia River Basalt Group. Initiation of the offshore Cascadia subduction zone appears to coincide with the sub-High Cascade unconformity.

INTRODUCTION

Purpose

The purpose of this paper is to apply sequence stratigraphy to the predominantly nonmarine Cenozoic sedimentary and volcanic cover sequences of central and eastern Washington. I will show that four interregional unconformity-bounded sequences (UBS) younger than about 55 Ma exist. Each of these has less extensive regional sequences. An interregional unconformity is herein defined as one that has been recognized in localities more than 580 km apart (arbitrarily chosen as the approximate east-to-west length of Washington).

The interregional and regional sequences are laterally discontinuous due to multiple periods of uplift, post-depositional folding and faulting, and erosion; thus, they are preserved in structural basins. They also are locally obscured by overlying units. In contrast, most presently accepted models assert that the Cenozoic rocks, especially those of the Challis sequence, were deposited in local basins, few of which were more than 100 to 125 km long.

Figure 1 shows that interregional sequence stratigraphy presented here is not entirely new. Wheeler and Mallory (1963, 1970) were the first to suggest that the Cenozoic successions of the Pacific Northwest could be divided into a few interregional UBS. Armentrout (1987) applied this model to southwestern Washington and adjacent Oregon.

My interest was stimulated by having Harry Wheeler and Stan Mallory as colleagues and by noting the work of others who were not necessarily thinking in terms of regional and interregional UBS. Pearson and Obradovich (1977) made the provocative observation that the Eocene sedimentary and volcanic rocks of the Republic graben of northeastern Washington occur throughout northeastern Washington and adjacent British Columbia. Armstrong (1978) noted that the radiometric ages of volcanic rocks in the Pacific Northwest are episodic: 55–36 Ma, 36–20 Ma,

Wheeler and Mallory (1963 & 1970) Sequences	Armstrong (1979) Volcanic Episodes	Armentrout (1987) Sequences	This Paper Sequences	Examples of Lithostratigraphy	Age Ma
unnamed	High Cascade	V	High Cascade	Vashon Drift Logan Hill Fm.	
					2
Walpapi	unnamed Columbia	IV	Walpapi	Columbia River Basalt Group	
					19
unnamed	Cascade	III	Kittitas	Fifes Peak Fm. Ohanapecosh Fm.	
					36
unnamed	Challis	II	Challis	Wenatchee Fm. Roslyn Fm. Teanaway Fm. Taneum Fm. Swauk Fm.	
					55

Figure 1. Comparison of past and present regional stratigraphic models. Fm, Formation.

20–13 Ma, and younger than 13 Ma; he called these the Challis, Cascade, Columbia, and High Cascade episodes, respectively. (See Fig. 1.) The interregional extent of at least the first three of these episodes seemed to be confirmed by Proffett's description (1977, fig. 3, table 1) of the Tertiary stratigraphy of western Nevada. My colleague Gresens (1976) described the Wenatchee Formation and its unconformable position on the Eocene clastic rocks and pre-Tertiary gneisses of the Wenatchee area of central Washington. Gresens (1981) then rediscovered an unconformity noted by Wheeler and Mallory (1963) that extends throughout the American part of the Cordillera and into Mexico; this he named the Telluride erosion surface. Significantly, Hanneman and Wideman (1991) used sequence stratigraphy to subdivide and correlate Cenozoic continental lithostratigraphic units of southwestern Montana, and they (1991, fig. 8) correlated the resultant five sequences with those in southwestern Washington and the Great Plains.

The Cenozoic sequence stratigraphy described here is the outgrowth of research initiated with R. J. Stewart in the early 1980s. Since then, the models for local depositional basins have grown ever more popular (for example, see Tabor and others, 1984; Johnson, 1985; and numerous examples cited below). Nonetheless, during the past dozen years as more regional mapping and compilations have become available and as fission-track and radiometric dates have become more numerous, the evidence for interregional and regional UBS has increased.

Scope

Because Armentrout (1987) discussed the Cenozoic sequence stratigraphy of southwestern Washington and adjacent Oregon, my discussion will concentrate on the area of Washington east of Interstate Highway 5. The similarity of the sequences I describe to the four youngest sequences recognized by Armentrout demonstrates the continuity of the interregional sequences across the state.

This paper consists of four major parts. Because sequence stratigraphy has not been practiced by previous regional mappers and compilers working in Washington (Tabor and others, 1982a, 1987; Frizzell and others, 1984; Walsh and others, 1987; Swanson and others, 1989; Reidel and others, 1989a, 1989b; Tolan and others, 1989; Smith, 1989; Stoffel and others, 1991; Schuster, 1992), the first section is a brief review of the topic. Then I discuss each of the four major interregional sequences: Challis, Kittitas, Walpapi, and High Cascade. Next, I show that the Challis sequence in Washington and adjacent British Columbia is composed of at least eight smaller regional UBS. Finally, I consider a few of the implications of sequence stratigraphy to the geology of Washington.

CONCEPTS OF SEQUENCE STRATIGRAPHY

Table 1 reviews some of the major aspects of sequence stratigraphy. Fairly recent references that could be consulted for more extended discussions are Mitchum and others (1977), International Subcommittee on Stratigraphic Classification (ISSC) (1987), Sloss (1988a, b), and Cheney and others (1990).

Table 1 is biased toward sequences in the cratonic environment, rather than those in marginal and deep marine settings. On continental margins and in deep-sea environments sequences tend to be thicker and more conformable. In the cratonic environment, unconformities are more numerous and geographically extensive, so that stratigraphic patterns and thicknesses are preservational, not depositional (Wheeler, 1958, 1963).

Several figures in this paper are area–time diagrams promoted by Wheeler (1958) and increasingly known as Wheeler diagrams. Conventional stratigraphic columns and cross sections are area–depth diagrams that show the thickness of each unit. In contrast, the vertical axis of a Wheeler diagram is geologic time, not thickness. Gaps between formations in such diagrams indicate intervals of time for which no physical (lithostratigraphic) unit exists; these gaps are caused by non-deposition, erosion, or both. Technically, Wheeler diagrams should only be constructed from drill holes or individual traverses across strike. In practice, the columns shown in some of the figures are composites for a given area; this facilitates comparison of units in just a few columns. However, composite columns make sequences appear to be more continuous than they really are.

Table 1. Some elements of sequence stratigraphy

PRINCIPLES	REMARKS
(1) An unconformity-bounded sequence can be of interregional extent.	(1) Sequences may contain many lithostratigraphic units (that is, formations). Formations rarely are of interregional extent.
(2) Locally derived units do not disprove the existence of interregional sequences.	(2) Local relief can exist even on cratonic interregional unconformities, especially in volcanic environments.
(3) Interregional sequences of sedimentary, volcanic, or volcaniclastic rocks are not restricted to areas underlain by pre-Phanerozoic sialic crust.	(3) The North American craton grew westward with time; so the pre-Tertiary basement in Washington is neither entirely pre-Phanerozoic nor crystalline. Furthermore, sequences can be traced into marginal successions bordering cratons.
(4) The size of an unconformity is measured by its lateral extent and the amount of section it eliminates somewhere in that extent.	(4) Size is not measured by the amount of angular discordance in outcrop. In this paper, an interregional sequence is one that somewhere rests on pre-Tertiary rocks and has a strike length greater than the east-west length of Washington (580 km).
(5) Large unconformities can be difficult to recognize in the field.	(5) The angular discordance of some unconformities can only be recognized on 1:50,000- to 1:125,000-scale geological maps.
(6) A regional unconformity indicates flat paleotopography and that the subsequent depositional system most likely was laterally extensive. Concepts such as the "ancestral Cascades" or the "ancestral Columbia River" cause misunderstandings.	(6) Present topography and present fluvial systems are not keys to the past.
(7) Because cratonic sequences have upper bounding unconformities, original depositional thicknesses are rarely preserved.	(7) Changes in thickness are as likely to be erosional (preservational) as depositional, especially in the uppermost units of a sequence. Depositional centers and axes are difficult to reconstruct. A minimum of syndepositional paleotopography need be assumed if changes in thickness are erosional.
(8) Due to upper bounding unconformities, the original extent and continuity of a sequence is rarely preserved.	(8) The present geographic outline of a succession rarely matches its original depositional extent. Cratonic stratigraphic patterns are preservational, not depositional.
(9) Although bounding unconformities commonly are time-transgressive, the sequence between them represents a slice of time not represented by rocks in sequences above and below.	(9) Unconformity-bounded sequences (not chronostratigraphic units) are the natural subdivisions of the stratigraphic record and are idea for correlation. Lithostratigraphic units on opposite sides of an unconformity cannot be facies equivalent (correlative) even if they do contain similar minor lithologies).
(10) Sequences commonly have distinctive or characteristic lithologies.	(10) Sequences commonly represent distinct tectonic cycles. Reports of "rapid" changes of cratonic facies commonly imply an unrecognized unconformity or other forms of miscorrelation.
(11) Any report on a thick stratigraphic section that has not considered unconformities is suspect.	(11) A reported thickness in excess of 1 to 3 km should prompt the search for unrecognized unconformities, structural repetitions of the section, or other forms of miscorrelation.
(12) Differential erosion beneath unconformities can cause distinctly different neighboring successions.	(12) Distinctly different neighboring successions do not necessarily imply separate (local) depositional basins. Composite sections are rarely preserved in nature.
(13) Paleocurrent directions derived from cross bedding and imbricated clasts are variable.	(13) More reliable indicators of depositional directions are the distribution of facies, the overall strike of fluvial channels, mineralogically determined provenance, and coarsening- (or fining-) upward parasequences.
(14) Apparently local patterns of facies, isopachs, and paleocurrent indicators can be regional.	(14) Each pattern must be mapped to test whether it is more extensive than the area in which it was first described or is spatially related to the proposed source.

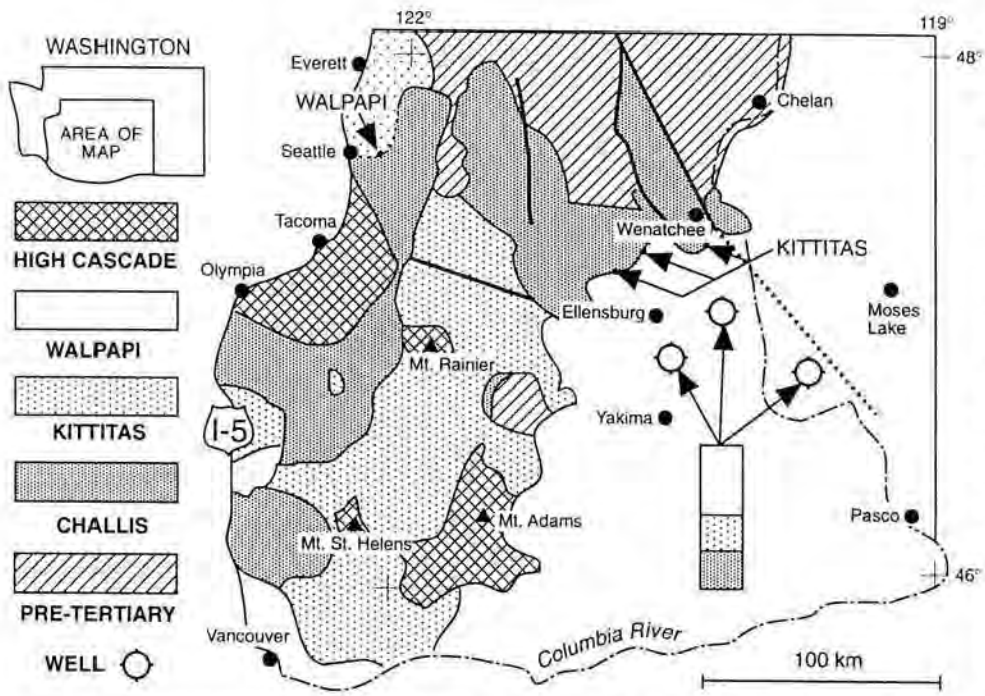


Figure 2. The distribution of four interregional sequences in south-central Washington. The three oldest sequences in deep holes near the Columbia River are shown schematically. Covered contacts are dotted; solid triangles are Pleistocene volcanoes. Principal sources of data are Walsh and others (1987), Catchings and Mooney (1988), Campbell (1989), Yount and Gower (1991), and Schuster (1992). I-5, Interstate Highway 5.

In this paper, a major unconformity may be conveniently defined as one that somewhere causes the overlying section to rest on pre-Cenozoic intrusive or metamorphic rocks. An interregional unconformity as herein defined is one with a strike length >580 km. Using these criteria, four major Cenozoic, interregional sequences occur in central and eastern Washington (Fig. 1).

INTERREGIONAL SEQUENCES

Introduction

The presence of pre-Cenozoic crystalline rocks in the northern (McGroder, 1991) and southern (Miller, 1989) Cascade Range of Washington demonstrates the cratonic setting of the Cenozoic sequences. Figure 2 shows that all four sequences occur in south-central Washington. The unconformable nature and stratigraphic superposition of the Challis, Kittitas, and Walpapi sequences are particularly obvious near Cle Elum (Fig. 3).

Challis Sequence

As shown in Figures 2 and 3, Eocene arkosic and volcanic rocks occur unconformably above pre-Cenozoic crystalline rocks and unconformably below Oligocene or Miocene rocks. These Eocene rocks must, therefore, represent one or more UBS. Armstrong (1978) referred to the interval represented by these rocks as the "Challis volcanic episode" and noted that the volcanic assemblage was "referred to as the Challis Volcanics in many areas...." Armentrout (1987) referred to this succession of rocks in southwestern Washington as Sequence II. I suggest that this sequence of arkosic and volcanic rocks between 55 and 36 Ma be formally designated the Challis sequence.

Kittitas Sequence

The post-Challis sequence of intermediate to felsic volcaniclastic rocks is best exposed in the southern Cascade Range (Fig. 2), where it consists of the Oligocene Ohanapecosh Formation and the unconformably overlying upper Oligocene and lower Miocene Fifes Peak Formation (Swanson and others, 1989). Sources for these rocks were coeval granitic plutons (Tabor and others, 1982b; Walsh and others, 1987) along the axis of the present Cascade Range.

This 36–20 Ma sequence is not restricted to the Cascade Range, as is commonly supposed. Figure 2 shows it at the surface and in the subsurface (Campbell, 1989) east of Ellensburg and Yakima; west of the range it occurs near Seattle (Yount and Gower, 1991). Wheeler and Mallory (1963) and Armstrong (1978) noted that the John Day Formation of north-central Oregon also belongs in this sequence. In fact, this sequence is widespread in the western United States (Wheeler, 1956; Wheeler and Mallory, 1963; Gresens, 1981).

Armstrong (1978) referred to the area represented by these volcanic rocks as the "Cascade volcanic arc". Other authors have used such terms as Cascade assemblage, Cascade suite, and Cascade volcanic rocks. Even worse, the name Cascade plutons has been applied to rocks younger than 20 Ma. Such usages reinforce erroneous beliefs that the Cascade magmatic and volcanic arc has been continuously (instead of episodically) active since 36 Ma, that the Cascade Range has been a topographic high since then, and that the rocks of this UBS are limited to the Cascade Range.

Although Armstrong's original temporal definition of the Cascade episode does coincide with the Oligocene to Miocene interregional sequence, to rectify the various and misleading uses of "Cascade", a different name should be given to this sequence. To avoid confusion with chronostratigraphic and lithostratigraphic names, Sloss (1963) chose American Indian or Spanish names for interregional sequences. The counties in Washington that have Indian names and that are extensively underlain by this sequence are Yakima, Skamania, and Kittitas; of these only Kittitas has not been seriously pre-empted as a stratigraphic name (cf. Tabor and others, 1982a; Walsh and others, 1987). The term Kittitas system was abandoned when the Swauk and Roslyn Formations were named (Russell, 1900, p. 118); Kittitas Drift is used for Pleistocene deposits in the area of Figure 3 (Waitt, 1979; Tabor and others, 1982a; Walsh and others, 1987). Because the term Kittitas sequence seems easily distinguishable from these previous usages, I informally use it for the Oligocene to Miocene unconformity-bounded sequence, pending agreement on a better name.

Walpapi Sequence

Although an interregional mid-Miocene unconformity has long been known (Wheeler, 1956; Wheeler and Mallory, 1963, 1970), its presence in Washington is still underappreciated. In central and eastern Washington the most voluminous lithostratigraphic unit above this unconformity is the Columbia River Basalt Group (CRBG). Wheeler and Mallory (1970) referred to the deformational event marked by this sub-CRBG unconformity as the Ochocoan orogeny and the sequence above the unconformity as the Walpapi sequence. This is Armentrout's (1987) Sequence IV of southwestern Washington and adjacent Oregon (Fig. 1).

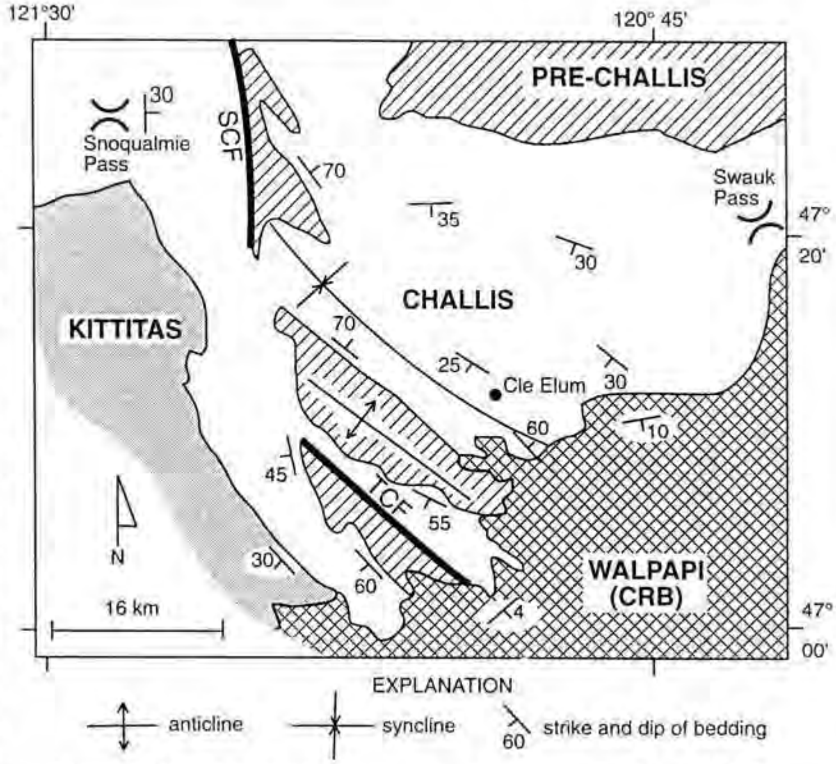


Figure 3. Distribution of the interregional sequences in the central Cascade Range of Washington. Principal sources of data are Foster (1960), Tabor and others (1982a), and Frizzell and others (1984). CRB, Columbia River Basalt Group; SCF, Straight Creek fault; TCF, Taneum Creek fault.

Figures 2, 3, and 4 show that the CRBG overlies basement rocks and the Challis and Kittitas sequences. The two lowest units, the Imnaha and Grande Ronde Basalts, comprise the bulk of the CRBG (Tolan and others, 1989) and were extruded between 17.3 and 15.6 Ma (Baksi, 1989).

Sources of most of the CRBG were northwesterly trending dikes in the area of adjacent Idaho, Washington, and Oregon (Hooper and Swanson, 1987; Tolan and others, 1989; Swanson and others, 1989). The lavas flowed westward, and some reached the sea (Hooper and Swanson, 1987). Volcaniclastic material between some basalts originated in the area now occupied by the Cascade Range, but epiclastic rocks between the basalts were derived from at least as far east as Idaho (Fecht and others, 1987; Reidel and others, 1989b; Swanson and others, 1989). In southeastern Oregon, post-CRBG eruptive centers of felsic volcaniclastic rocks in the Walpapi sequence migrated northwestward with time (MacLeod and Sammel, 1982; Swanson and others, 1989).

The regional unconformity above the Walpapi sequence also is underappreciated, perhaps because the CRBG is commonly regarded as undeformed. The large number of folds and faults in the CRBG (Walsh and others, 1987; Campbell, 1989; Reidel and others, 1989b; Tolan and Reidel, 1989) clearly demonstrates deformation. In addition, the CRBG is regionally arched along the eastern margin of the Cascade Range (Tabor and others, 1982a; Campbell, 1989; Swanson and others, 1989). For example, the map of Tabor and others (1982a) shows that the base of magnetostratigraphic unit N_2 of the CRBG has an altitude of 350 m on the west side of the Columbia River south of Wenatchee; 26 km to the west at Mission Ridge this contact is at 1,800 m. Similarly, D. A. Swanson (U.S. Geological Survey, oral commun., 1991) has pointed out that from the Columbia River on the Oregon border to the west of Yakima (Fig. 2) the structural relief of N_2 is almost 2 km.

The well-documented synclinal map pattern of the formations of the CRBG (Choiniere and Swanson, 1979, fig. 1) also attests to regional folding. Most observers probably regard this pattern as being due to successive fillings of a paleotopographic low or basin (cf. Reidel and others, 1989a,b). However, the presence of unconformities at the base of and in each of the two youngest formations, Wanapum Basalt and Saddle Mountains Basalt (Reidel and others, 1989b, fig. 1), favors episodic deformation.

High Cascade Sequence

In the Cascade Range of southern Washington, rocks younger than 2 Ma rest unconformably on Walpapi and older rocks including pre-Cenozoic metamorphic rocks (Walsh and others, 1987). Sources of the volcanic and volcaniclastic rocks of the sequence are a few stocks and volcanic centers, including the present stratovolcanoes in the Cascade Range. However, glacigene sediments are the most widespread component.

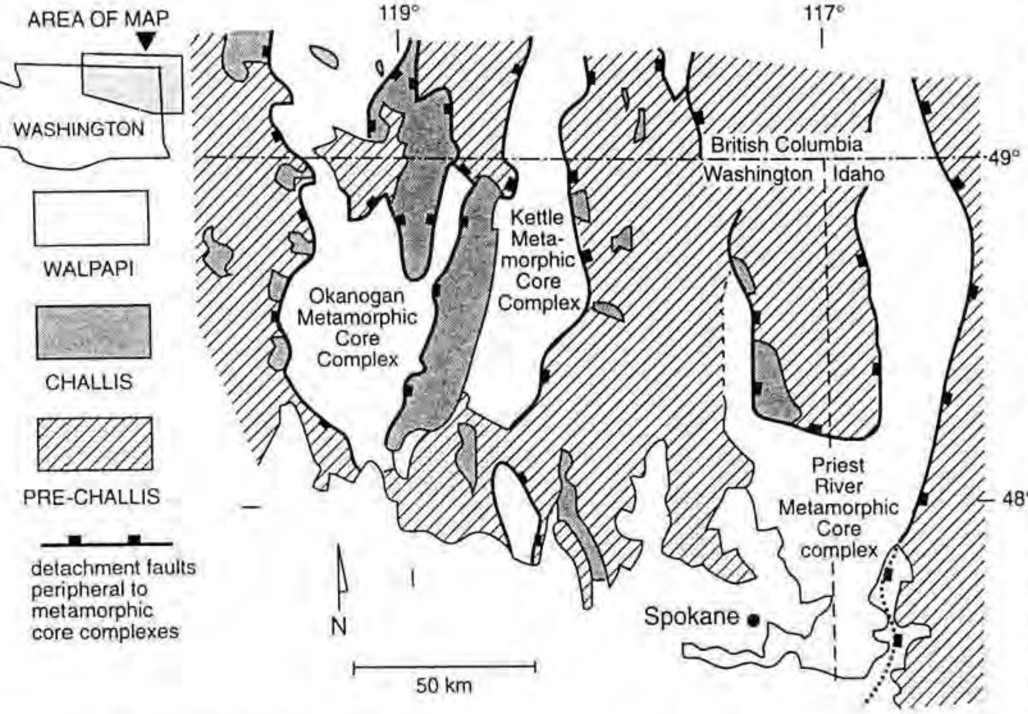


Figure 4. Tectonic map of northeastern Washington and adjacent map. Note that the detachment faults (blocks on the upper plate) bordering the metamorphic core complexes cut the Challis sequence. Source of data: Stoffel and others (1991).

Wheeler and Mallory (1970) referred to the deformational event at the base of this sequence as the Cascade orogeny. This name emphasizes that the topographic relief of the present Cascade Range is post-Walpapi.

I informally refer to these post-2 Ma rocks as the High Cascade sequence, as is the custom in Oregon (Peck and others, 1964; Armstrong, 1978). This is Armentrout's (1987) Sequence V.

REGIONAL CHALLIS SEQUENCES

Introduction

The Challis sequence has at least eight regional UBS. Each of the eight has a characteristic lithologic composition, but some lithologies are so similar that if one sequence rests upon another with little angular discordance, the intervening unconformity is difficult to recognize. This paper uses the principles of sequence stratigraphy (in conjunction with available paleontologic and radiometric data) for regional correlation. Foremost among these principles are numbers 1, 4, 5, 8, 9, and 10 of Table 1. Correlations between areas are particularly strong where two or more sequences occur in the same stratigraphic order, even if some intervening sequences may be missing.

The growing number of fission-track and radiometric dates are most useful. Because the temporal difference between some sequences is slight compared to the precision of some dates and because some dates have been reset, published radiometric and fission-track dates do not always permit the recognition of sequences. Where my assignment of rocks to a certain sequence is in conflict with published ages, I explain the discrepancy; otherwise, I incorporate published dates without much comment.

According to Principle 9 of Table 1, each sequence represents a certain interval of time; thus, if a sequence has been dated in one area, I extend its bounding dates to other areas in which undated representatives exist. In the Wheeler diagrams, this is done by assigning ages to the regional unconformities. The hiatuses represented by some regional unconformities may be 1 Ma or less. The ages shown for most of the unconformities on the Wheeler diagrams are probably ± 1 Ma.

In the Wheeler diagrams unconformities are represented by the wavy lines. If a previous author did not recognize an unconformity, but one is evident in the distribution of units on a map, I show it as a wavy line. Otherwise, the contact between units is shown as conformable (the conventional straight line). Because some of the contacts represented by straight lines may be unconformities, unconformities may be under-represented in the diagrams.

The maximum age of the Challis sequence east of Interstate Highway 5 is not well established. Omitting fission-track ages from detrital zircons, the oldest known and seemingly reliable dates presently available are a fission-track age from zircon of 52.7 ± 2.5 Ma from a rhyolite in the Chuckanut Formation of northwestern Washington (Whetten and others, 1988) and a fission-track age of 54 ± 2.1 Ma for zircon in the Silver Pass volcanic rocks of central Washington (Frizzell and others, 1984). I argue below that the Chuckanut Formation is part of the same basal sequence as the Swauk Formation and that the Swauk Formation unconformably underlies the Silver Pass volcanic rocks. For convenience, I adopt a younger limit of 53 ± 1 Ma for the basal sequence and infer that its oldest possible age is 55 ± 1 Ma.

Northeastern Washington

Figure 4 shows that in northeastern Washington and adjacent British Columbia, areas of Challis rocks are commonly bounded by faults, including the low-angle detachment faults that border the metamorphic core complexes. Pearson and Obradovich (1977) were the first to suggest

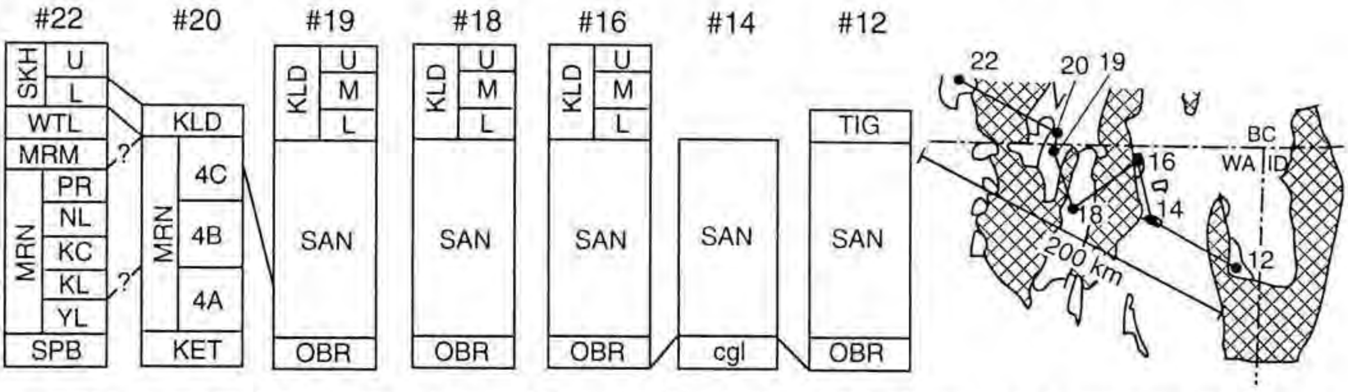


Figure 5. A portion of the correlation diagram of Pearson and Obradovich (1977) for Eocene rocks in northeastern Washington and adjacent British Columbia. Sources of data: Muessig (1967) for column 18; Church (1973) for column 22; and Pearson and Obradovich (1977) for all other columns. The location for each column is shown on the location map, which is compiled from Figures 1 and 4. Abbreviations for formations are the same as those listed in the caption of Figure 6 with the following exceptions: cgl, conglomerate without volcanic clasts, probably in part correlative with the O'Brien Creek Formation; members of the Marron Formation are KC (Kearns Creek), KL (Kitley Lake), NL (Nimpit Lake), PR (Park Rill), WL (White Lake), and YL (Yellow Lake); 4A, 4B, and 4C are divisions of the Marron Formation; L, M, and U are the lower, middle, and upper members of the Klondike Mountain Formation.

that the Challis stratigraphic units and their facies are more extensive than the grabens in which they are preserved (Fig. 5). Pearson and Obradovich (1977), therefore, suppressed many local lithostratigraphic names.

Pearson and Obradovich did not present their interpretations in terms of sequence stratigraphy, but this is easy to do (Fig. 6A) because of the unconformities described by Muessig (1967) in the Republic district and by Church (1973) in the White Lake basin of British Columbia. The descriptions of Pearson and Obradovich (1977) show that the formational names with precedence (from the base upward) are O'Brien Creek (tuffaceous arkose and shale), Marron (alkalic volcanic rocks), Sanpoil (rhyodacitic volcanic rocks), Klondike Mountain (quartz latitic to rhyodacitic volcaniclastic and volcanic rocks), and Tiger (arkose).

For simplicity, some of the columns in the original figure of Pearson and Obradovich are omitted from Figures 5 and 6A. A column could have been added to Figure 6A for the Okanogan Valley in the United States, which Pearson and Obradovich (1977) noted is stratigraphically similar to the Republic area. Instead, I have added (as column 34 in Fig. 6A) the succession along the Chewack–Pasayten fault described by White (1986); this eastern bounding fault of the Methow graben offsets the O'Brien Creek and Sanpoil units but is overlain by the unnamed basaltic to andesitic unit (White, 1986).

Cle Elum Area

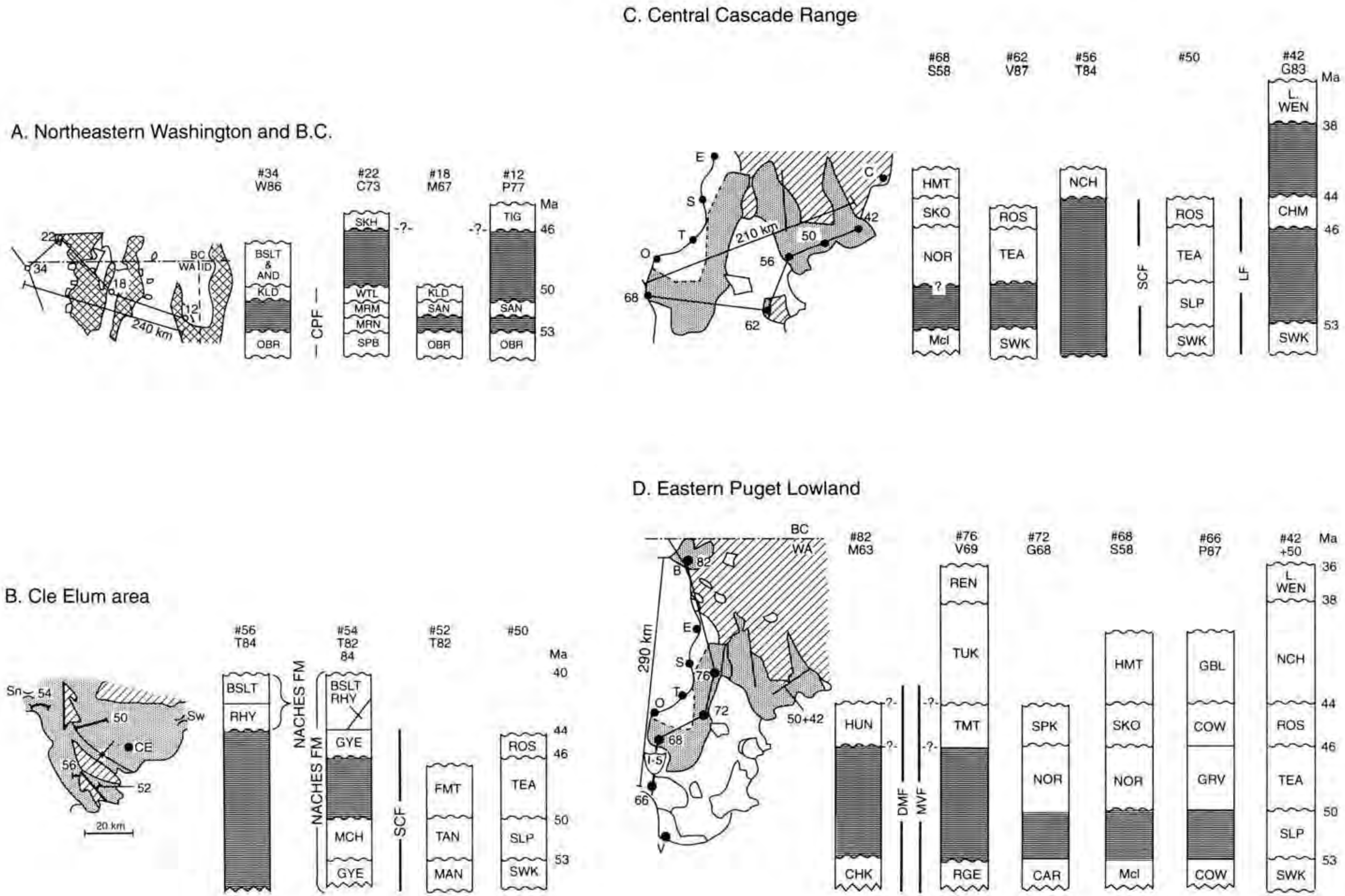
In the central Cascade Range, the Challis section is best documented in the vicinity of Cle Elum (Fig. 7). Here the oldest unit is the Swauk Formation, which is composed of at least 1.3 km of arkosic sandstone, conglomerate, and shale (Foster, 1960). Figure 8 lists the available K-Ar and fission-track ages of the units.

The type locality of the dacitic to andesitic Silver Pass volcanic rocks is about 24 km northwest of Cle Elum (Foster, 1960). Because the Silver Pass volcanic rocks overlie both the Swauk Formation and pre-Tertiary schists (Fig. 7), their base must be an unconformity, as recognized by Foster (1960). Including interbedded volcaniclastic and arkosic strata, the Silver Pass Formation attains a maximum thickness of 1.8 km northwest of Cle Elum, but it pinches out eastward beneath the Teanaway Formation (Foster, 1960, pl. 1; Tabor and others, 1984, fig. 3). Although not shown by Tabor and others (1982a), a belt of the Silver Pass reappears beneath the Teanaway northeast of Cle Elum (J. Margolis, Univ. of Oregon, oral commun., 1991).

Because arkosic interbeds occur in the Silver Pass Formation, whereas tuffs and volcaniclastic rocks occur in the Swauk Formation, Tabor and others (1982a, 1984) inferred that the two interfinger and that the Silver Pass is a member of the Swauk Formation. Due to the unconformity between the Silver Pass and the Swauk, such a correlation is untenable (Table 1). Accordingly, fission-track dates (54.1 ± 2.1 and 52.2 ± 1.9 Ma) at the type locality of the Silver Pass (Frizzell and others, 1984) date not the Swauk, as assumed by Tabor and others (1982a, 1984), but the Silver Pass.

Tabor and others (1982a) reported four fission-track ages from Silver Pass-like volcanic rocks within the Swauk Formation. Of these, the one with the oldest maximum age (49.1 ± 5.2 Ma) is from the previously unrecognized eastern portion of Silver Pass rocks (J. Margolis, Univ. of Oregon, oral commun., 1991). The three remaining dates from volcanic rocks in the Swauk (50.5 ± 1.2, 48.6 ± 2.3, 43.6 ± 1.1 Ma) are younger than the Silver Pass (Fig. 8) and, therefore, rather than reliably dating the Swauk Formation as supposed (Tabor and others, 1982a, 1984), must have been reset.

The basaltic to andesitic Teanaway Formation unconformably overlies the Swauk Formation (Foster, 1960; Tabor and others, 1982a, 1984). Foster could not determine whether the Teanaway is unconformable upon the Silver


A. Northeastern Washington and B.C.
#34 W86
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Ma
BSLT & AND
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B. Cle Elum area
#56 T84
#54 T82 84
#52 T82
#50
BSLT
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NACHES FM
GYE
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TAN
MAN
ROS
TEA
SLP
SWK
SCF
CE
Sn
Sw
20 km
C. Central Cascade Range
#68 S58
#62 V87
#56 T84
#42 G83
HMT
SKO
NOR
McI
NCH
L. WEN
CHM
LF
210 km
D. Eastern Puget Lowland
#82 M63
#76 V69
#72 G68
#66 P87
#42 +50
HUN
CHK
REN
TUK
TMT
RGE
SPK
CAR
GBL
COW
GRV
DMF
MVF
290 km
I-5
50+42

Figure 6. *(facing page)* Area–time diagrams of the Challis sequence. The location map for Figure 6A is compiled from Figures 1 and 4. Portions of the location maps of Figures 6B, 6C, and 6D are compiled from Figures 2, 3, or 4. The names of formations are those used by Walsh and others (1987), Stoffel and others (1991), or the U.S. Geological Survey (Vine, 1969; Easterbrook, 1976; Tabor and others, 1982a). Exceptions are the Guye formation (Foster, 1960) and the formations in British Columbia (Church, 1973). See this text and Figure 8 for an explanation of column #50.

Wavy lines between formations indicate unconformities described in the literature or in the text; straight lines indicate that no local evidence for an unconformity is yet known; saw teeth indicate that the bottom of the formation is not exposed. The location for each column can be determined by matching its number on the accompanying location map.

Sources of data:

C73	Church, 1973
G68	Gard, 1968
G83	Gresens, 1983
M63	Miller and Misch, 1963
M67	Muessig, 1967
P77	Pearson and Obradovich, 1977
P87	Phillips, 1987
S58	Snavely and others, 1958
T82	Tabor and others, 1982a
T84	Tabor and others, 1984
V69	Vine, 1969
V87	Vance and others, 1987
W86	White, 1986

Abbreviations for formations:

CAR	Carbonado Formation
CHK	Chuckanut Formation
CHM	Chumstick Formation
COW	Cowlitz Formation
FMT	basalt of Frost Mountain
GBL	Goble Volcanics
GRV	Grays River volcanic rocks of the Cowlitz Formation
GYE	Guye Formation
HMT	Hatchet Mountain Formation
HUN	Huntingdon Formation
KLD	Klondike Mountain Formation
MAN	Manastash Formation
MCH	Mount Catherine Rhyolite
McI	McIntosh Formation
MRM	Marama Formation
MRN	Marron Formation
NCH	Naches Formation
NOR	Northcraft Formation
OBR	O'Brien Creek Formation
REN	Renton Formation
RGE	Raging River Formation
ROS	Roslyn Formation
SAN	Sanpoil Volcanics
SKH	Skaha Formation
SKO	Skookumchuck Formation
SLP	Silver Pass volcanic rocks
SPB	Springbrook Formation
SPK	Spiketon Formation
SWK	Swauk Formation
TAN	Taneum Formation
TEA	Teanaway Formation
TIG	Tiger Formation
TMT	Tiger Mountain Formation
TUK	Tukwila Formation
WTL	White Lake Formation
WEN	lower member of the Wenatchee Formation

Abbreviations for faults:

CPF	Chewack–Pasayten
DMF	Devils Mountain
LF	Leavenworth
MVF	Mount Vernon
SCF	Straight Creek

Abbreviations for rock types

AND	andesite
BSLT	basalt
RHY	rhyolite

Geographic abbreviations:

B	Bellingham
BC	British Columbia
C	Chelan
CE	Cle Elum
E	Everett
ID	Idaho
I-5	Interstate Highway 5
O	Olympia
S	Seattle
Sn	Snoqualmie Pass
Sw	Swauk Pass
T	Tacoma
V	Vancouver
WA	Washington

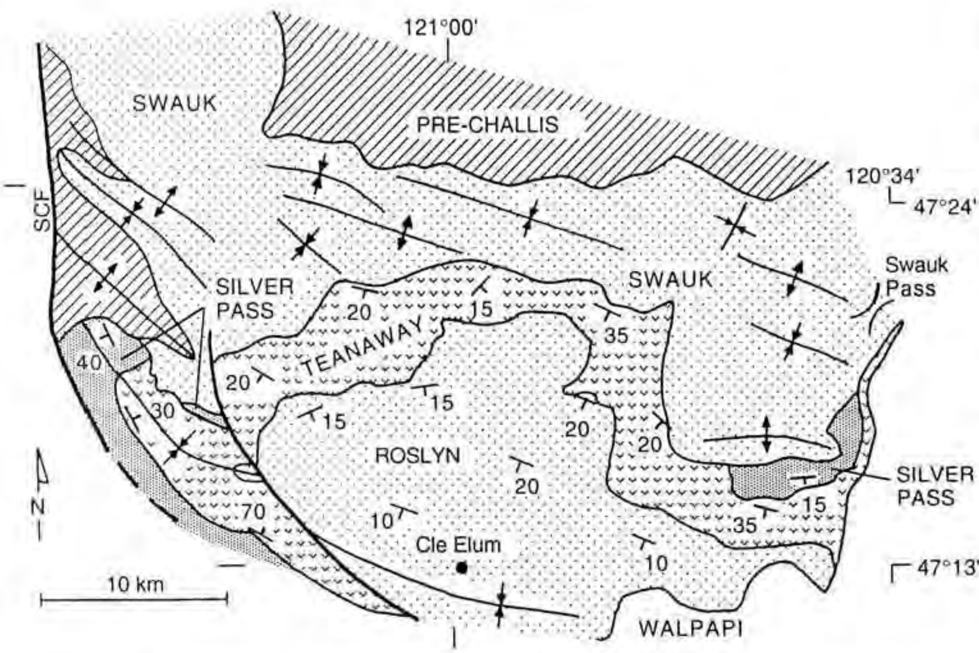


Figure 7. Distribution of Challis sequences in the Cle Elum area. Sources of data: Foster (1960), Tabor and others (1982a), Frizzell and others (1984), and Jacob Margolis (Univ. of Oregon, oral commun., 1991). Figure 8 is the explanation for this figure. SCF, Straight Creek fault.

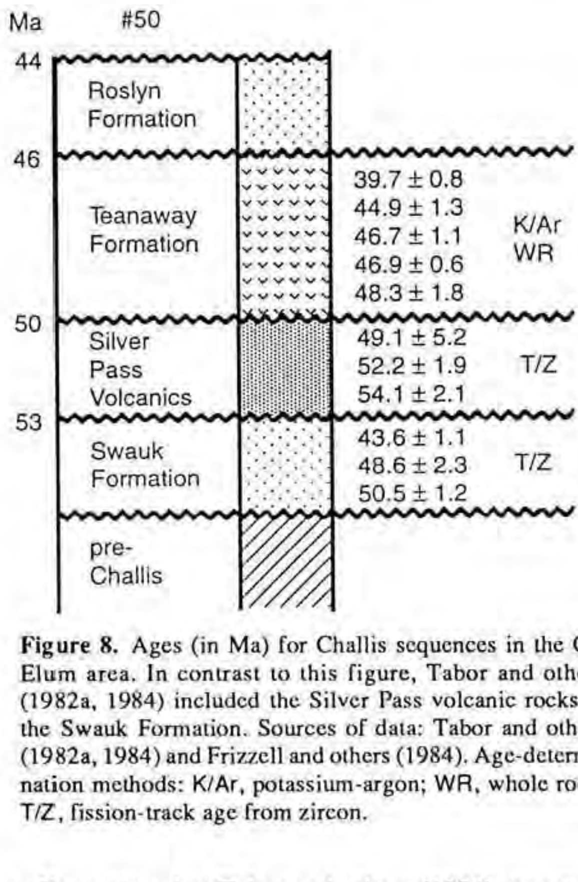


Figure 8. Ages (in Ma) for Challis sequences in the Cle Elum area. In contrast to this figure, Tabor and others (1982a, 1984) included the Silver Pass volcanic rocks in the Swauk Formation. Sources of data: Tabor and others (1982a, 1984) and Frizzell and others (1984). Age-determination methods: K/Ar, potassium-argon; WR, whole rock; T/Z, fission-track age from zircon.

Pass Formation, but Tabor and others (1984) reported that the Swauk and Silver Pass are more tightly folded than the Teanaway. Figure 8 shows that the K-Ar ages for the Teanaway Formation are younger than fission-track ages from the Silver Pass Formation. Furthermore, the Teanaway overlies both the Swauk and the Silver Pass Formations (Fig. 7). In the 40 km across Figure 7, the Teanaway thins eastward from a maximum of 2.5 km to less than 10 m (Tabor and others, 1984).

The Roslyn Formation near Cle Elum consists of poorly exposed nonmarine arkosic rocks. Bressler (1957) divided the formation into a lower conglomerate-bearing unit (700 m), a middle unit of mostly medium-grained sandstone (760 m), and an upper unit (475 m) of sandstone and siltstone containing eight (previously mined) coal seams. Tabor and others (1982a) thought that the Roslyn was conformable upon the Teanaway Formation. However, Bressler (1957) recognized as much as 60 m of red, fine-grained clastic strata locally along the basal contact and thought that they represented regolithic detritus from the weathering of the underlying Teanaway basalts. Thus, he concluded that the lower Roslyn lies disconformably upon the Teanaway.

Figure 6B compares the stratigraphy of the Cle Elum (column 50) and nearby areas. Cle Elum is near the axis of a northwesterly striking syncline of Challis rocks (Fig. 7). Column 52 of Figure 6B shows that in the syncline to the southwest (in the Manastash River block of Tabor and others, 1984) the stratigraphic order and gross lithologies of the Manastash Formation, Taneum Formation, and basalt of Frost Mountain are similar to those of the Swauk, Silver Pass, and Teanaway, respectively (Tabor and others, 1982a, 1984; Frizzell and others, 1984). Furthermore, one of the two fission-track ages on zircon from the Taneum (51.8 ± 1.0, 46.2 ± 1.1 Ma) is similar to ages from the Silver Pass.

The Naches Formation is a thick unit of interbedded volcanic rocks and feldspathic sandstone with a long history of confused definition (Tabor and others, 1984). As shown by Frizzell and others (1984) and Tabor and others (1984), and hence by Walsh and others (1987), the Naches (column 54 of Fig. 6B) consists of all Challis rocks west of the Straight Creek fault (the Cabin Creek block of Tabor and others, 1984). Hence, the stratigraphic discontinuity across the fault seems impressive.

However, Foster (1960) showed that west of the fault at Snoqualmie Pass (column 54 of Fig. 6B) the lowest Challis unit, the chert-bearing arkose of the Guye Formation, is unconformably overlain by the Mount Catherine Rhyolite. He concluded that the Swauk and Guye Formations and the Silver Pass and Mount Catherine volcanic successions are correlative. Tabor and others (1984) also entertained the idea that Manastash and Guye are equivalent. Thus I equate the Guye with the Swauk and Manastash, and the Mount Catherine with the Silver Pass and Taneum. The names Swauk and Taneum have precedence (cf. Tabor and others, 1982a).

On the basis of chert-bearing arkoses within the Mount Catherine Rhyolite, Tabor and others (1984) concluded that the Guye and Mount Catherine are interbedded. Frizzell and others (1984) and Tabor and others (1984) also mapped an interval of arkose above the Mount Catherine Rhyolite. Unfortunately they extended the name Guye to this upper arkose, whereas Guye as defined by Foster (1960) occurs unconformably below the rhyolite. The likely identity of this upper arkose is Roslyn (without mafic volcanic rocks of the Teanaway/Frost Mountain between it and the Mount Catherine Rhyolite).

In column 56 of Figure 6B, rhyolites overlie pre-Tertiary rocks (Stout, 1964; Goetsch, 1978; Frizzell and others, 1984; Tabor and others, 1984), and basalts overlie the rhyolites. Four fission-track ages ranging from 39.9 ± 1.6 to 44.6 ± 3.2 Ma in the rhyolite and two K-Ar ages (40 ± 0.3 Ma and 43.2 ± 3.1 Ma) for the overlying basalt (Tabor and others, 1984) imply that these rocks are younger than Silver Pass/Taneum. Evidently, these units are yet another sequence that locally rests on pre-Tertiary rocks. Thus, the definition of Naches Formation in column 56 differs from the definition in column 54, which spans at least three regional unconformities. Obviously, the name Naches Formation needs to be modified or abandoned.

Chiwaukum Graben

The Chiwaukum graben is the swath of Challis rocks extending northwesterly from Wenatchee in Figure 2. The stratigraphy and age of strata in the graben (column 42 of Fig. 6C) require some discussion. The Chumstick Formation is a thick section of sandstone, conglomerate, and shale restricted to (and the predominant formation in) the graben (Gresens and others, 1981; Gresens, 1983). Geologic estimates of the thickness of the formation range from 4 km (Willis, 1953) to 9 km (Gresens and Stewart, 1981); on the basis of gravity measurements north of 47°30′, Silling (1979) concluded that the most reasonable thickness is 2 km. Conglomerates in the Chumstick typically have a few to as much as 90 percent felsic volcanic clasts (Laravie, 1976; Whetten and Laravie, 1976; Whetten and Waitt, 1978; Gresens, 1983). This implies uplift and erosion of the Taneum (Fig. 6B) or the Marron, Sanpoil, and Klondike Mountain sequences (Fig. 6A), none of which occur in the Chiwaukum graben.

Ash-fall tuffs less than 10 m thick are minor units in the Chumstick Formation (Gresens, 1983; McClincy, 1986). The range of nine fission-track ages (48.8 ± 7.2 to 42.5 ± 5.1 Ma) from these tuffs (Tabor and others, 1987) is similar to that for four K-Ar ages (47.3 ± 1.8 to 43.5 ± 1.6 Ma) reported by Gresens (1983) and Margolis (1989) for a felsic intrusion and a felsic volcanic complex in the Chumstick Formation south of Wenatchee. These ages suggest that the Chumstick Formation is correlative with the Roslyn Formation of the Cle Elum area, as Tabor and others (1982a) and Gresens (1983) supposed (Fig. 6C).

South of Wenatchee, arkosic rocks host three epithermal gold-silver deposits (Patton and Cheney, 1971; Margolis, 1989; Cameron and others, 1992). These arkosic rocks occur in a faulted anticline; they are stratigraphically below the dated volcanic complex and a distinctive conglomerate of the Chumstick Formation that contains cobbles of vein quartz, felsic volcanic rocks, and plutonic rocks (Gresens, 1983).

The age and stratigraphic identity of the mineralized rocks in the core of the anticline are in dispute. Gresens (1983) provided a list of mineralogical, lithologic, and structural criteria to show that these rocks are Swauk rather than Chumstick, but neither Tabor and others (1982a) nor Margolis (1987, 1989) were convinced. Ott and others (1986) obtained a K-Ar date of 50.9 ± 3.5 Ma for an andesite, which is either a dike or volcanic rock (Cameron and others, 1992), in the mineralized arkoses. Ott and others (1986), therefore, inferred that the arkoses are Swauk. I also interpret this age to be pre-Chumstick and, considering the reset ages of the Swauk Formation below the Silver Pass volcanic rocks, possibly a reset Swauk age. Thus, in column 42 of Figure 6C these mineralized rocks are shown as Swauk Formation. I regard the distinctive cobble conglomerate as the basal conglomerate of the Chumstick Formation.

Near Wenatchee, the 300-m-thick Wenatchee Formation is unconformable upon the Chumstick Formation, the pre-Chumstick mineralized arkosic rocks (Swauk), and pre-Tertiary metamorphic rocks (Gresens, 1983). Quartz-rich sandstones and variegated shales also permit the Wenatchee Formation to be distinguished from the Chumstick Formation (Tabor and others, 1982a). Hauptman (1983) showed that the Wenatchee Formation consists of a basal arkosic fluvial interval (with an eastern provenance) and an overlying lacustrine unit of dark mudstone and siltstone with interbedded sandstones. Three fission-track ages on zircons from tuffaceous beds range from 39.8 ± 9.0 to 33.4 ± 1.4 Ma, whereas a fourth is 49.1 ± 2.3 Ma (Tabor and others, 1982a). The Wenatchee and the Chumstick formations are cut by dikes and sills of hornblende andesite (Gresens, 1983), for which there are five K-Ar dates ranging from 35.1 ± 3.2 to 25.1 ± 0.3 Ma (Tabor and others, 1982a, table 1). Gresens (1983) also recognized an upper member of the Wenatchee Formation. This is 15 m of predominantly conglomerate composed mostly of clasts of felsic volcanic rocks, including pumice. This member is <5 degrees discordant with the lower member of quartz-rich sandstone and variegated shale.

In the Wenatchee–Cle Elum area, three other small patches of tuffaceous rocks occur above Challis rocks and below the CRBG. These localities at Malaga, Lion Rock, and Swauk Prairie are indicated by arrows in Figure 3. Tabor and others (1982a) labeled these Wenatchee Formation but were reluctant to include them with the shaly and quartz-rich sandstones of the lower member of the Wenatchee Formation; so was Gresens (Tabor and others, 1982a). The Malaga locality has a Miocene flora (Gresens, 1983). Fission-track ages from zircons from the Lion Rock and Swauk Prairie localities are 34.4 ± 2.3 Ma and 32.8 ± 0.6 Ma, respectively (Tabor and others, 1982a). On the basis of their tuffaceous nature, stratigraphic position, and ages, I assign these rocks to the Kittitas sequence and suggest that the upper member of the Wenatchee Formation might also be Kittitas.

Columbia Plateau

Deep wells (Campbell, 1989) and geophysical surveys (Glover, 1985; Catchings and Mooney, 1988) show that the Challis and Kittitas rocks exist in the subsurface in central Washington (Fig. 2). This is important because the Kittitas sequence is commonly thought to be restricted to the Cascade Range, and the Challis sequence to the Roslyn basin and the Chiwaukum graben.

Central Cascade Range

Figure 6C shows the Challis sequence stratigraphy across 210 km of the Cascade Range. Because Armentrout (1987) used the Centralia area in defining the sequences of southwestern Washington and adjacent Oregon, column 68 ties the sequences of central Washington into Armentrout's previously defined Sequence II (Fig. 1). Vance and others (1987) recognized the existence of the four Cenozoic sequences in the Mount Rainier–Rimrock Lake area. In their review of the Tertiary rocks of the central and southern Cascade Range, Swanson and others (1989) subdivided the Tertiary rocks into the same four successions described here as sequences, and specifically showed these as unconformity-bounded sequences in the Rimrock Lake inlier (1989, fig. 25); however, they (1989, fig. 4) did not recognize these as regional or interregional sequences.

Because the Hatchet Mountain Formation of Roberts (1958) was not described by Snavely and others (1958) in the Centralia area, an explanation of column 68 is necessary. Schasse's compilation of the Centralia area (1987, table 2) includes K-Ar ages of 38.3 ± 1.9 and 38.8 ± 1.9 Ma for the Northcraft Formation. These ages, if valid, would preclude the correlations shown in Figure 6C of the Northcraft with the Teanaway and of the arkosic, coal-bearing Skookumchuck Formation with the similar Roslyn Formation. Significantly, these two K-Ar ages are not from the type area of the Northcraft Formation northeast of Centralia but from the Newaukum and Tilton River areas southeast of Centralia. I suggest that the two sampled localities are most likely the northern edge of the basal basaltic portion of the Hatchet Mountain Formation described by Roberts (1958) about 20 km to the south.

Column 62 shows that the Challis sequence is unconformable on the pre-Tertiary rocks of the western part of the Rimrock Lake inlier; the names shown in column 62 are those inferred by Vance and others (1967). The rhyolitic Spencer Creek welded tuff (dated 41.8 ± 3.8 Ma by Vance and others (1987) using fission tracks in zircon) unconformably overlies the southeastern part of the inlier (Swanson and others, 1989); this unconformity is, therefore, roughly equivalent to the one at the base of the Naches Formation shown in column 56 of Figure 6B. The outcrop area of the Spencer Creek tuff is too small to show in Figure 2. Figure 2 does show that the Kittitas and Walpapi sequences unconformably overlie parts of the inlier. A part of the High Cascade sequence does as well, but is too small to show.

The Challis sequence is not well exposed in the southern Cascade Range due to the extensive cover of Kittitas and younger sequences. Magnetotelluric surveys suggest the Kittitas rocks are commonly about 5 km thick and that a conductive zone 5 to 7 km thick beneath these volcanic rocks most likely is composed of Challis rocks (Stanley and others, 1990).

Puget Lowland

Figure 6D illustrates the Challis sequences for 290 km on the eastern side of the Puget Lowland and compares them with the Cle Elum area. Correlations are based on available radiometric ages, the fission-track ages of detrital zircons reported by Brandon and Vance (1992), and the principles of sequence stratigraphy (especially numbers 8, 9, and 10 of Table 1). According to Frizzell and others (1984), the Raging River Formation is the temporal equivalent of the Swauk and Manastash Formations. The lower and upper arkosic members of the Cowlitz Formation (column 66) correlate with the McIntosh and Skookumchuck Formations of column 64 (Phillips, 1987).

The two arkosic members of the Cowlitz Formation are separated by the Grays River basaltic rocks. A whole-rock K-Ar age of 37.3 ± 2.2 Ma for the Grays River volcanics (Phillips, 1987) would seem to preclude the correlation in Figure 6D. However, as the matrix of the Grays River is devitrified glass containing chlorite, zeolites, and calcite (Phillips, 1987), and the single dated sample is younger than four of the eight dates for the Goble Volcanics, which regionally overlie the Grays River volcanics, 37.3 ± 2.2 Ma may not be a reliable age. Gard (1968, figs. 7 and 8) demonstrated the correlation of the units in the Carbonado area (column 72) with those of the Centralia (column 68) and intervening areas. Columns 66 to 76 also are consistent with Armentrout's (1987) interpretations.

The ages of the Goble, Hatchet Mountain, and Tukwila formations are compatible with radiometric and fission-track ages discussed by Armentrout (1987), given by Schasse (1987) for the Northcraft Formation (as noted above), and reported by Turner and others (1983), respectively. Flows of the Goble Volcanics are traceable into the Hatchet Mountain Formation (Phillips, 1987). In contrast to Figure 6D, Phillips (1987, table 4) listed eight K-Ar ages ranging from 32.2 ± 0.3 to 45.0 ± 1.4 Ma for the Goble Volcanics. Because these rocks contain chlorite and zeolites, perhaps some dates have been reset, as implied by

Phillips (1987, p. 41). Although Phillips (1987) characterized the contact between the Goble and the Cowlitz as gradational, he also stated that the Goble overlies both the Cowlitz arkosic rocks and the Grays River volcanic rocks; therefore, this contact is shown as an unconformity in column 66 of Figure 6D. The Tukwila Formation of column 76 has a zircon fission-track age of 41.3 ± 2.3 Ma and a K-Ar age of 42.0 ± 2.4 Ma (Turner and others, 1983). The Tukwila is contiguous with the volcanic rocks of Mount Persis (Tabor and others, 1982b; Frizzell and others, 1984), which yielded a K-Ar age of 41.7 ± 4.3 Ma and a fission-track age on apatite of 47 ± 4 Ma (Tabor and others, 1982b).

The youngest group of fission-track ages from detrital zircons from four localities in the Chuckanut ranges from 55 to 58 Ma (Johnson, 1984a); these and fission-track ages of 49.9 ± 1.2 Ma (Johnson, 1984a) and 52.7 ± 2.5 Ma (Whetten and others, 1988) of zircons in tuffs imply that the Chuckanut is coeval with the Swauk Formation. The exact radiometric or paleontological age of the Huntingdon Formation is unknown. Chuckanut strata are truncated by and occur northeast of the Devils Mountain fault (Cheney, 1987, fig. 2; Whetten and others, 1988). Additionally, the Chuckanut and Huntingdon Formations are bounded on the southwest by the northwesterly striking Mount Vernon fault, which dextrally offsets the Devils Mountain fault by 47 km (Cheney, 1987).

Despite being separated from the Raging River Formation by at least these two major faults (Fig. 6D), the Chuckanut Formation can be traced by a more circuitous route into the lowest Challis arkosic sequence of the southeastern part of the Puget Lowland. Northeast of the Devils Mountain fault, the Chuckanut Formation strikes southeastward in a discontinuous belt toward the Straight Creek fault (Huntting and others, 1961). In the southern end of this belt west of the Straight Creek fault (the central belt of Challis rocks shown at the latitude of Seattle in Fig. 2), Tabor and others (1982b) mapped the arkosic rocks as Swauk Formation (overlain by Silver Pass volcanic rocks). Tabor and others (1982a) also mapped Swauk Formation east of the Straight Creek fault in the Cle Elum area. The Chuckanut and Swauk formations also are lithologically similar (Johnson, 1984b). Figure 6C correlates the Swauk Formation east of the Straight Creek fault (column 50) with the McIntosh Formation (column 64) in the southeastern part of the Puget Lowland. This correlation is made on the basis of the lithologic similarity of the Swauk and McIntosh Formations, their stratigraphic position beneath basaltic units (Teanaway and Northcraft Formations, respectively), and the revised age of the Northcraft Formation.

DISCUSSION

Thermal Maturity of Sequences

Thermal maturity can be evaluated by examining the ranks of coals within the Challis and Kittitas sequences. Except where contact metamorphosed, Challis coals range from subbituminous to medium volatile bituminous (Beikman and others, 1961; Walsh and Phillips, 1983; Walsh and Lingley, 1991). In contrast, lignite in beds more than 10 ft thick occurs in the Toutle Formation at Cedar Creek at 122°46'W, 46°25'N (Roberts, 1958; Beikman and others, 1961). The Toutle Formation is part of Armentrout's Sequence III, the Kittitas Sequence. Thus, the maturity of coals in western Washington is sequence-dependent.

Thermal Maturity in the Challis Sequence

Until now, no attempt has been made to determine whether thermal maturity of coals decreases upward in the Challis sequence. To do this, it is necessary to recognize that coal has been mined from each of the three arkosic sequences in the Challis rocks of central and western Washington. Coal was mined from the Carbonado and Chuckanut Formations (Beikman and others, 1961; Gard, 1968) of the basal sequence. Mining in the Skookumchuck, Spiketon, and Roslyn Formations (Beikman and others, 1961; Schasse and others, 1984, *in* the reprint of Beikman and others, 1961) was from the middle arkosic sequence. Coal mined from the Puget Group in the Green River area (Beikman and others, 1961; Vine, 1969) probably is from both the middle and the upper sequences (Vine, 1969, fig. 2, pl. 2). The Huntingdon Formation is unconformable upon the Chuckanut Formation near Bellingham (Miller and Misch, 1963). A comparison of the maps of Miller and Misch (1963), Moen (1969, figs. 8 and 9), and Easterbrook (1976) show that the coal mining at Bellingham was in the Huntingdon Formation, as Vonheeder (1975) suggested. The question marks adjacent to column 82 of Figure 6D indicate that I am not sure whether the Huntingdon Formation belongs in the middle or the upper arkosic sequence. Coal mined from the Renton Formation (Beikman and others, 1961; Vine, 1969) is from the upper arkosic sequence.

Walsh and Phillips (1983, figs. 3, 7, and 9) and Walsh and Lingley (1991) showed that the ranks of individual seams in the Chuckanut, Carbonado, and Roslyn Formations increase toward the present crest of the Cascade range; these authors attributed the increase to thermal overprinting by Kittitas-age volcanism and plutonism. However, figure 6 of Walsh and Lingley (1991) hints that the rank of Challis coals may be sequence-dependent; it shows that the lowest ranks of the Chuckanut and Carbonado coals tend to be high volatile C and B bituminous, whereas only the highest values of the stratigraphically higher Challis coals attain these ranks.

Other data exist to test whether thermal maturity decreases upward in the Challis sequence. Beikman and others (1961) tabulated analyses of British Thermal Units per pound (BTU/lb) and the percent fixed carbon of various Challis coals; these data are from mined coals. The data of Beikman and others (1961) obviously favor the thicker, cleaner, and more continuous seams. Unfortunately, the data for vitrinite reflectance of arkosic Challis rocks (Walsh and Lingley, 1991, table 1) are from weathered rock (outcrops), not fresh rock. Another potential limitation of comparing the two sets of data is that the analyses by Beikman and others (1961) and by Walsh and Lingley

Table 2. Statistical summary of data from rocks and coals of the Carbonado Formation, Puget Group, and Renton Formation. The source of data for vitrinite reflectance is Walsh and Lingley (1992, table 1). The entries for percent fixed carbon of coals and British Thermal Units per pound of coal are calculated from Beikman and others (1961, tables 9, 10, 17, and 19). Std. Dev. is one standard deviation.

Parameter	Statistics	Carbonado Formation	Puget Group	Renton Formation
Mean random vitrinite reflectance (percent)	Number	12	12	7
	Mean	1.02	0.56	0.50
	Std. Dev.	0.27	0.13	0.10
BTU/pound x 10^3	Number	9	37	24
	Mean	12.9	11.6	10.4
	Std. Dev.	0.4	1.2	0.9
Fixed carbon (percent)	Number	9	37	24
	Mean	53.1	43.8	40.8
	Std. Dev.	3.5	4.2	3.5

(1991) almost certainly were not performed on adjacent rocks.

Although Beikman and others (1961) and Walsh and Lingley (1991) did report data from other coals and strata, the only lithostratigraphic units with enough data to treat statistically ($N \geq 7$) are the Carbonado Formation, Puget Group, and Renton Formation. Accordingly, only the data for these three units are summarized in Table 2. These three units are in the same general area southeast of Seattle.

The data for the Puget Group are from the section along the Green River, which is between the areas represented by columns 72 and 76 of Figure 6D. Due to the lack of any volcanic interval at least a few tens of meters thick (that might be equivalent to either the Northcraft Formation or the Tukwila Formation), the stratigraphic position of this part of the Puget Group cannot be unambiguously determined lithostratigraphically. The paleoflora of the lower two-thirds of this part of the Puget Group appears to correlate with those of the Raging River and Tiger Mountain formations, and the upper one-third seems to correlate with the Renton Formation (Vine, 1969, fig. 12, pl. 2). K-Ar and fission-track dates (41.2 ± 1.8 to 45.0 ± 2.1 Ma) from partings of volcanic ash in the Green River section (Turner and others, 1983) imply that the Puget Group is younger than the age of Carbonado Formation shown in Figure 6D. A zircon fission-track age of 41.3 ± 2.3 Ma and K-Ar age of 42.0 ± 2.4 Ma from the Tukwila Formation (Turner and others, 1983) that underlies the Renton Formation suggest that the Renton could be younger than the portion of the Puget Group in the Green River section. Based on sequence stratigraphy (Fig. 6D), the Renton Formation is younger than the Carbonado Formation.

Some of the data of Beikman and others (1961) are excluded from Table 2 and Figure 9. Beikman and others (1961) correlated the coals at Niblock with the Renton Formation. However, their data for the Niblock coals are omitted because Tabor and others (1982) showed that the Niblock area is underlain by the Tiger Mountain Formation, not the Renton Formation. Because Vine (1969) suspected that the Kummer coals of the Puget Group in the Green River area correlate with the Renton Formation, the values for Kummer coals in Beikman and others (1961, table 17) are not used for the Puget Group in Table 2, nor are they included in the data for the Renton Formation.

The data of Table 2 are shown graphically in Figure 9. Figure 9A shows that the thermal maturity of the Carbonado Formation is statistically greater (at one standard deviation) than that of the Puget Group and the Renton Formation (which are statistically indistinguishable). Figure 9B shows that vitrinite reflectance and the BTU/lb of the coals provide a better discrimination between the sequences. Given the previously mentioned caveats about the data, the results of Figure 9 must be considered preliminary, but they seem to indicate that in the area southeast of Seattle, thermal maturity of the Challis rocks is sequence-dependent.

Extent of Challis Sequences

Figure 10, which summarizes the columns shown in Figure 6, shows that the Challis rocks consist of eight sequences.

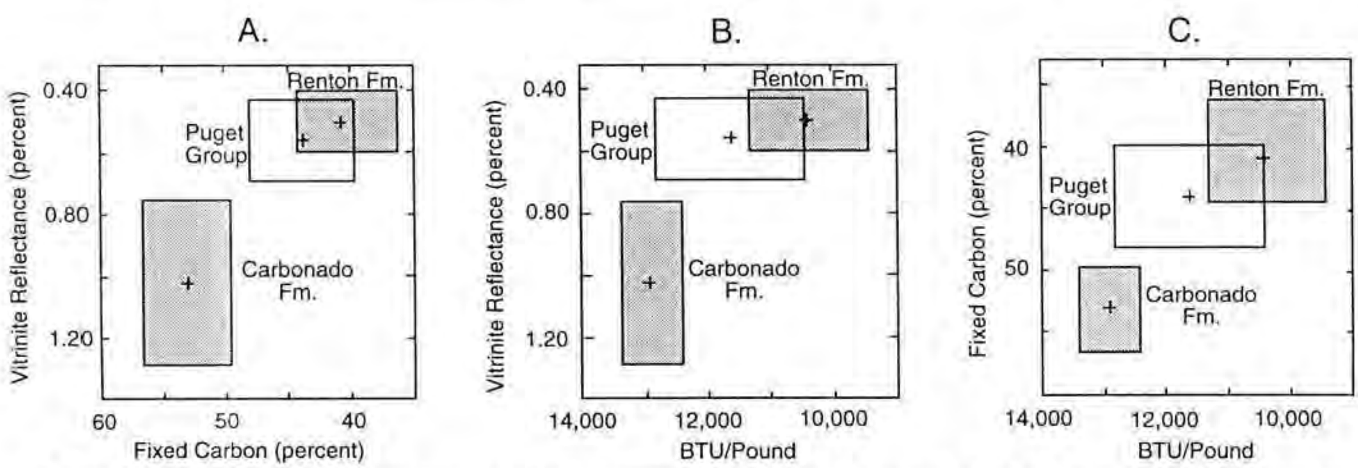


Figure 9. Comparison of some characteristics of Challis formations southeast of Seattle. See Table 2 for data. Crosses are means; boxes show one standard deviation.

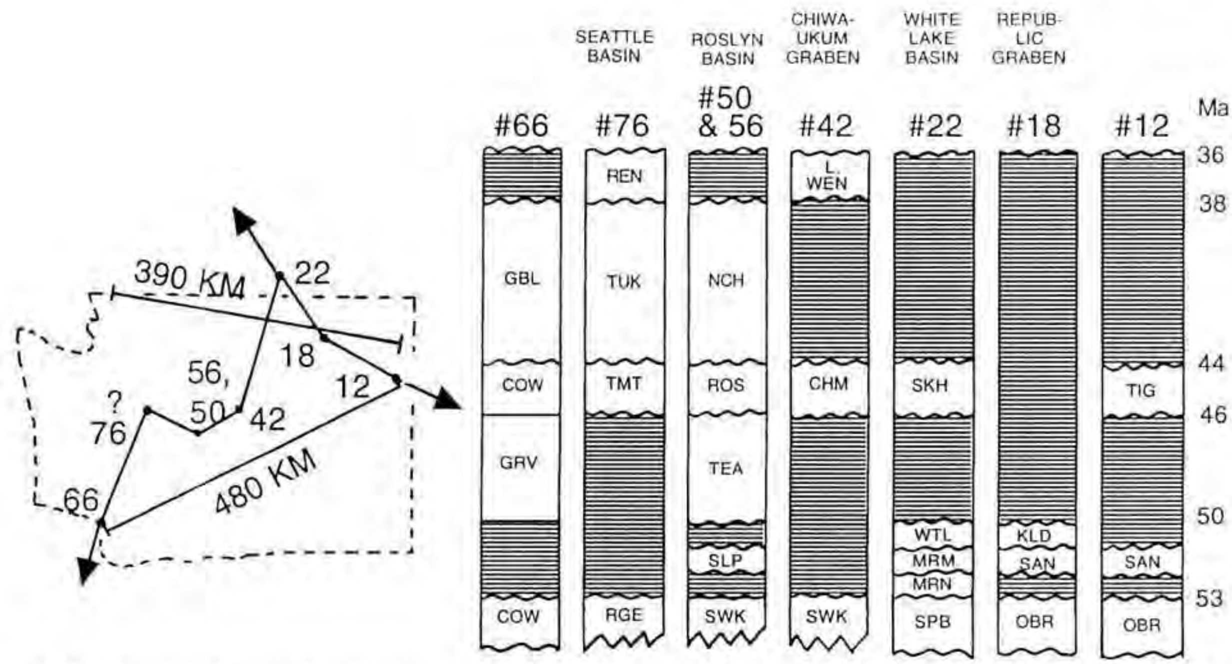


Figure 10. Summary of regional sequences within the Challis sequence in Washington. For sources of data and names of formations, see Figure 6.

The sequence represented by the Naches, Tukwila, and Goble formations in Figure 10 has the longest duration and probably is the least well mapped. If additional sequences exist in the Challis rocks, they are most likely to be discovered in this interval.

The presently known maximum strike length of a sequence in Washington is about 480 km. Similar rocks of similar age in the Princeton, BC, area 70 km west of the White Lake and in the Hat Creek area 200 km northwest of White Lake (Church and others, 1983) suggest that these sequences could be more extensive. If some of the sequences extend as far southwest as Coos Bay, OR (Armentrout, 1987), their strike length is at least 730 km.

The eight Challis sequences clearly are not coextensive. The younger sequences appear to be absent to the northeast. Three silicic volcanic sequences (the alkalic Marron, rhyodacitic Sanpoil, and volcaniclastic and quartz latitic to rhyodacitic Klondike Mountain) occur above the basal clastic sequence in British Columbia, but only two (Sanpoil and Klondike Mountain formations) exist in northeastern Washington, one (Silver Pass volcanic rocks) in central Washington, and, apparently, none in the Puget Lowland.

Figure 6 and the preceding discussions show that the older Challis sequences are cut by a number of major faults (in addition to the detachment faults bounding the metamorphic core complexes). The persistence of these sequences across the faults shows that the geographic extents of the sequences are greater than any strike-slip displacements on the faults.

Challis Basins

Present models stress that in Washington the Cenozoic strata, especially the Challis rocks, are local successions deposited in syntectonically active grabens or subsiding basins (Swanson and others, 1989). Well-known examples of such models are:

- *Chiwaukum graben* (Willis, 1953; Whetten, 1976; Gresens, 1980; Gresens and others, 1981; Tabor and others, 1982a, 1987; Johnson, 1985; Evans and Johnson, 1989; Margolis, 1989)
- *Manastash graben* (Johnson, 1985)
- *Roslyn basin* (Bressler, 1957; Foster, 1960; Tabor and others, 1982a, 1984)
- *Swauk basin* (Evans and Johnson, 1989)
- *Republic graben* (Muessig, 1967; Holder and others, 1990)
- *Chuckanut basin* (Johnson, 1984b, 1985; Evans and Johnson, 1989; Walsh and Lingley, 1991; Heller and others, 1992)
- *Everett, Seattle (Puget), Tacoma, and Chehalis depocenters* (Johnson, 1984b, 1985; Walsh and Lingley, 1991; Heller and others, 1992)
- *Columbia basin* (Reidel and others, 1989a, b; Walsh and Lingley, 1991).

Many of these models were directly (Gresens, 1983; Tabor and others, 1984; Johnson, 1985) or indirectly influenced by Crowell (1974), who deduced syntectonic sedimentary patterns in pull-apart basins modeled after the strike-slip faulting in California. In Washington, such models seemed especially attractive because several high-angle, regional faults are well known (Huntting and others, 1961), and by the 1980s, several of these faults were known to have, or suspected of having, some strike-slip movement.

Figure 10 illustrates that the regional sequences within the Challis are more extensive than the local basins in which they commonly are inferred to have been deposited. The basins must, therefore, be post-depositional (preservational), rather than depositional. The apparently local stratigraphy of individual basins must be due to facies changes and to differential erosion of various sequences beneath various unconformities. Erosion, clearly, was episodic and extensive.

Additional reasons for doubting that the local basins are depositional are:

(a) Eocene transcurrent movements have not yet been documented on any of the inferred basin-bounding faults.

(b) As noted by Johnson (1985), the Eocene basins inferred in Washington are as much as 90 km wide. This is considerably wider (by a factor of 3) than the widest pull-apart basin listed by Aydin and Nur (1982), who described the 30-km-wide Lago de Izabul basin of Guatemala as "huge".

(c) At least in northeastern Washington and adjacent British Columbia, the geographic extent of facies in the Challis strata appears to be larger than the grabens and to be spatially unrelated to bounding faults of individual grabens (Pearson and Obradovich, 1977).

(d) Deformation in some of the basins is so severe that the original depositional shape and extent of the basin can hardly have been retained. Dips of Challis strata are (at least locally) ≥60 degrees in the Republic graben (Muessig, 1967), Chiwaukum graben (Tabor and others, 1982a, 1987), Roslyn basin (Tabor and others, 1982a), the Carbonado area (Gard, 1968), Green River gorge (Vine, 1969), and the Chuckanut basin (Miller and Misch, 1963).

(e) Most of the areas of Challis rocks in northeastern Washington and adjacent British Columbia are bounded on one or both sides by detachment faults marginal to the metamorphic core complexes (Orr and Cheney, 1987; Parrish and others, 1988; Box, 1992; Cheney and others, 1994, this volume). Cumulative extension across these detachment faults is on the order of 80 km or 25–30 percent (Parrish and others, 1988) and locally may be 40–120 percent (Harms and Price, 1992). Therefore, Challis successions now bounded by detachment faults may once have been considerably closer together, or even contiguous (Parrish and others, 1988, pl. 2).

The cause for the great thickness of strata in some areas remains uncertain. Being near the former continental margin, the Challis sequences may be considerably thicker than if they had been in the cratonic interior. Alternatively, more than one Challis sequence may be present in a given area; for example, if an arkosic sequence were deposited on top of an older arkosic sequence, the unconformity might be very difficult to detect, and the stratigraphic section would appear to be unusually thick.

Another possibility is that some stratigraphic sections may have been tectonically repeated. A thrust system is well known along the northeastern edge of the Chiwaukum graben (Patton and Cheney, 1971; Cameron and others, 1992). Other thrusts within the graben might be as yet undetected.

Provenance of Challis Sequences

The provenance of the Challis sequence was not primarily local. Bressler (1957) inferred that the Roslyn Formation (Fig. 3) is derived from the Cretaceous Mount Stuart batholith to the north; however, the batholith has abundant plagioclase and hornblende, whereas the Roslyn Formation has abundant K-feldspar and biotite. Biotite and hornblende K-Ar ages of the crystalline rocks in the core of the northern Cascade Range are as young as 50 to 40 Ma (McGroder, 1990, fig. 7), indicating that these rocks were deeply buried at the time the nearby Chuckanut Formation was deposited. Because the Challis sequences in northeastern Washington are cut by the detachment faults bordering the metamorphic core complexes (Muessig, 1967; Orr and Cheney, 1987; Parrish and others, 1988), they must be older than the core complexes. Cooling ages of biotites in the metamorphic core complexes are 51–42 Ma (Parrish and others, 1988; Harms and Price, 1992), confirming that these rocks were not available for erosion until late in Challis time.

The abundance of chert and volcanic fragments (Johnson, 1984b; Heller and others, 1992) in the basal clastic sequence (represented by the Chuckanut and Swauk formations) shows that much of the detritus consists of supracrustal rocks. Heller and others (1992) concluded that the basal clastic sequence resulted from sources in eastern Washington or east or north of Washington. The sources might have been uplifted by Laramide events or the earliest detachment faults (54–60 Ma) of the metamorphic core complexes described by Parrish and others (1988).

In contrast, the mineralogy of Challis volcanic rocks in the Methow and Republic areas does appear to match the mineralogy of nearby dikes or granitic plutons (White, 1986, and Holder and others, 1990, respectively). A swarm of Teanaway dikes is well known in the Swauk Formation north of the Teanaway Formation (Tabor and others, 1982a).

Conceivably, the Tiger Formation, although cut by the Newport detachment fault, is syndeformational (Harms and Price, 1992). If so, the middle UBS of the Challis (represented by Tiger, Skaha, Chumstick, Roslyn, Cowlitz, Tiger Mountain, and possibly the Puget and Huntingdon successions) may represent erosion of the metamorphic core complexes (Orr and Cheney, 1987). For this to be possible, erosion of the core complexes and deposition of the arkosic rocks must have begun before the last rocks in the core complexes cooled to blocking temperatures about 45 Ma.

Challis Volcanic Centers

Just as many authors have regarded thick sedimentary sections to be depositional basins, so too have some authors regarded thick volcanic successions to be volcanic centers, constructional volcanic hills on an arkosic plain, or the filling of paleovalleys. The synformal nature of the thick section of Northcraft volcanic rocks west of Carbonado (Gard, 1968) shows that some thick sections are merely structurally preserved remnants of formerly more extensive volcanic units. Clearly other criteria (such as the presence of plutons, ring dikes, radial dikes, densely welded tuffs, and extremely coarse volcaniclastic rocks) are better criteria for a volcanic center.

Volume of Columbia River Basalt

The upper bounding unconformity of the Walpapi sequence, as well as unconformities within the sequence, indicate that the volume and rate of extrusion of CRBG are currently underestimated. Tolan and others (1989) revised previous estimates of the areal extent and volume of the CRBG (200,000 km^2 and 325,000–382,000 km^3) downward to 164,000 km^2 and 174,300 ± 31,000 km^3, respectively. These may be reasonable estimates of the preserved extent and volume of CRBG as presently defined, but Hull (1991) suggested that the CRBG is more extensive in southeastern Oregon. Although a volume of about 200,000 km^3 may seem impressive, some other flood basalts and oceanic plateau basalts are at least five times larger (Richards and others, 1989, 1991).

Tolan and others (1989, p. 14) recognized that one of the major problems in constructing isopach maps for the CRBG flows is "defining the degree of erosional stripping around the margins of the units." Nonetheless, they used the present erosional edge of the CRBG in their calculations of the volume of the four magnetostratigraphic units (R_1, N_1, R_2, N_2) of the Grande Ronde Basalt, which together constitute 85 percent of the estimated volume of the CRBG.

A number of unconformities are known in the CRBG above the Grande Ronde Basalt (Tolan and others, 1989, fig. 1; Reidel and others, 1989a, fig. 1). Judging from figures 8 and 9 of Reidel and others (1989b), another major, but unrecognized, unconformity occurs beneath magnetostratigraphic unit N_2 of the Grande Ronde: N_2 truncates portions of R_1, N_1, and R_2. This truncation also can be inferred from the maps of figure 2 of Reidel and others (1989a), which show that N_2 is far more common along the present western and northern margins of the CRBG than R_1, N_1, or R_2. This relation is confirmed by the 1:100,000 maps of Frizzell and others (1984) and Tabor and others (1982a, 1987). Thus, significant amounts and extents of basalt were eroded in pre-N_2 time.

The volume of basalt within the Walpapi sequence is considerably larger than for the narrowly defined CRBG alone. Steens Mountain and related basalts in southeastern Oregon and adjacent Nevada are part of the Walpapi sequence. The 2–3 and 6–10 Ma Chilcotin basalts of south-central British Columbia (Bevier, 1983) could be part of the Walpapi sequence. Inclusion of these units would significantly increase the pre-2 Ma area (and volume) of the Walpapi basalts, especially if the areas between CRBG and the Chilcotin and between the CRBG and Steens Mountain were once largely covered by basalt. For example, the present northwesterly strike length of the Steens Mountain basalt plus CRBG is about twice that of the CRBG; the present northwesterly strike length of the Steens Mountain basalt plus CRBG plus Chilcotin Group is about four times that of the CRBG.

The rate of extrusion of the CRBG is typically shown (Reidel and others, 1989b, fig. 12; Swanson and others, 1989, fig. 16) as tapering off dramatically with time (stratigraphically upward). These rates were calculated using the presently preserved volumes of stratigraphic units. Because erosion proceeds from the top down, the uppermost units of the CRBG have probably been eroded the most. If so, the rate of extrusion did not decline as dramatically as commonly supposed.

Challis/Crescent Relations

Armentrout (1987) considered the basaltic Crescent Formation and associated overlying sedimentary rocks to be a separate sequence, I, below his sequence II. II is the Challis sequence (Fig. 1). However, many of the radiometric dates for the Crescent and related rocks are younger than the oldest Challis rocks. Some of the basalts are as young as 46 Ma (Armentrout, 1987; Babcock and others, 1992). One interpretation of these young ages and of micaceous arkosic rocks associated with the Crescent basalts is that the Challis and Crescent rocks interfinger and, thus, are not separate sequences.

The following suggest that the Crescent and related rocks are not part of the same sequence as the Challis rocks but are a separate terrane:

(a) Although arkosic rocks in the Coast Range rocks and the Challis may be lithologically and temporally similar, to date no Challis rocks above the lowest basaltic unit (Northcraft-equivalent) have been described as overlying the Coast Range rocks. That is, no appreciable section of Challis rocks is known to overlie the Crescent.

(b) Johnson (1984b) noted that a petrologic mismatch exists between the arkosic Challis sandstones and coeval arkosic rocks associated with the Crescent basalts.

(c) The Challis rocks are virtually unmetamorphosed (Brandon and Vance, 1992) except for various zeolites, whereas the Crescent basalts contain prehnite, pumpellyite, and epidote.

(d) East of the Crescent Formation, basaltic debris is lacking in Challis sedimentary rocks but is present in the overlying rocks (Armentrout, 1987; Yount and Gower, 1991) of the Kittitas sequence.

(e) Minor element profiles of the Crescent basalts appear to be more MORB-like than those of Challis volcanic rocks (Phillips and others, 1989).

To prove that the Crescent rocks are a separate terrane with respect to the Challis, a fault must be demonstrated between the two. Unfortunately, virtually all contacts of the Crescent rocks with Challis and older rocks are covered. An exception is southern Vancouver Island where Crescent-equivalent rocks are juxtaposed against the Leech River schist by the Leech River fault. The Leech River schist has a metamorphic age of 42 Ma, and the fault is overlain by rocks that have a fossil content equivalent to 24–30 Ma (Brandon and Vance, 1992). If this is the terrane-bounding fault, as seems likely (Cheney, 1987; Brandon and Vance, 1992), docking of the Crescent rocks must post-date most Challis rocks.

The most likely place to find a faulted contact of the Crescent rocks with the Challis sequence is between Centralia and the Oregon border at about 123°W. The steep northerly trending gravity gradient here is the obvious place to draw the contact (Finn, 1990). This is an area of poor outcrop where both the Challis and Coast Range rocks seem to contain arkosic rocks. Because no fault has yet been observed (Walsh and others, 1987), the fault is shown as a dashed line in Figure 11.

Regional Geology of Washington

Figure 11 shows the extent of the four regional Cenozoic sequences beyond the areas previously discussed. This is the first map of Washington to show both superterranes and major sequences as regional units. These are natural regional units. In contrast, conventional regional maps, such as the one by Schuster (1992), show lithologies divided into manmade chronostratigraphic units (such as Cambrian, Mesozoic, Miocene, etc.). Most chronostratigraphic divisions have natural boundaries (unconformities) in the areas in which they were originally defined but have evolved into manmade units of convenience for correlation. Thus, the boundaries of chronostratigraphic units do not necessarily correspond to natural boundaries in various parts of the world (Sloss, 1988b), including Washington. The age of each superterrane shown in Figure 11 is the age of its accretion to the next regional unit to the east; obviously, rocks within a terrane can be considerably older than the age of accretion.

One of the most important features of Figure 11 is that the Kittitas, Walpapi, and High Cascade sequences unconformably overlie the Coast Range superterrane, but the Challis sequence does not. Thus, some discussion is worthwhile. However, the relations east of 120°W discussed by Cheney and others (1994, this volume) and the geology, history, and stacking order of the Insular and Northwest Cascade superterranes described by McGroder (1991) will not be repeated.

The Coast Range superterrane is generally regarded as the area underlain by the Crescent Formation and its correlatives (Brandon and Vance, 1992). The Coast Range superterrane as shown in Figure 11 includes the Crescent Formation, the Eocene sedimentary rocks above the Crescent but below the basal unconformity of the Kittitas sequence, Eocene metasedimentary rocks in the core of the Olympic Peninsula that Snavely and Kvenvolden (1989) and Brandon and Vance (1992) consider to be distal equivalents of the Crescent basalts that have been subducted beneath them, and the <80-km^2 Sooes terrane of Jurassic to Tertiary rocks (Snavely and Kvenvolden, 1989) on the northwestern coast of the Olympic Peninsula. This definition excludes the Hoh assemblage and the Kittitas and younger sequences.

Figure 11. *(facing page)* Geologic map of Washington. Sources of data: Figure 2, Huntting and others (1961), Rau (1967), Tabor and others (1987), Walsh and others (1987), Cheney (1987), Catchings and Mooney (1988), Finn (1990), McGroder (1991), Yount and Gower (1991), Schuster (1992), P. D. Snavely *in* Christiansen and Yeats (1992), Brandon and Vance (1992), and Cheney and others (1994, this volume). To better show subsurface relations in the Puget Lowland, the extent of the High Cascade sequence is less than shown by Schuster (1992).

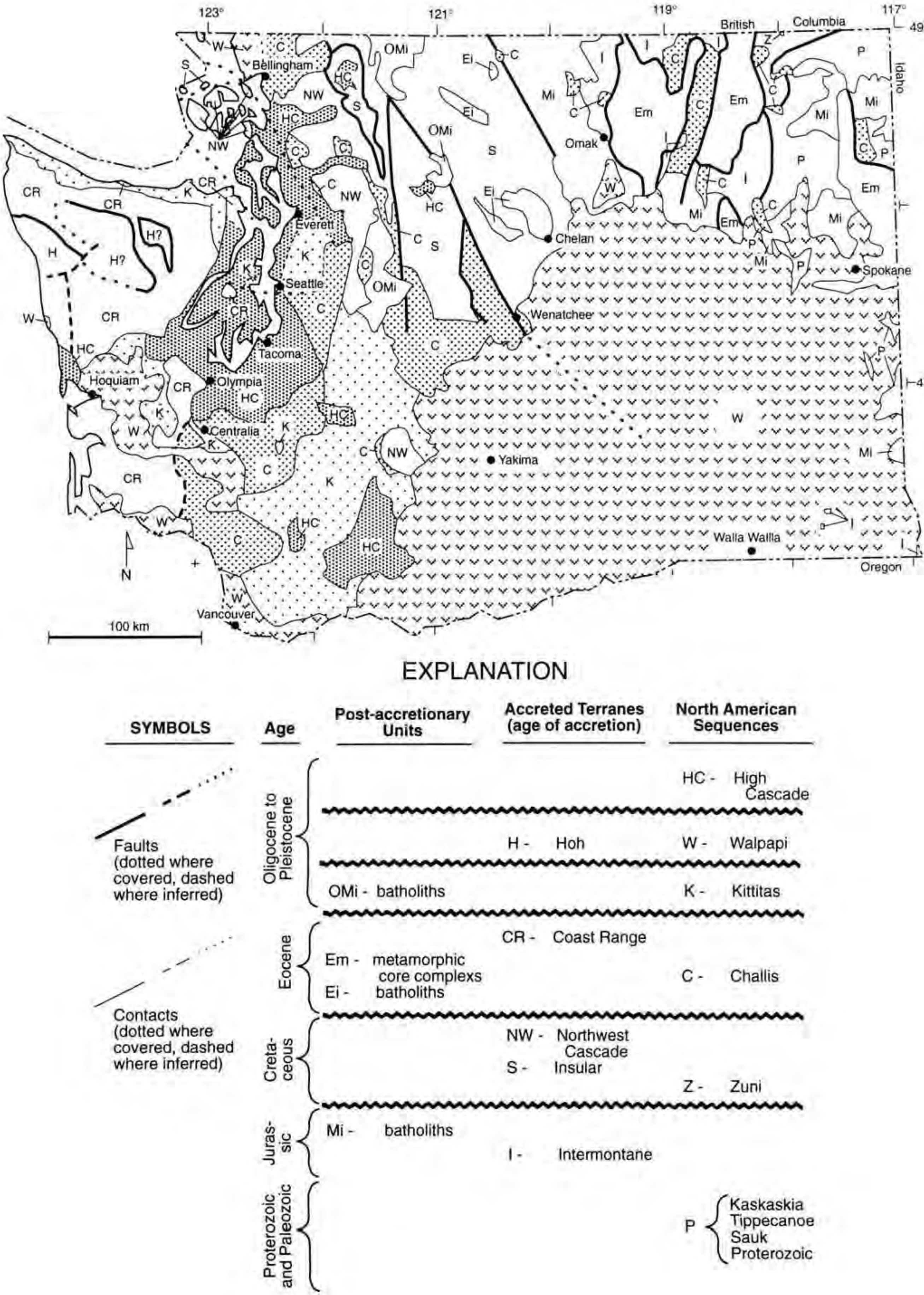
The Hoh consists of upper Oligocene to middle Miocene broken formation and mélange on the west coast of the Olympic Peninsula (Snavely and Kvenvolden, 1989). Some metasedimentary rocks in the core of the Olympic Mountains have unreset fission-track ages from detrital zircons of 27–19 Ma and reset ages on zircons as young as 14 Ma. Brandon and Vance (1992) suggested that these might be Hoh-equivalent rocks.

In the Puget Lowland, the eastern extent of the Coast Range rocks is well marked by steep gravity gradients caused by the Crescent basalts (Finn, 1990; Yount and Gower, 1991). However, as any subsurface faulted contact of the Coast Range rocks with the Challis sequence may be unconformably overlain by younger sequences, no contact is drawn in Figure 11 along the southwesterly trending gravity gradients on the southeastern side of the Olympic Peninsula and from Seattle to Olympia.

Another notable feature of Figure 11 is that the Walpapi sequence extends almost to the Pacific Ocean. Within this sequence, the western edge of the CRBG is northwest of Hoquiam (Walsh and others, 1987). Rather than flowing down the Columbia River valley, up the Cowlitz valley, and down the Chehalis valley to Hoquiam, it seems more likely that the basalts originally extended at least this far northwest before later uplift caused them to be eroded from the present site of the Cascade Range. If so, the area and volume of the CRBG were once larger than estimated by Tolan and others (1989).

Terrestrial Miocene strata more than a kilometer thick occur in the subsurface northwest of Bellingham (Cheney, 1987). These strata (and the unconformably underlying Challis sequence) are truncated on the southwest by the northwesterly trending Mount Vernon fault (Cheney, 1987, fig. 2), which is shown as a dotted line in Figure 11.

A significant intra-Walpapi unconformity occurs in western Washington. The Wilkes Formation, which unconformably overlies the CRBG south of Centralia (Roberts, 1958), appears to be similar to the Miocene strata north-


123°
121°
119°
117°
49°
47°
British
Columbia
Idaho
Oregon
Bellingham
Everett
Seattle
Tacoma
Olympia
Hoquiam
Centralia
Vancouver
Omak
Chelan
Wenatchee
Yakima
Spokane
Walla Walla
N
100 km
EXPLANATION
SYMBOLS
Faults (dotted where covered, dashed where inferred)
Contacts (dotted where covered, dashed where inferred)
Age
Oligocene to Pleistocene
Eocene
Creta-ceous
Juras-sic
Proterozoic and Paleozoic
Post-accretionary Units
Accreted Terranes (age of accretion)
North American Sequences
HC - High Cascade
H - Hoh
W - Walpapi
OMi - batholiths
K - Kittitas
CR - Coast Range
Em - metamorphic core complexes
Ei - batholiths
C - Challis
NW - Northwest Cascade
S - Insular
Z - Zuni
Mi - batholiths
I - Intermontane
P
Kaskaskia
Tippecanoe
Sauk
Proterozoic

west of Bellingham. The marine equivalent of the Wilkes, the Montesano Formation (Armentrout, 1987), rests unconformably upon Kittitas rocks south of the Olympic Mountains (Rau, 1967; Armentrout, 1987). On the west coast of the Olympic Peninsula, the Quinault Formation, which is correlative with the Montesano Formation (Tabor and Cady, 1978; Armentrout, 1987), unconformably overlies Hoh and Coast Range rocks (Tabor and Cady, 1978; Snavely and Kvenvolden, 1989). The unconformity at the base of the Wilkes Formation shows that at least some erosion of the CRBG was intra-Walpapi.

Causes of Interregional UBS

The map and explanation of Figure 11 provide insights for the probable causes of some unconformities. The sub-Challis unconformity developed after the amalgamation of the Northwest Cascade and Insular superterranes. Unconformities within the Challis sequence most likely formed in response to development of the metamorphic core complexes east of 120°W and to movement on major faults east of 122°W.

The sub-Kittitas unconformity could be the result of either of two events. The youngest ages of the Crescent basalts, the post-42 Ma and pre-30 Ma age of the Leech River fault, and basaltic debris in basal Kittitas strata suggest that the sub-Kittitas unconformity was the result of uplift and erosion caused by the docking of the Coast Range superterrane with North America (and its covering Challis sequence). However, in southern Oregon this docking occurred by 50 Ma (Snavely and Wells, 1991) and thus may be another cause of intra-Challis unconformities.

The Makah Formation, which is the basal formation of Armentrout's (1987, p. 300) Sequence III, the Kittitas sequence, and which occurs on the northwestern tip of the Olympic Peninsula, provides an alternative explanation for the sub-Kittitas unconformity. The Makah Formation unconformably overlies the Crescent Formation and the Crescent thrust, along which the Eocene rocks of the core of the Olympic Mountains were subducted below the Crescent Formation (Snavely and Kvenvolden,1989). This underthrusting must postdate the youngest unreset zircons (32.6 ± 2.4 Ma) in the footwall rocks (Brandon and Vance, 1992). Assuming that any partial resetting of this fission-track age caused a <10 percent reduction of the age of these zircons (Brandon and Vance, 1992, p. 591), these zircons might be ≥36 Ma. Thus, the sub-Kittitas unconformity may have been caused by uplift accompanying this subduction below the Crescent Formation (Snavely and Kvenvolden, 1989). Possibly this subduction and the docking of the Coast Range superterrane (as marked by the Leech River fault) are so nearly of the same age that the sub-Kittitas unconformity represents both.

Whatever caused the sub-Walpapi unconformity is not particularly obvious in Washington. It and the overlying CRBG most likely are related to extension in the Basin and Range Province (Swanson and others, 1989, fig. 19), as are structures within the CRBG (Hooper and Conrey, 1989). If the CRBG is due to a mantle plume (now expressed as the Yellowstone hotspot), the sub-Walpapi unconformity also might be due to the uplift and erosion that preceded eruption of the plume.

Snavely and Kvenvolden (1989) inferred that the contacts between the Hoh rocks and Eocene sedimentary rocks in the western part of the Olympic Peninsula are tectonic, and Brandon and Vance (1992) made the same inference with respect to possible Hoh rocks in the core of the Olympic Mountains. If so, accretion of the Hoh rocks probably caused the intra-Walpapi unconformity beneath the Wilkes, Montesano, and Quinault Formations and correlative rocks.

Presumably, the sub-High Cascade unconformity was caused by initiation of the Cascadia subduction zone farther offshore. A seismic reflection profile across the abyssal plain and lower continental slope southwest of Vancouver Island shows the Cascadia subduction zone as a landward dipping thrust that has folded and uplifted Pliocene and Pleistocene abyssal sediments by about 700 m, but this fault has been inactive during the Holocene (Snavely and Wells, 1991). Onshore, erosion continues, but a regional peneplain has not developed yet.

CONCLUSIONS

Four major interregional Cenozoic unconformity-bounded sequences occur in, and extend beyond, Washington. These are the Eocene Challis sequence, the Oligocene to Miocene Kittitas sequence, the Miocene to Pliocene Walpapi sequence, and the Pliocene to Pleistocene High Cascade sequence. Even though Figure 1 shows that each of the major sequences, or parts of them, were recognized by previous authors, they commonly have not been recognized by regional mappers and compilers.

Each of the four major sequences contains regional unconformity-bounded sequences. The regional Challis sequences are more extensive than the grabens or basins in which they commonly are inferred to have been deposited. The same stratigraphic order of regional Challis sequences in many areas implies that these areas were not separate local depositional basins, but post-depositional structural lows that have preserved remnants of regionally extensive units.

I suspect that the major reasons so many authors have adopted the model of local basins include the following: (1) these Cenozoic sedimentary and volcanic rocks are near the present continental margin; (2) the Challis sequence, in particular, has very thick intervals of clastic rocks; (3) the present topography of western Washington is not "cratonic-looking"; and (4) Crowell (1974) discussed thick sections of Neogene clastic rocks in the aulacogenic margin of California. Without an obvious lithostratigraphy or marker units to guide interpretations in Washington, Crowell's aulacogenic model seemed attractive. However, the most important reason probably is that, although sequence stratigraphy is more than four decades old (Sloss, 1988c), only recently has it been dramatically applied (Mitchum and others, 1977), recognized (ISSC, 1987), and documented (Haq and others, 1987). Furthermore, the applicability of sequence stratigraphy to active and passive continental margins is generally assumed to be minimal (Sloss, 1988a).

Another important conclusion is that the Cascade volcanic arc has not been continuously active for the last 36 Ma. The volcanic and volcaniclastic rocks of the southern Cascade Range consist of three distinct unconformity-bounded sequences (Kittitas, Walpapi, and High Cascade) and are punctuated by two major, interregional unconformities (the sub-Walpapi and the sub-High Cascade, of about 36 Ma and 2 Ma respectively). Other regional unconformities occur within the three sequences. The bulk of the volcanic and volcaniclastic rocks of the southern Cascade Range are part of the Kittitas sequence, which is 18–34 Ma older than the present topographic entity known as the Cascade Range, including its <2 Ma volcanoes. Furthermore, the Kittitas sequence is not geographically restricted to the present (<2 Ma) Cascade volcanic arc.

I anticipate that, in the next decade, sequence stratigraphy will be as widely applied in Washington as elsewhere in North America. Part of the stimulus will be the availability of seismic reflection lines in the western part of the state and offshore. Even more important will be the geological and geochronological databases created by current 1:100,000 regional mapping by the U.S. Geological Survey and by the 1:250,000 compilations of the Washington Division of Geology and Earth Resources.

Future applications of sequence stratigraphy will outnumber the few discussed here. The associated unconformities will emerge as the natural datum planes to measure the extent, magnitude, and timing of tectonic events. Sequences, like terranes, will be used as the natural units (as opposed to manmade chronostratigraphic units) on regional and interregional geologic maps.

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Tree split by ground cracking in the head scarp of a landslide on Van Zandt Dike, Whatcom County. There are about 30 annual growth rings in the tree. This most recent slide is a reactivation of an older slide that reached the Nooksack River. Photo by Timothy J. Walsh, 1987.

The Crescent "Terrane", Olympic Peninsula and Southern Vancouver Island

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ABSTRACT

The lower Eocene to Pliocene Crescent "terrane" is exposed on the Olympic Peninsula of Washington and southern Vancouver Island, British Columbia. The Crescent "terrane" consists of three lithotectonic units: (1) the lower Eocene Blue Mountain unit that is composed mainly of continentally derived turbidites; (2) the Eocene Crescent Formation that consists mainly of voluminous submarine and subaerial tholeiitic basalts partly intertongued with the Blue Mountain unit and interbedded with a variety of intrabasinal and terrigenous sedimentary rocks; and (3) an upper Eocene to Pliocene section of marine sedimentary rocks. Although the Crescent "terrane" is fault-bounded and lithologically distinct, we do not interpret it as an accreted microplate. We consider this package of rocks to be the product of rifting along the continental margin of North America as a result of plate kinematics, possibly influenced by the regional effects of the Yellowstone(?) hotspot.

Rifting began at about 60 Ma with subsidence of a marginal marine basin in which sediments of the Blue Mountain unit were deposited. From about 58 to 45 Ma, continued lithospheric extension induced mantle upwelling and decompression partial melting, which generated an estimated 100,000 cubic kilometers of tholeiitic basalt—the Crescent volcanic rocks. Several extrusive centers were involved, and many formed emergent islands in subsiding basins, as indicated by the transition from submarine to subaerial flows.

Lithologies in the upper sedimentary section of the northern Olympics indicate that Crescent underthrusting and resultant uplift of southern Vancouver Island persisted from the time of Aldwell deposition (ca. 45 Ma) through deposition of the Makah Formation (ca. 23 Ma). Thereafter the Pysht and Clallam Formations (and the Sooke Formation) reflect the rapid filling of the Tofino–Fuca basin to shallow marine and emergent conditions. The Tofino–Fuca basin was apparently isolated from the Cascade Range and other eastern sources until deposition of the Clallam Formation.

No sedimentary rocks younger than Miocene are found in the northern part of the Crescent "terrane", but to the south sedimentation continued in shallow marine basins through the Pliocene. Lithologies of these rocks demonstrate that exposure of plutonic sources in the Cascade Range occurred in the early to middle Miocene and that uplift of the Olympic subduction complex into its present domal form began at about 10 to 12 Ma.

INTRODUCTION

At least five major lithotectonic units (terranes?) make up the Olympic Peninsula of northwest Washington State (Fig. 1). Toward the continental margin are four highly disrupted, subduction-related, mélange and turbidite units (Snavely and others, 1986; Snavely, 1987): (1) the upper middle Miocene to upper Oligocene Hoh mélange; (2) the upper and middle Eocene Ozette mélange; (3) a middle and lower Eocene unit situated between the Crescent and Calawah faults, informally named the Waatch Point assemblage; and (4) the upper Eocene to Jurassic(?) rocks of the Sooes "terrane". Collectively, these "terranes" are called the "core rocks" by Tabor and Cady (1978) and the "Olympic subduction complex" by Brandon and Calderwood (1990).

The fifth "terrane" of the Olympic Peninsula and adjacent Vancouver Island is composed of the Crescent Formation and associated sedimentary rocks, which form a horseshoe-shaped outcrop belt surrounding the inboard side of the Olympic subduction complex (Fig. 2). In this paper, we define the Crescent "terrane" to consist of three distinctive lithotectonic units that formed along the western margin of North America from the early Eocene to the Pliocene. Early Eocene marine sedimentary rocks of the Blue Mountain unit (Tabor and Cady, 1978) form a lower unit. These are partly interbedded with the overlying Crescent Formation (Arnold, 1906; Brown and others, 1960), which consists predominantly of basalt flows and breccias interbedded with subordinate clastic sediments and limestone. Overlying and partly interbedded with the Crescent Formation are marine sedimentary rocks that range in age from late Eocene to Pliocene. This definition of the Crescent "terrane" corresponds to that of the "peripheral rocks" of Tabor and Cady (1978) but differs from the designation of Silberling and others (1987), who included only the Crescent Formation in the Crescent "terrane". Paleomagnetic data from the Crescent basalts of the central and northern Olympics (Babcock and others, 1992; Warnock, 1992) indicate that there has been no significant rotation or translation of these rocks since extrusion. In contrast, basalts of the Black Hills, Willapa Hills, and the Siletz River terrane show a consistent post-Eocene 40–80 degrees clockwise rotation

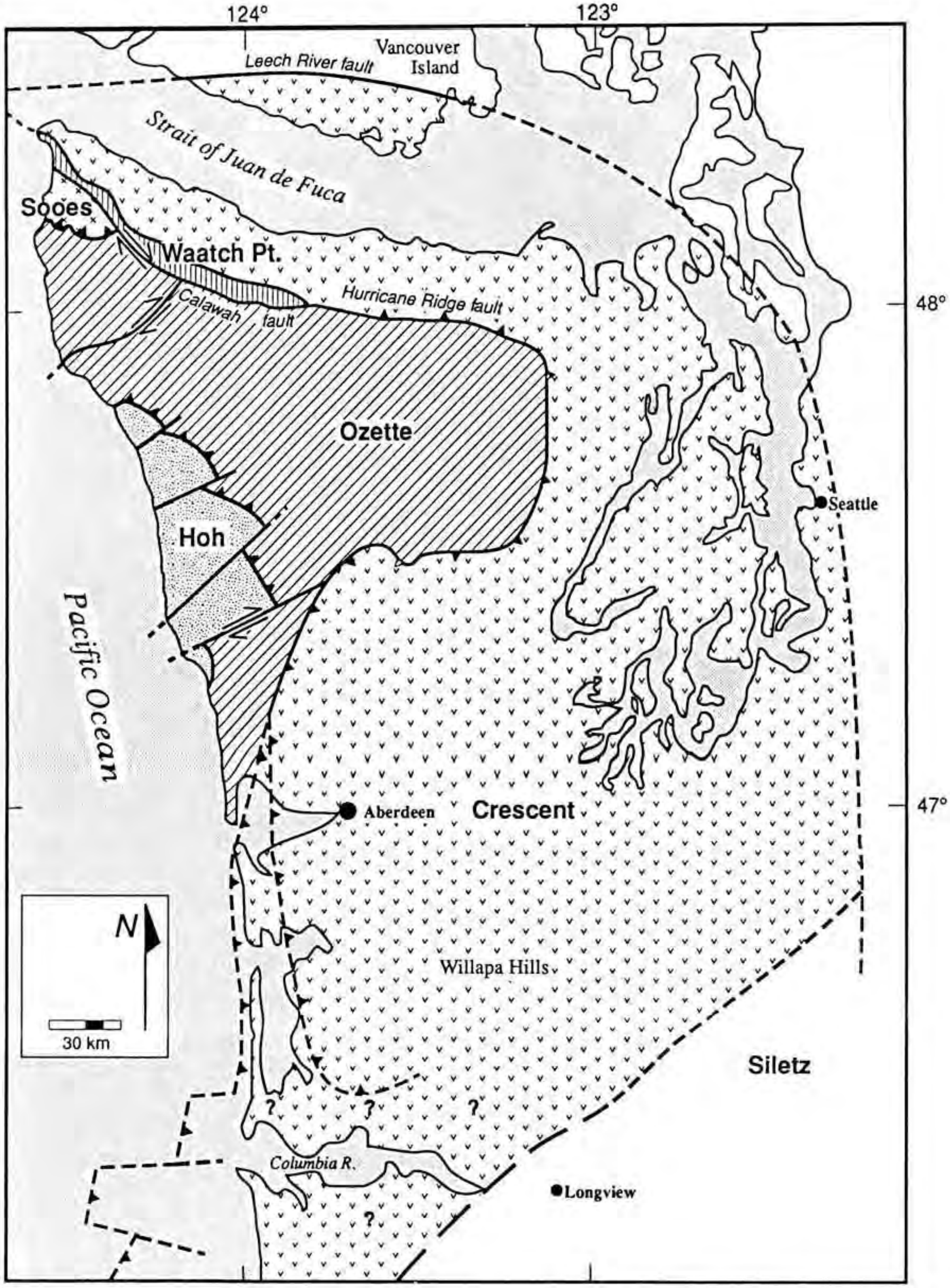

Figure 1. Lithotectonic units (terranes?) composing the Olympic Peninsula and southern Vancouver Island. Boundaries of the Sooes, Ozette, and Waatch Point "terranes" are as defined by Snavely and others (1986) and Snavely (1987). The term "Waatch Point" used in this paper corresponds to the unnamed "terrane south of Crescent thrust fault and north of Calawah fault" of Snavely and others (1986). The northern and western boundaries of the Crescent "terrane" are as inferred from geophysical data by Clowes and others (1987), except for the portion on southern Whidbey Island, which is drawn to account for the probable presence of Crescent Formation basalt in the Kingston well (W. W. Rau, Wash. Div. Geol. and Earth Res., written commun., 1992). The boundary between the Crescent "terrane" and the Siletz River terrane is uncertain and is placed as shown by Silberling and others (1987).

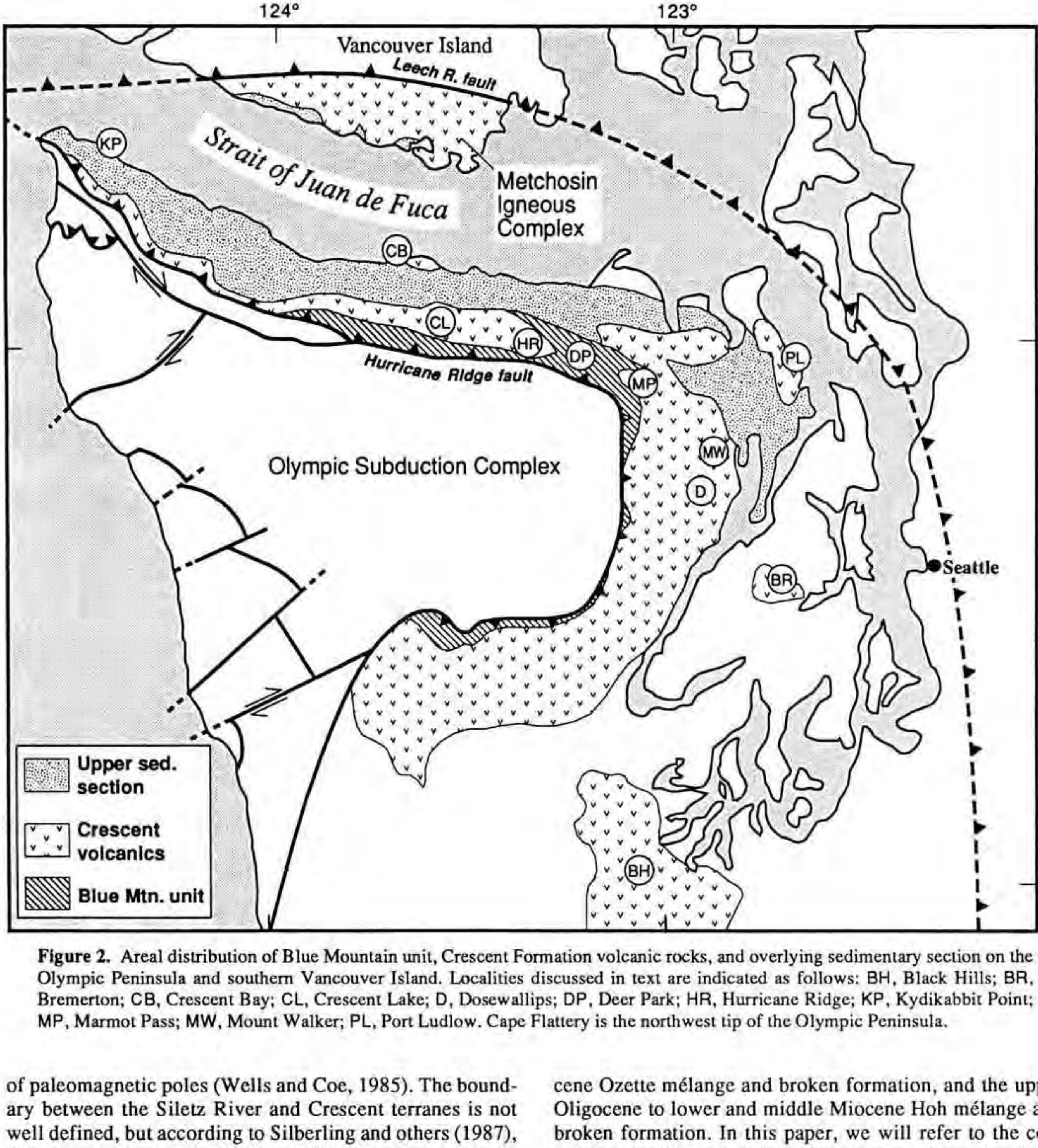

Figure 2. Areal distribution of Blue Mountain unit, Crescent Formation volcanic rocks, and overlying sedimentary section on the Olympic Peninsula and southern Vancouver Island. Localities discussed in text are indicated as follows: BH, Black Hills; BR, Bremerton; CB, Crescent Bay; CL, Crescent Lake; D, Dosewallips; DP, Deer Park; HR, Hurricane Ridge; KP, Kydikabbit Point; MP, Marmot Pass; MW, Mount Walker; PL, Port Ludlow. Cape Flattery is the northwest tip of the Olympic Peninsula.

of paleomagnetic poles (Wells and Coe, 1985). The boundary between the Siletz River and Crescent terranes is not well defined, but according to Silberling and others (1987), it lies somewhere in the Willapa Hills of southwest Washington (Fig. 1).

The Crescent "terrane" is everywhere separated by faults from adjacent, coeval rocks of the "core" terrane, which consists of several distinctive lithic assemblages as defined by Tabor and Cady (1978). The core rocks range from pervasively sheared greenschist-facies metasedimentary rocks to relatively undeformed turbidites with minor interbeds of basalt and tephra. Snavely (1987) subdivided the core rocks into two terranes: the middle to upper Eocene Ozette mélange and broken formation, and the upper Oligocene to lower and middle Miocene Hoh mélange and broken formation. In this paper, we will refer to the core rocks as the Olympic subduction complex, following Brandon and Calderwood (1990).

Snavely and others (1986) also delineated two other terranes separated by faults from the Crescent "terrane" in the northern Olympic Peninsula. The Sooes "terrane" consists of a variety of upper Eocene to Jurassic(?) sedimentary, volcanic, and plutonic (granodiorite) rocks. Another terrane, located between the Crescent fault and the Calawah fault, consists of lower and middle Eocene sandstones, siltstones, and conglomerates plus minor basalt flows and breccia that we informally call the Waatch Point "terrane".

The Crescent "terrane" extends across the Strait of Juan de Fuca to Vancouver Island, where exposures of basalt, gabbro, and sheeted diabase dikes are called the Metchosin Igneous Complex (Massey, 1986). The Metchosin rocks are in fault contact with the Leech River schist, which has a 38–41 Ma metamorphic age (Fairchild and Cowan, 1982). Seismic refraction data show that the Crescent–Metchosin volcanic rocks are thrust beneath Vancouver Island along the Leech River fault (Clowes and others, 1987).

Seismic and aeromagnetic data indicate that the Crescent basalts also extend to the east under the Puget Lowland and the westernmost Cascades (Finn, 1990; Clowes and others, 1987). In the northern Puget Sound, gravity interpretation suggests that subsurface normal faults offset the Crescent basalts with vertical displacements as much as 15 km (Roberts, 1991).

The purpose of this paper is to describe the lithology and tectonic history of rocks that make up the Crescent "terrane". This review draws heavily on previously published work, most notably Brown and others (1960), Gower (1960), Tabor and Cady (1978), Cady (1975), Rau (1981), Snavely and others (1986), Snavely (1987), Armentrout (1987), Tabor (1989) and Palmer and Lingley (1989). We also summarize two decades of thesis research in the Olympics by graduate students at Western Washington University. With the exception of radiometric ages for basalts, we will designate ages by the West Coast Standard foraminiferal stages (Rau, 1964, 1981; Squires and others, 1992) with reference to the time scale of Harland and others (1989).

LITHOLOGY

Blue Mountain Unit

The Blue Mountain unit of the Crescent Formation is an informal name for marine sediments that underlie and interfinger with the overlying Crescent volcanic rocks (Tabor and Cady, 1978). There are no direct radiometric or paleontological data on the age of the Blue Mountain unit, so it can only be assumed to be as old as or older than the early to mid-Eocene Crescent volcanic rocks (Rau, 1964).

The Blue Mountain unit generally crops out in a basal stratigraphic position between the Hurricane Ridge fault (the probable terrane boundary) and the overlying volcanic rocks of the Crescent Formation (Fig. 2). However, on the northeast side of the "Crescent horseshoe", the Blue Mountain unit appears to fill the gap between two separate volcanic centers and is directly overlain by rocks of the upper sedimentary section (Aldwell Formation). The contact between the Blue Mountain unit and the Crescent volcanic rocks is varied. Tabor and others (1972), Cady and others (1972a, 1972b) and Tabor and Cady (1978) provide field evidence for a depositional contact along the eastern part of the Peninsula. In contrast, detailed studies by Einarsen (1987) in the northwest and eastern parts show that the contact is both faulted and depositional. The most convincing evidence of a depositional contact is in the Marmot Pass area. Along the trail to the Tubal Cain mine, Blue Mountain unit siltstones can be found sandwiched between two pillow basalt flows. A baked zone is irregularly developed in the siltstone near the base of the upper flow, and siltstone fills the interstices between pillows of the lower flow. A baked zone in shales at the base of a pillowed flow also can be found along the road to Deer Park.

The Blue Mountain unit is dominated by thin-bedded turbidites that probably represent a middle to outer fan depositional environment. Massive sandstone beds also occur throughout the Blue Mountain unit. These generally lack sedimentary structures except for grading, and they show evidence for deposition in inner or middle fan channels. Subordinate massive siltstone beds are finely laminated or structureless and represent deposition on distal interchannel portions of a fan. A major submarine channel in the middle of the Blue Mountain unit is marked by pebbly sandstone and conglomerate (Einarsen, 1987).

Two end-member petrofacies and mixtures thereof are recognized by Einarsen (1987): a plagioclase-rich feldspathic arenite and a chert-rich lithic arenite. According to Einarsen (1987), the plagioclase-rich facies had a plutonic source, most likely the Coast Plutonic Complex, and the chert-rich facies had a San Juan Islands provenance. Derivation from sources to the north or northeast is supported by sparse paleocurrent data that indicate a southwest flow direction near Blue Mountain (Snavely and Wagner, 1963).

Crescent Formation Volcanic Rocks

The Crescent Formation was originally named by Arnold (1906) for outcrops of basalt at Crescent Bay, whereas equivalent rocks on southern Vancouver Island were called the Metchosin Volcanics and Sooke Gabbro by Clapp and Cooke (1917). Brown and others (1960) redefined the Crescent Formation to include rocks between Lake Crescent and Hurricane Ridge. The term is now generally applied to all lower to middle Eocene basalts from the Willapa Hills to southern Vancouver Island. However, despite the obvious correlation with the Crescent volcanic rocks and the precedence of Arnold's name, Canadian geologists prefer the terms Metchosin Volcanics (Muller, 1977, 1980) or, more properly, the Metchosin Igneous Complex (Massey, 1986) because the unit includes gabbroic intrusive rocks, a sheeted dike complex, and a thick section of basalt flows.

Cady and others (1972a, 1972b) and Tabor and others (1972) subdivide the Crescent volcanic rocks into a subaerial upper basalt member and a submarine lower basalt member. Their lower basalt member is composed mainly of pillowed flows with numerous diabase or gabbroic dikes and sills. Interbeds of basaltic breccia, hyaloclastites, basaltic sandstone, chert, and limestone are subordinate, but they occur throughout the lower member. Their upper basalt member consists predominantly of columnar to massive flows that have oxidized tops and paleosols developed in places.

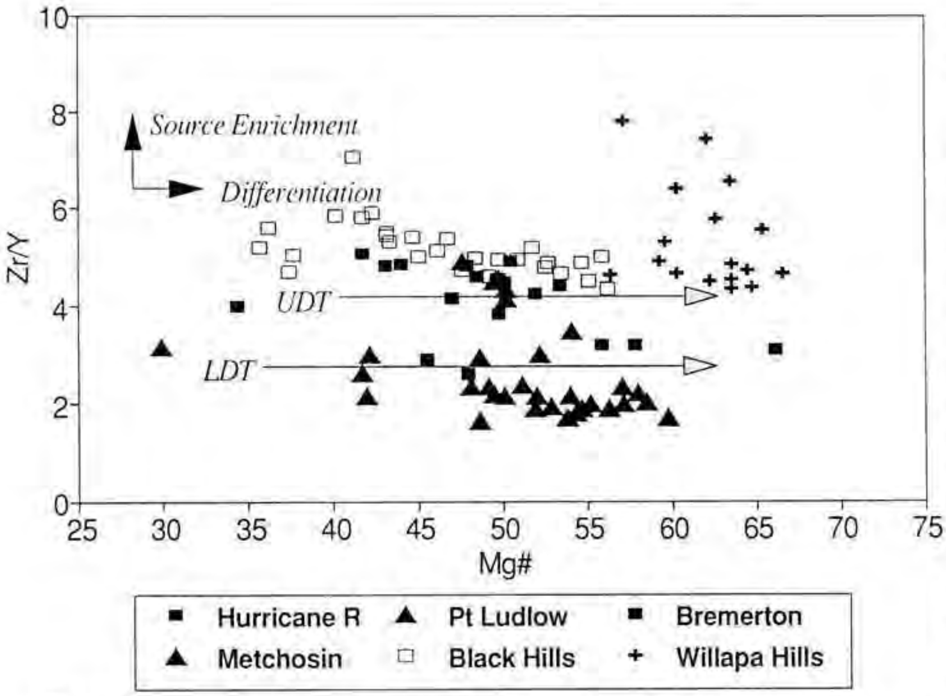

Figure 3. Regional variation in the trace element geochemistry of the Crescent basalts. Trend lines for the submarine (LDT) and subaerial (UDT) basalt sections near the Dosewallips River valley are shown by arrows. Mg#, mol % MgO/FeO-total + MgO. Note the systematic increase in the degree of source enrichment from north (Metchosin) to south (Willapa Hills). Data plotted from Babcock and others (1992), Clark (1989), Globerman (1981), Muller (1980), and Phillips and others (1989).

Glassley (1974) argues that the lower member represents typical oceanic crust (MORB) and the upper member resembles oceanic island basalt (OIB). Similarly, Muller (1980) proposes that the Metchosin represents conditions equivalent to an Icelandic setting where a plume-dominated oceanic island is superimposed on a mid-oceanic ridge. Cady (1975) disagrees with this interpretation and considers the Crescent to be a continuous volcanic sequence because he found no clear geochemical, petrologic, or structural distinction between the upper and lower members. As we will discuss later, our geochronology and geochemistry of the Crescent volcanic rocks indicate that subdivision into upper and lower members is invalid.

Babcock and others (1992) describe a measured composite section of 8.4 km of predominantly submarine flows and 7.8 km of predominantly subaerial flows in and near the Dosewallips River valley. In this section, the subaerial flows are distinctly enriched in TiO_2, Sr, Nb, and light rare-earth elements compared to the submarine flows. However, this difference does not support subdivision into upper and lower members, nor Glassley's (1974) contention that two separate packages of basalt were tectonically juxtaposed. Figure 3 shows that there is a systematic increase in the degree of "enrichment" from the Metchosin on the north to the Willapa Hills to the south. Only in the extremely thick Dosewallips section is there a transition from submarine normal-MORB type basalts to more enriched-MORB type in the overlying subaerial flows.

Foraminifera from interbeds in the Crescent range in age from Penutian to Ulatisian (Rau, 1981). This corresponds with ^{40}Ar-^{39}Ar geochronometry (Fig. 4 and summarized below), which shows a similar range. In the Dosewallips section, the subaerial sections range from 50.5 ± 1.6 Ma near the top to 51.0 ± 4.6 Ma near the base. The submarine basalts have an age of 56.0 ± 1.0 Ma near to top. Fossils in sediments interbedded with the uppermost Crescent basalts at Pulali Point give ages equivalent to Penutian to Ulatisian foraminiferal stages (Squires and others, 1992), which is consistent with the 50.5 Ma age, using the time-scale correlations of Harland and others 1989). At Crescent Lake, a submarine flow just below the Aldwell contact yielded an age of 52.9 ± 4.6 Ma, while the base of the submarine flows exposed on the Hurricane Ridge road was dated at 45.4 ± 0.6 Ma. Elsewhere, Duncan (1982) gives an age of 57.8 ± 0.8 Ma for the Metchosin Igneous Complex and ages of 55.0 ± 0.9 Ma and 51.7 ± 2.4 Ma for subaerial flows in the Bremerton Igneous Complex. Clark (1989) reports ages of 50.4 ± 0.6 Ma and 49.2 ± 0.8 Ma for Bremerton complex leucogabbros. Only conventional K-Ar ages are available for basalt flows correlative to the Crescent in the southern part of the Olympic Peninsula. These may be unreliable due to thermal resetting, but the oldest ages range from 56 Ma in the Black Hills to 47 Ma in the Willapa Hills (Fig. 4). In summary, basalts of the Crescent Formation range in age from 45 to 58 Ma, and there is no apparent geographic progression in the age of extrusion. Exposures of basalt that have been mapped as contiguous (for example, those at Hurricane Ridge and Crescent Lake) must be part of separate extrusive centers. More geochronology is necessary before any valid reconstruction of the stratigraphy and paleogeography of the Crescent volcanic rocks can be accomplished.

POST-CRESCENT MARINE SEDIMENTARY SECTION

As volcanism waned at the various extrusive centers, sedimentation became dominant in shallow to deep marine basins and on-lapped both submarine and subaerial basalt piles. Our description of these sedimentary rocks will concentrate on the northern Olympic coast where most of our work has been done. We will also describe correlative sedi-

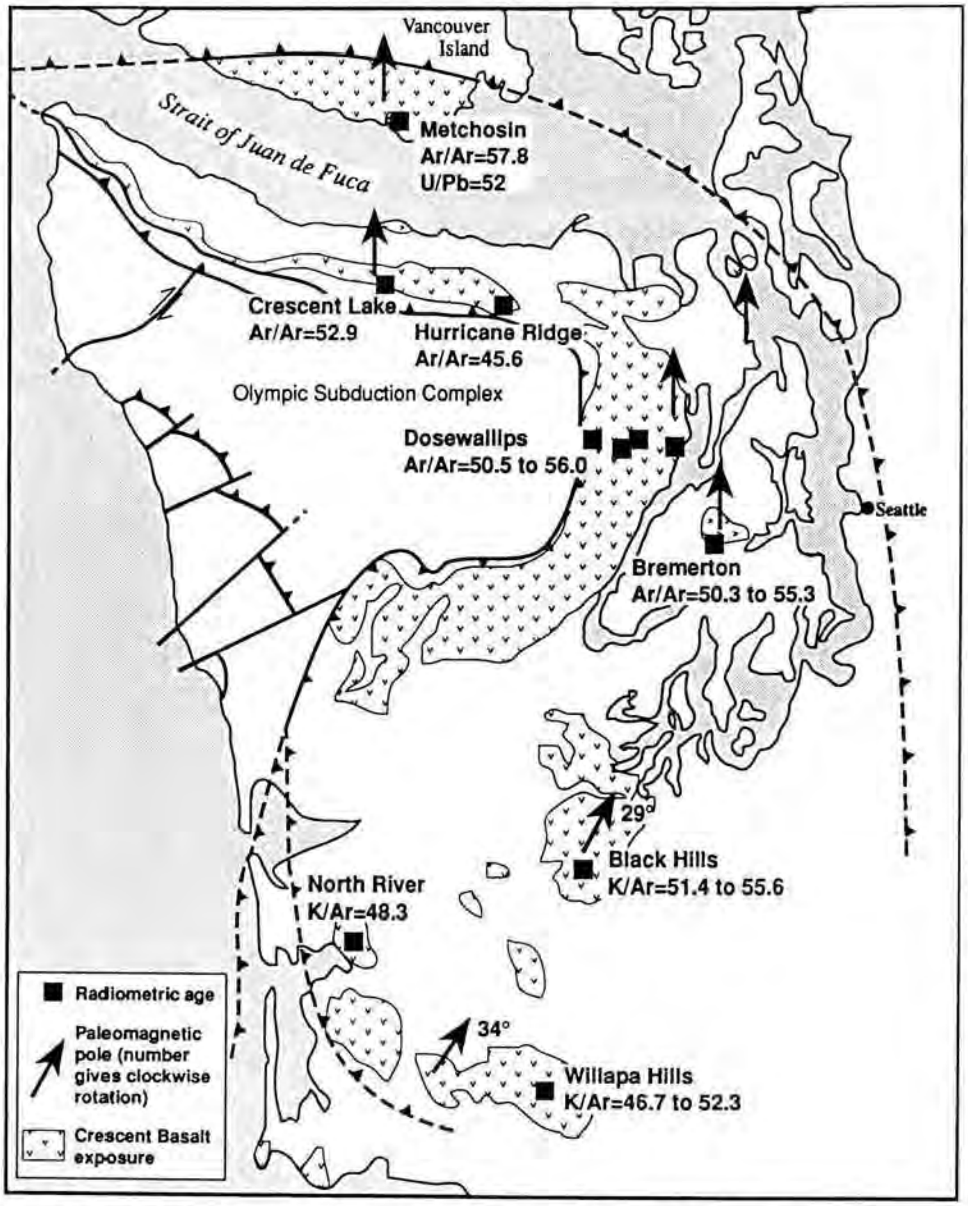

Figure 4. Paleomagnetic poles and radiometric ages obtained from the Crescent Formation volcanic rocks. Ar/Ar indicates $^{40}Ar/^{39}Ar$ plateau ages. Arrows and degrees show amount of clockwise rotation relative to an Eocene pole for cratonic North America. Data from Babcock and others (1992), Duncan (1982), Clark (1989), Massey (1986), Warnock (1992), Globerman and others (1982), and R. A. Duncan (Ore. State Univ., written commun., 1990).

mentary units elsewhere in the Olympics and southern Vancouver Island.

Northern Olympic Coast

Figure 5 shows the stratigraphy and areal distribution of post-Crescent marine sedimentary rocks on the northern Olympic coast as defined by Tabor and Cady (1978) and Snavely and others (1978, 1980, 1986). The pattern of northwest-trending belts reflects the macro-structure, which is the southern limb of a syncline that has an axis roughly parallel to the Strait of Juan de Fuca. The northern limb of the syncline is represented by exposures of basalt and overlying sedimentary rocks on southern Vancouver Island (Gower, 1960; Snavely and Wagner, 1981). Deposition of the upper sedimentary section occurred in a basin variously called the Juan de Fuca basin, the Tofino–Juan de Fuca basin, and the Tofino–Fuca basin. The last two terms refer to a deep marginal basin along the north side of the Olympic Peninsula that extended northward along the west coast of Vancouver Island. In this paper we will use the term Tofino–Fuca basin.

Aldwell Formation

The Aldwell Formation (Brown and others, 1960) is about 900 m thick in the type area. It was originally described as being unconformable on Ulatisian sediments of the Crescent Formation (Snavely and Lander, 1983) and assigned an age of early to late Narizian (Rau, 1981). However, at Pulali Point on the Hood Canal, we have found that Crescent flows are interbedded and intertongued with sedimentary rocks that are mapped as Aldwell(?) and that have been assigned a Penutian to Narizian age (Squires and others, 1992, p. 12, 14–19). The Aldwell consists mostly of poorly indurated, thin-bedded siltstone with interbedded sandstone, plus minor lenses of conglomerate and pillow basalt (Tabor and Cady, 1978; Marcott, 1984). It also includes olistostromes of Striped Peak volcanic rocks that may have been derived from areas uplifted by the Striped Peak thrust fault (Snavely, 1983).

Deposition of Aldwell sediments occurred mainly in cold, deep water (Rau, 1964); these sediments are interpreted to be mostly submarine landslides or mudflows (Brown and others, 1960; Squires and others, 1992). Snavely (1987) describes a discontinuous basal conglomerate derived from the underlying Crescent Formation. However, Babcock and others (1992) describe the Crescent volcanic rocks at Pulali Point as being interbedded with Aldwell sedimentary rocks. This is disputed by Squires and others (1992), who place the Crescent–Aldwell contact at an angular unconformity well above the uppermost Crescent flows.

Marcott (1984) delineates two lithofacies in Aldwell rocks exposed along the northwestern Olympic coast. An

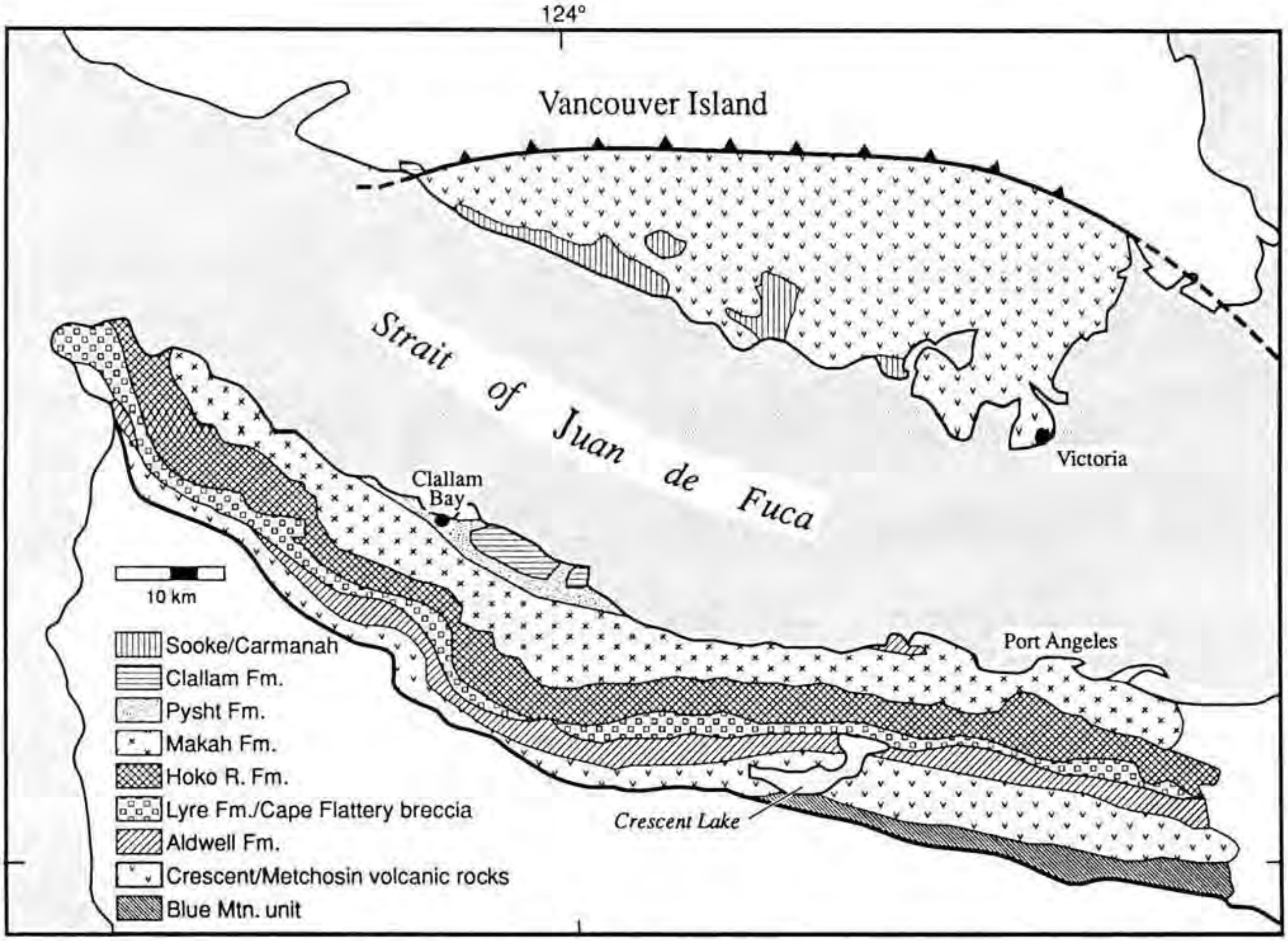

Figure 5. Geologic sketch map of the northern Olympic coast and southern Vancouver Island showing areal distribution of rock units in the Crescent "terrane". Compiled from Tabor and Cady (1978), Snavely and others (1980, 1986), Snavely (1983), and De Chant (1989).

eastern facies consists of thin-bedded laminated siltstone and lithic graywacke probably derived from Crescent "terrane" basalt on Vancouver Island and deposited in an outer fan environment (Marcott, 1984). A western facies consists of chert-rich lithic arenite deposited in the middle region of a fan separate from the eastern facies. Marcott (1984) interprets the source of the western facies to be uplifted portions of southern Vancouver Island. If so, the lithology of the Aldwell Formation may reflect the initiation of underthrusting of Crescent basalts along the Leech River fault.

Lyre Formation

The Lyre Formation (Weaver, 1937; redefined by Brown and others, 1956, 1960; Tabor and Cady, 1978) ranges from "<300 to >600 m" thick (Ansfield, 1972) and includes the distinctive breccia of Cape Flattery. The Lyre is in depositional contact with the underlying Aldwell Formation. On the basis of its microfossils, the Lyre is late Eocene (late Narizian) in age (W. W. Rau, cited in Snavely, 1983). Sandstones in the Lyre consist predominantly of lithic arenite and lithic wacke (Ansfield, 1972; Shilhanek, 1992) that were deposited under shallow- to deep-water conditions (Tabor and Cady, 1978). Spencer (1984) provides evidence that the basin deepened as deposition continued in the northeast Olympic Peninsula. The Cape Flattery breccia and laterally equivalent Lyre Formation represent submarine fan or base-of-slope mass sediment gravity flows in much shallower water than the Aldwell (Ansfield, 1972; Shilhanek, 1992). The restricted presence of angular boulders to the west and rounded, nearly polished boulders to the east indicate two separate sources with different distances of transport (Shilhanek, 1992). Two separate sources are also indicated by provenance studies that show Leech River and Pandora Peak clasts only in the western unit. Continued uplift of southern Vancouver Island during the late Narizian (ca. 40 Ma) is indicated by the presence of these metamorphic clast types in the Lyre Formation.

Hoko River Formation

The Twin River Formation, originally defined by Arnold and Hannibal (1913) and redefined by Brown and Gower (1958), was raised to group rank by Snavely and others (1978). The Twin River Group consists of three formations, the Hoko River, the Makah, and the Pysht, that can be traced more than 100 km along strike in the northern Olympic Peninsula (Fig. 5). The 1,600–2,300-m-thick, late Narizian Hoko River Formation is the lowest formation of the Twin River Group. It conformably overlies and (near

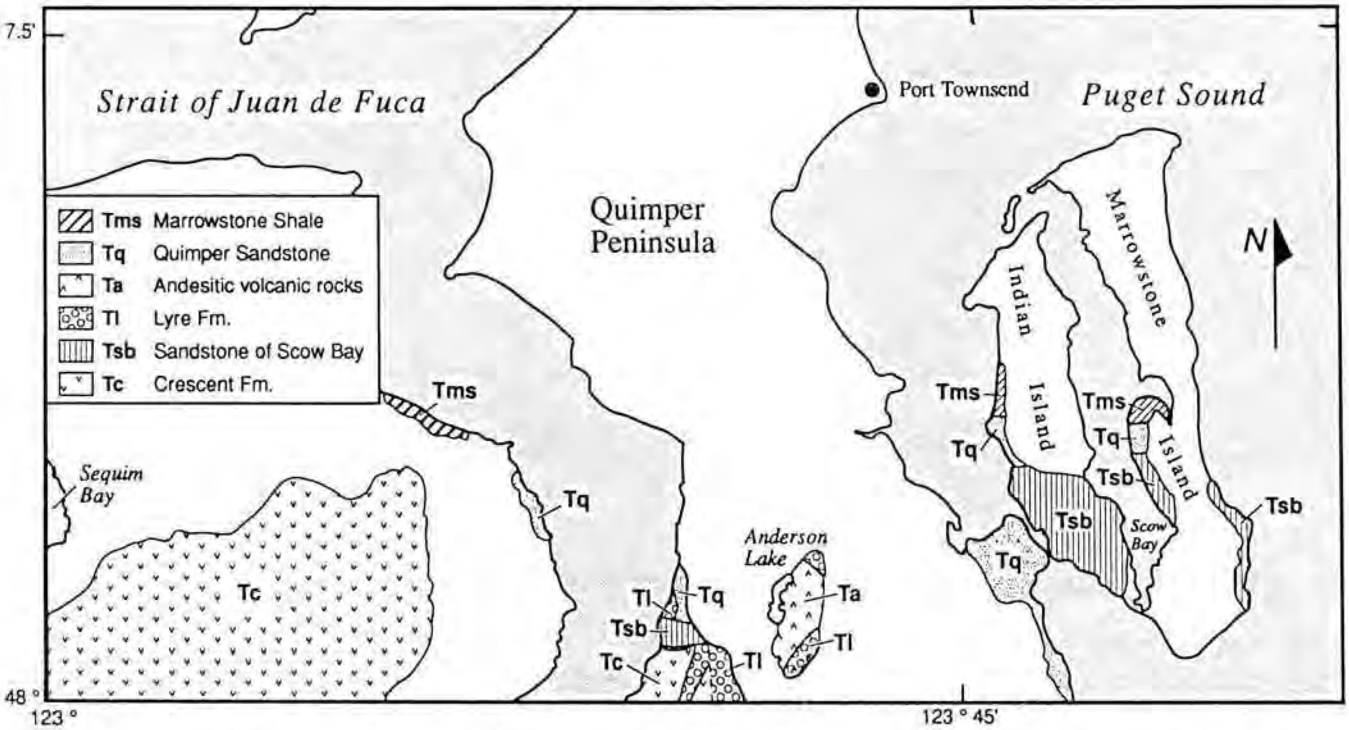

Figure 6. Sketch map of the bedrock geology of the Quimper Peninsula and adjacent areas. Compiled from Armentrout and Berta (1977), Rauch (1985), and Whetten and others (1988).

Cape Flattery, 48°23'N, 124°42'W) intertongues with sandstone and conglomerate of the Lyre Formation (Snavely and others, 1978). Most of this formation consists of massive to thin-bedded siltstone, with lesser phyllite-rich and basalt-rich sandstone and conglomerate. Hackly fracturing in the siltstone and sparse conglomerate interbeds distinguish the Hoko River from the overlying Makah Formation. Sandstone interbeds are lithic arenite and lithic graywacke (De Chant, 1989). A Vancouver Island provenance appears to dominate deposition from Aldwell and Lyre time through the Hoko River, although paleocurrent directions are inconsistent. For most of the Hoko River deposition, transport from the north is indicated, but some exposures show flow directions from the southwest and from the east-southeast (De Chant, 1989). De Chant (1989) interprets the depositional regime to have been outer neritic to bathyal depths in middle-fan channels, inner-fan channels, and depositional lobes. The Hoko River sediments represent greater water depths in the Tofino–Fuca Basin than the Lyre, but not as deep as most of the Aldwell (Rau, 1964).

Makah Formation

The Makah Formation, named by Snavely and others (1978) and described in detail by Snavely and others (1980), is late Eocene to Oligocene (late Narizian(?) to late Zemorrian) in age (Snavely and others, 1978). The Makah consists of more than 2,800 m of predominantly thin-bedded siltstone with sandstone interbeds defining six mappable members: four turbidite units, an olistostrome, and a water-laid tuff. The contact with the underlying Hoko River Formation is generally gradational and conformable (Snavely and others, 1978; De Chant, 1989). Deposition occurred at upper to middle bathyal depths on mid-fan depositional lobes and channels. Paleocurrent directions from the north or northwest and clast compositions indicate a Vancouver Island source, although some sandstone units seem to be more compatible with derivation from granitic rocks of the Coast Plutonic Complex.

Pysht Formation

The Pysht Formation, defined by Snavely and others (1978), is upper Oligocene (upper Zemorrian to lower Saucesian) and gradational with the underlying Makah Formation. The base of the Pysht is marked by channel deposits of pebble and boulder conglomerate with clasts of distinctive felsic tuff (Snavely and others, 1978). The Pysht consists of about 1,400 m of poorly indurated mudstone to sandy siltstone with calcareous concretions and conglomeratic channel deposits. The depositional environment is interpreted by Snavely and others (1980) to be a submarine channel system extending from shallow to mainly deep marine conditions and with a Vancouver Island source. The Pysht sediments show evidence for shoaling upward into shallow marine conditions and are gradational and conformable with the overlying Clallam Formation (Snavely and others, 1987).

Clallam Formation

The Clallam Formation, as restricted by Gower (1960), is an approximately 800-m-thick section of lower Miocene sandstone and conglomerate that contains abundant wood fragments and a shallow-water marine fauna. A detailed study by Anderson (1985) showed that the Clallam consists of feldspathic and lithic wacke plus arenite sandstones with subordinate conglomerates and minor siltstones that were deposited as shoaling-upward deposits in a prograding deltaic distributary environment. Anderson also recognized that the Cascade Range or San Juan Islands as well as Vancouver Island may have contributed sediment to the Clallam Formation. If correct, this would represent a distinct change in the Tofino–Fuca basin that previously had sediment sources restricted to Vancouver Island or the Coast Plutonic Complex(?). The general shallowing of the basin during deposition of the Pysht and Clallam Formations indicates cessation of any tectonic deepening due to thrusting of the Crescent "terrane" under Vancouver Island.

Quimper Peninsula Exposures

Another section of post-Crescent marine sedimentary rocks about 1 km thick occurs on the Quimper Peninsula and on Indian and Marrowstone Islands (Fig. 6). The lowermost unit consists of the informally named sandstones of Scow Bay (Allison, 1959). These are middle Eocene (Ulatisian and lower Narizian), according to Armentrout and Berta (1977). The exposed rocks consist of very thick bedded, typically structureless lithic arenite interbedded with minor siltstone, shale, and limestone. The sandstones probably represent mid-fan channel-fill deposits, possibly with a nearby volcanic center indicated by a vitric tuff bed (Melim, 1984). Several basaltic dikes that cut the sandstones of Scow Bay (Melim, 1984) are petrographically similar to subaerial Crescent flows exposed to the south near Port Ludlow, so the sandstones of Scow Bay could possibly be intertongued with the upper Crescent.

An angular unconformity separates the Scow Bay from the overlying Quimper Sandstone and Marrowstone Shale (Armentrout and Berta, 1977). These rocks, which are upper Eocene (Refugian), elsewhere unconformably overlie the Lyre Formation and are probably correlative with the Makah Formation. The Quimper contains a tuff layer (Rauch, 1985) that may be equivalent to the Carpenter Creek Tuff Member of the Makah. Three facies are recognized by Rauch (1985); overall, these show fining and deepening upward from outer shoreface and inner shelf to offshore or outer shelf. The lower Quimper is composed of amalgamated sandstone with hummocky cross-bedding indicating shallow deposition in a high energy environment. The upper Quimper consists of finely laminated silt and shale that were deposited below storm wave base, while the Marrowstone Shale was deposited in deeper, colder water as the basin rapidly subsided. The provenance of these rocks is indefinite but likely involved Vancouver Island, Coast Plutonic Complex, and North Cascades sources (Rauch, 1985). If the Discovery Bay fault zone is the northern boundary of the Crescent Terrane (as proposed by MacLeod and others, 1977), the Quimper Sandstone constrains the age of terrane accretion because it overlies the fault zone. However, an interpretation of gravity data on the Quimper Peninsula by Roberts (1991) indicates that the actual terrane boundary is not the Discovery Bay fault zone, but must be another structure (as yet unnamed) lying farther east under Puget Sound.

Post-Metchosin Section on Southern Vancouver Island

The Carmanah Group is subdivided into three formations by Muller (1977). From oldest to youngest these are: the Escalante Formation, the Hesquiat Formation, and the Sooke Formation. The Escalante Formation (Cameron, 1971) is upper Eocene in age and consists of about 300 m of calcareous sandstone with minor lenticular, shelly conglomerate and argillaceous sandstone (Jeletsky, 1975). It grades into the upper Eocene to Oligocene Hesquiat Formation, which is 1,200 m thick and ranges from a basal calcareous sandstone to argillaceous sandstone and siltstone with shale interbeds (Cameron, 1980). The Escalante and Hesquiat unconformably overlie the Leech River Complex, which has a metamorphic age of 38–41 Ma (Fairchild and Cowan, 1982). The Hesquiat, in turn, is unconformably overlain by the Oligocene Sooke Formation.

There is disagreement about the depositional environment of the Escalante Formation. Jeletsky (1975) interprets a shallow marine estuarine environment from fossil assemblages. Cameron (1980) disputes Jeletsky's interpretation of the faunal assemblages and instead argues for deposition in upper bathyal to lower neritic water depths. Uplift had begun by the time of Hesquiat deposition, and these sediments represent either a bathyal submarine fan (Cameron, 1980) or a deltaic environment (Jeletsky, 1975). The Sooke Formation consists of shallow marine sandstones and conglomerates that were deposited along the Oligocene strand line, which lay on the northern side of the present-day Strait of Juan de Fuca (Bream, 1987). Depositional ages of the Escalante and Hesquiat Formations are roughly coeval with those of the Hoko River and Makah, while Sooke sedimentation occurred at the same time as that of the Pysht. In general, the Carmanah Group sediments represent deposition more proximal to the source and in shallower water than equivalent units in the northern Olympics.

Southern Olympic Peninsula

Our description of the southern part of the Crescent "terrane" is based primarily on the results of an M.S. thesis by Bigelow (1987) on the Montesano Formation in the northern part of the Grays Harbor basin. For more thorough regional descriptions of the complicated stratigraphy of these rocks, the reader is referred to summary papers by Rau (1964, 1981), Armentrout (1987), Snavely (1987), and Palmer and Lingley (1989).

The oldest rocks of the southern Olympic Peninsula are basalts and interbedded sediments of the Crescent Formation. These are unconformably overlain by a 4,500-m section of forearc sediments of the middle to upper Eocene Cowlitz and McIntosh Formations (Ulatisian to Narizian; Rau, 1981); these represent mostly deltaic deposition (Armentrout, 1987). Upsection is a 600–2,700-m section of marine siltstone and tuffaceous sandstone comprising the upper Eocene to uppermost Oligocene (Refugian to Zemorrian; Rau, 1981) Lincoln Creek Formation. The overlying shallow marine to deltaic sediments of the Miocene (Saucesian to Luisian; Rau, 1981) Astoria(?) Formation generally lie above a regional unconformity (Bigelow, 1987). However Armentrout (1987) finds that to the west, away from Oligocene structural highs, the Astoria(?) conformably overlies the Lincoln Creek Formation. The uppermost fluvial, lacustrine, and shallow-marine Montesano Formation (Mohnian to Delmontian; Rau, 1981) is also both conformable and unconformable, depending on the structural setting (Armentrout, 1987). Bigelow (1987) reports marine mammal fossils, including the sea lion *Allodesmus* and baleen whales, in the lower part of the Montesano.

A detailed petrologic study of the Astoria(?) and Montesano sediments (Bigelow, 1987) reveals a significant intermediate to silicic source component with abundant glass that probably reflects contemporaneous explosive volcanic activity in the Cascade Range. The presence of granodioritic clasts indicates sufficient uplift to expose plutonic sources in the southern Cascades. The clast composition of the Montesano indicates the onset of rapid uplift of the Olympic subduction complex (core terrane) in the late middle Miocene (about 12 Ma); this corresponds quite well to fission-track ages of uplift given by Brandon and Calderwood (1990).

PALEOMAGNETISM AND PALEOTECTONICS

Paleomagnetic studies in the Crescent basalt are hampered by post-extrusive alteration, metamorphism, and obscured deformation. The first successful paleopole determinations were made by Beck and Engebretson (1982), who show that Crescent-correlative subaerial basalts near Bremerton and Port Townsend have been neither significantly rotated nor translated northward with respect to the Eocene pole of cratonic North America. This contrasts with poles from generally coeval basalts in the Black Hills, the Willapa Hills, and the Oregon Coast Ranges, which show evidence for consistent clockwise rotations of as much as 70 degrees (Wells and Coe, 1985). Warnock (1992) used a small-diameter core drill to sample rims of pillow basalts in the submarine section of the Crescent. He also sampled subaerial basalts by conventional methods. His study of primary magnetizations of basalts, at sites extending from Mount Walker to Hurricane Ridge, confirms that there has been no significant movement since extrusion (Beck and Engebretson, 1982). Warnock (1992) also shows that erosion of a partial dome or arch open to the west could have produced the horseshoe outcrop pattern of the Crescent Formation.

According to Moyer (1985), secondary components of magnetization from sedimentary rocks of the northwestern Olympics record a later deformational event that involved about 40 degrees of clockwise rotation. He also reports evidence for 70 degrees of counterclockwise rotation at Kydikabbit Point on the northwestern tip of the Olympic Peninsula. These data are dubious and difficult to interpret, but such deformation could possibly be related to fragmentation of the Crescent "terrane" as it was thrust under southern Vancouver Island.

The plate tectonic setting in which rocks of the Crescent "terrane" formed is open to debate. Duncan (1982) proposes that the early to middle Eocene volcanic rocks of the Oregon and Washington Coast Ranges represent a chain of seamounts generated by a hotspot (Yellowstone?) centered on the Kula–Farallon ridge. Convergence of these oceanic plates resulted in accretion of the seamounts to North America by about 42 Ma. Wells and others (1984) provide more details about the plate motions involved and suggest that the Coast Range basalts could also have originated in a continental margin rift or in leaky transform faults on the oceanic plates. Brandon and Massey (1985) and Snavely (1987) support the rift origin, while Babcock and others (1992) explain why rifting would be expected in the environment of oblique convergence and ridge subduction in which Crescent extrusion occurred. Both hotspot (Engebretson and others, 1985) and plate-circuit type reconstructions (Stock and Molnar, 1988) indicate that the Kula–Farallon–North America triple junction must have been in the general vicinity of the Oregon–Washington Coast Ranges during the early Eocene.

For this paper, we have reconstructed Eocene plate interactions (Fig. 7), incorporating refined estimates of Kula plate motion by Cashman (1990). These calculations show that the onset of Crescent volcanism at about 58 Ma was accompanied by a marked increase in the velocity and obliquity of the convergence of the Kula plate with North America. At the same time, the orientation of the Kula–Farallon ridge rotated about 60 degrees from highly oblique to nearly normal to the reconstructed continental margin. By about 42 Ma, the ridge orientation had again become more parallel to the continental margin and the angle of subduction more normal. The rate of convergence also decreased dramatically. The northward passage of the Farallon–Pacific–North America triple junction and the Sedna and Aja fracture zones could also have had a significant effect on regional tectonics.

DISCUSSION

For interpretive purposes, the most important aspect of the Crescent "terrane" is its stratigraphic progression—a basal marine sedimentary section that is interfingered with and overlain by voluminous tholeiitic basalts, which are in turn interfingered with and overlain by another marine sedimentary section (Fig. 8). Our interpretation of this stratigraphy is that it represents initial subsidence of a basin or series of basins due to the onset of extensional rifting along the continental margin of North America at about 60 Ma (Wells and others, 1984; Brandon and Massey, 1985; Snavely, 1987). Continuation of rifting eventually induced convective upwelling of the underlying mantle and generation of voluminous basaltic magma by decompression partial

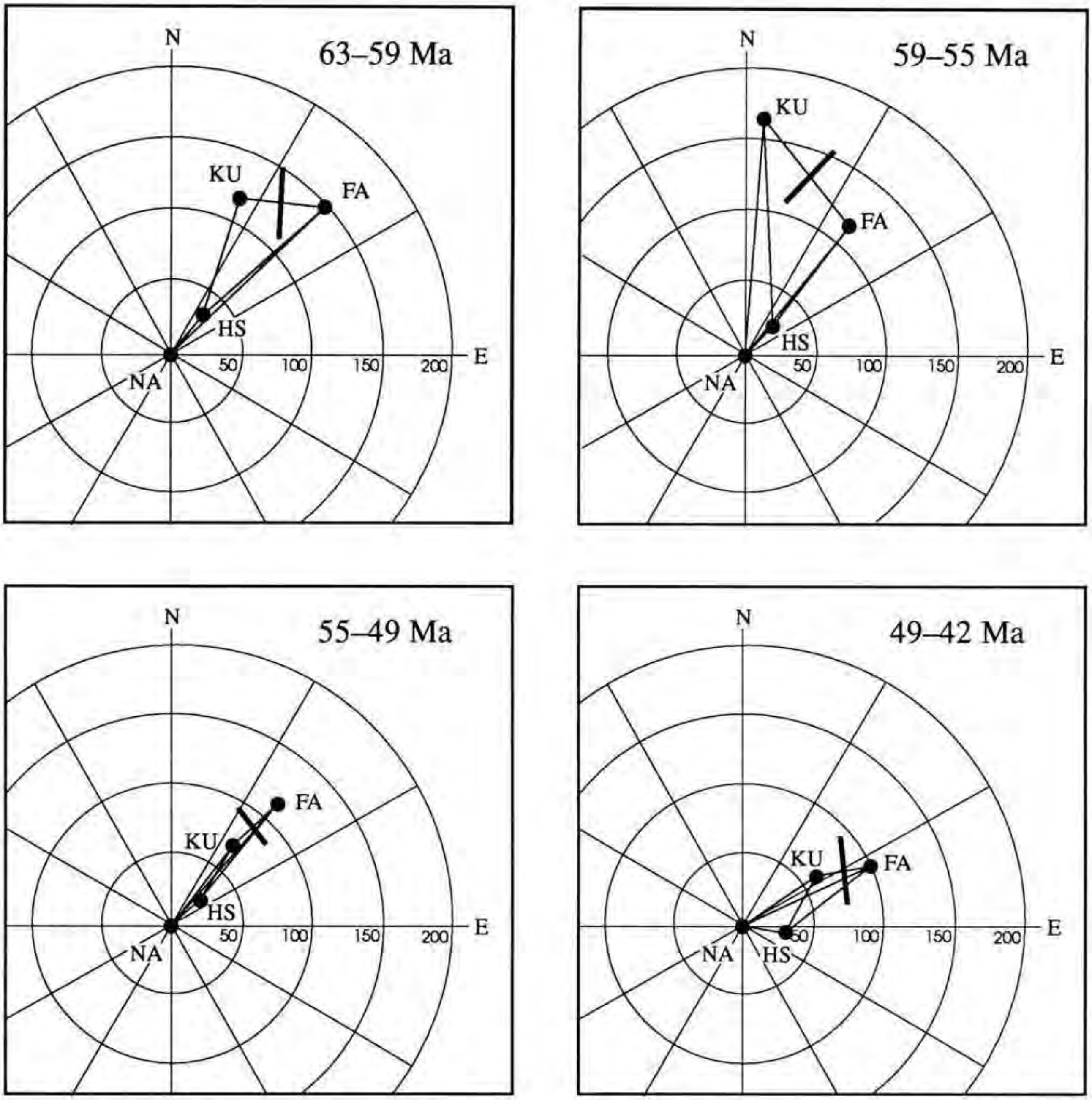

Figure 7. Relative plate motions along the western margin of North America at latitude 48° and longitude 124°. Lines connecting dots indicate vectors of relative motion. Length of lines indicates velocity, which is indicated by concentric circles scaled in increments of 50 km/m.y. KU, Kula plate; FA, Farallon plate; NA, North American plate; HS, hot spot position. Bold line bisecting the Kula–Farallon vector shows the orientation of the ridge that would be subducted along the Kula–Farallon–North America triple junction. Directions: N, north; E, east. Plate interactions are shown over the time intervals indicated in the upper right hand corner of each diagram. For example, between 59 and 55 Ma, the Kula plate was moving almost due north with respect to cratonic North America at a velocity of about 170 km/m.y. At the same time, the Kula plate was moving northwest at a velocity of about 100 km/m.y. relative to the Farallon plate and the Kula–Farallon ridge had a northeast trend. Plate motions were calculated using a hotspot reference frame in a manner similar to that of Engebretson and others (1985).

melting. As the rate of extension diminished, volcanism ceased and sedimentation in a thermally subsiding basin again prevailed. However, as indicated by Johnson and Yount (1992), thermal subsidence alone cannot account for the thickness of the post-volcanic sedimentary section in the northern Olympics. In the Tofino–Fuca basin, it appears that tectonic subsidence was related to the underthrusting of Crescent basalts beneath southern Vancouver Island along the Leech River fault (Clowes and others, 1987). The effects of tectonic activity are first recorded in Aldwell deposition by the presence of metamorphic clasts from the Leech River and Pandora Peak units, presumably

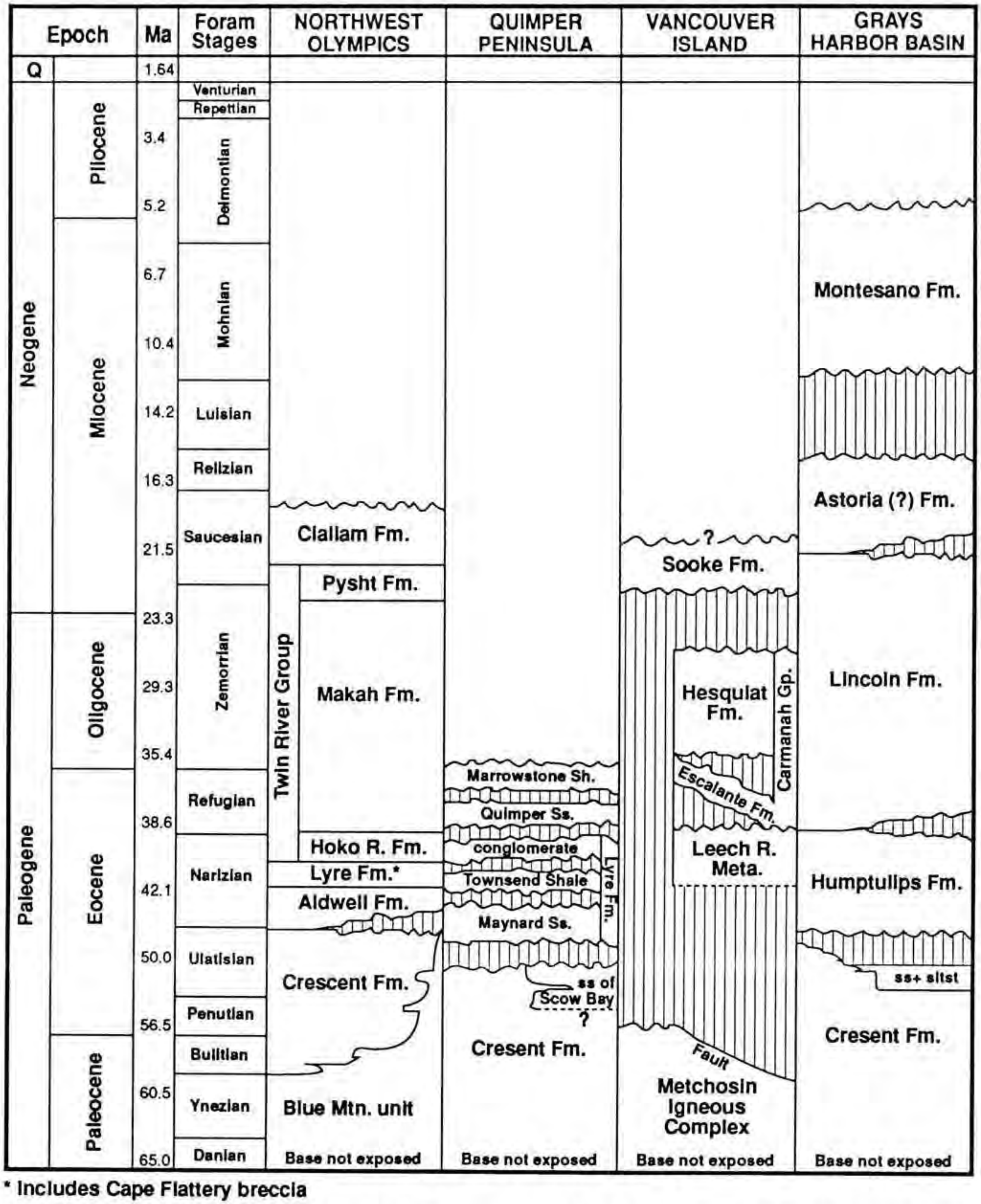

Figure 8. Stratigraphic columns for four sections in the Crescent "terrane". Data for the northwest Olympics from Tabor and Cady (1978), Snavely and others (1978, 1983), Snavely (1987), De Chant (1989), Palmer and Lingley (1989), Babcock and others (1992). Data for the Quimper Peninsula from Armentrout and Berta (1977), Melim (1984), Rauch (1985), and Whetten and others (1988). Data for Vancouver Island from Cameron (1971, 1980), Jeletsky (1975), Muller (1977), Fairchild and Cowan (1982), and Bream (1987). Data for the Grays Harbor Basin from Rau (1981), Armentrout (1987), Bigelow (1987), and Palmer and Lingley (1989). Time scale from Harland and others (1989).

derived from Vancouver Island thrust sheets. Active subsidence continued until deposition of the Pysht and Sooke Formations, which record a dominant shallowing-upward trend. Provenance studies indicate that sediments of the Tofino–Fuca basin were probably derived entirely from Vancouver Island or Coast Plutonic Complex sources until deposition of the Clallam Formation. Apparently, the basin was cut off from the Cascades and Kootenay sources that were supplying basins of the Chuckanut, Puget Group, and Olympic subduction complex. (See Heller and others, 1992.) Clast compositions in the Grays Harbor basin record uplift of both the Cascade Range during Astoria(?) deposition and of the Olympic subduction complex during Montesano deposition.

A widely accepted alternative hypothesis is that the tholeiitic volcanic rocks of the Oregon–Washington Coast Range represent a seamount chain developed on the Kula or Farallon plate and accreted to the margin of North America at about 50 Ma (Duncan, 1982; Armentrout, 1987; Heller and others, 1987). Given the thick section of terrigenous Blue Mountain unit sediments that underlie and interfinger with the Crescent volcanic rocks, it is difficult to accept this origin for the Crescent "terrane", although it may apply to the Siletz River terrane. If an accretion event did occur in the Crescent "terrane", it must have happened before deposition of the post-volcanic sedimentary section, which is in part interbedded with the underlying basalts. Snavely (1987) argues that lower and middle Eocene basalts plus interbedded sediments were deformed against the continental margin at about 52 Ma. We find no consistent evidence for this deformation, although this would be at about the time of initiation of subaerial volcanism in the Dosewallips section. Another compelling argument against an accreted seamount origin for the Crescent/Metchosin volcanics is the range in ^{40}Ar-^{39}Ar ages from about 45 Ma to 57 Ma with no apparent geographic progression. Hotspot-related volcanism with a lifetime of 12 m.y. would be quite unlikely given the 100+ km/m.y. relative velocities of the Kula and Farallon plates during the Eocene. Another feature inconsistent with seamount accretion is the apparent coeval extrusion of MORB-like Grays River volcanics with arc-like Northcraft volcanics (Phillips and others, 1989; Babcock and others, 1992). Although a rift origin implies in-situ development, it is possible—and in fact likely—that there has been as much as a few hundred kilometers of northward displacement of the Crescent "terrane" along a fault or faults in the Puget Lowland (Johnson, 1984, 1985).

We believe that plate kinematic relations can explain the development of the rift regime that we describe for rocks of the Crescent "terrane". The onset of rifting could have been related to either an increasingly oblique angle of subduction or the northward passage of the triple junction. For the latter, we assume that ridge subduction generated a window in the slab of oceanic crust descending beneath the margin of North America. This allowed upwelling of aesthenospheric mantle, rifting, and basalt generation. (See Dickinson and Snyder, 1979; Thorkelson and Taylor, 1989.) The 30–35-degree difference in the convergence vectors between the Kula and Farallon plates also could

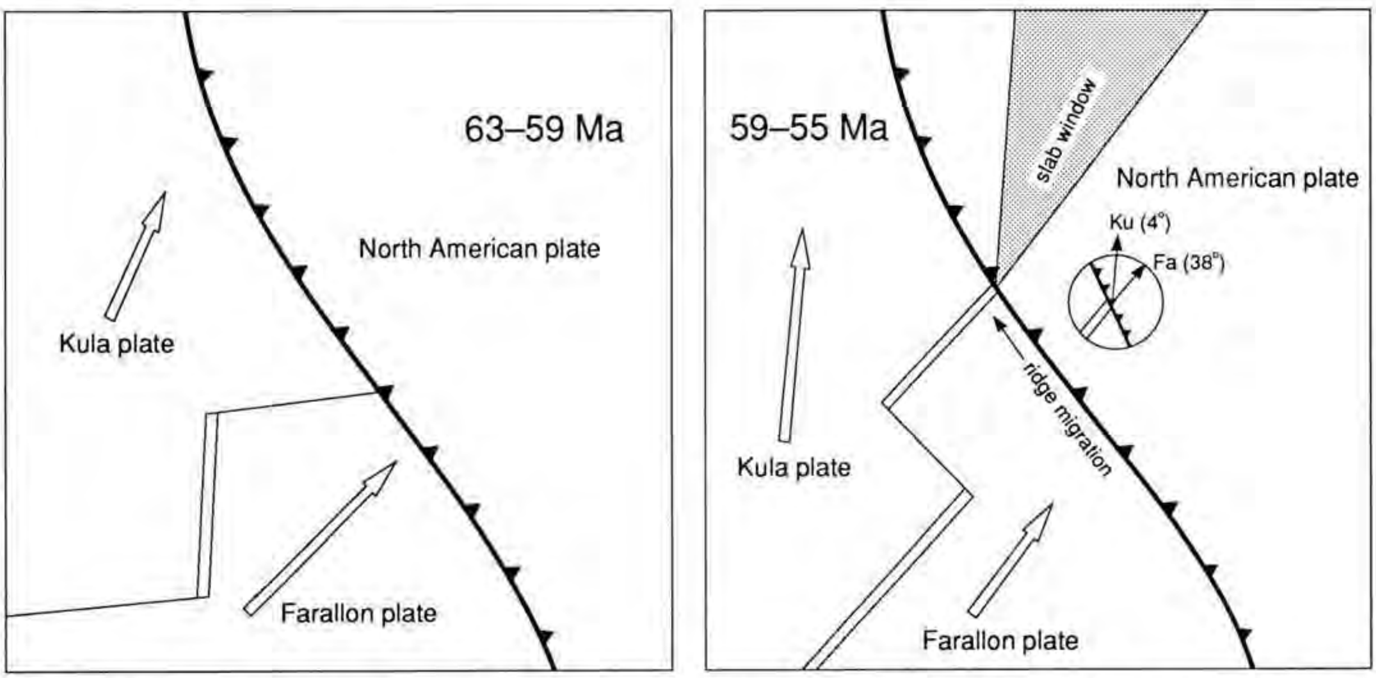

Figure 9. Inferred plate tectonic setting of the Oregon–Washington Coast Ranges during the early to middle Eocene. Note the inferred proximity of the Kula–Farallon–North American triple junction and the increase in the velocity and obliquity of Kula plate subduction during the time interval 59–55 Ma when Crescent magmatism was initiated. Note also the difference in slab traction vectors and the possible generation of a slab window along the path of the subducted Kula–Farallon ridge; these factors could induce extension in the upper plate.

have produced a rift-inducing differential slab traction effect along the triple junction (Fig. 9).

The role of a hot spot (Yellowstone or otherwise) is difficult to assess. White and Mackenzie (1989) show that a mantle plume can affect an area with a diameter of as much as 2,000 km. This would place the Crescent well within the range of the reconstructed path of the Yellowstone hotspot as it approached the margin of North America at a point that "best fit" calculations place near southern Oregon or northern California during the early Eocene. (However, note that the possible errors in reconstruction allow the hotspot track to be as far north as Vancouver Island.) The observed "enrichment trend" (Fig. 3) of the Crescent basalts toward the south would be consistent with an increasing contribution of a primitive mantle source—an expected influence of a mantle plume. Indeed, Pyle and Duncan (1992) observe that the Roseburg volcanic rocks in southern Oregon, which were extruded during the Eocene close to the reconstructed position of the Yellowstone hot spot, have a trace element composition and ^{3}He-^{4}He ratio (13:1) similar to that of oceanic islands centered over a plume (for example, Hawaii). The volume of basalt extruded is certainly comparable to other plume-related large igneous provinces, such as volcanic passive margins and continental flood basalts. Duncan (1982) estimates a volume of 250,000 km^3 for the Coast Range volcanic rocks as a whole. The volume of basalt in the Crescent "terrane" is at least 100,000 km^3, assuming an original area of 20,000 km^2 and an average thickness of 5,000 m.

Given the composition and available ages of the basalts analyzed in this study, the designation of a subaerial upper member and a submarine lower member of the Crescent basalt is meaningless on a regional scale. There is no correlation between measured ages of the basalts and their submarine versus subaerial nature, nor is there a correlation with chemical composition. Instead, it is apparent that there were several different centers of extrusion that developed in a series of subsiding basins. Overlap and poor exposure makes the boundaries of the basins and provenance of flows difficult to discern. The northern and central sections of Crescent basalt may represent at least two separate extrusive regimes within one basin. Between the two centers (at the northeast bend of the "horseshoe" near Deer Park) there was continuous sediment deposition with no intervening volcanic rocks, so that the upper sedimentary section lies directly on the Blue Mountain unit. Many of the extrusive centers became emergent islands where submarine basalts were succeeded by subaerial basalts, but some sections (for example, at Hurricane Ridge and Crescent Lake) consist entirely of submarine flows.

CONCLUSIONS

In summary, our interpretation is that the Crescent "terrane" is a package of sedimentary and volcanic rocks deposited during a period of extension along the continental margin of North America from the early Eocene to the mid-Miocene. In this sense the Crescent "terrane" is not accretionary, although the Crescent basalts and Blue Mountain unit may have moved northward as much as a few hundred kilometers and some of the basalts have clearly been thrust under the continental margin. These rocks record the interaction between the Kula, Farallon, and Pacific plates and the margin of North America, which involved varying degrees of oblique subduction and triple junction migration during the evolution of the Crescent "terrane". Rifting began during the early Eocene with subsidence of a marginal marine basin in which terrigenous sediments of the Blue Mountain unit were deposited from a nearby continental source.

Continued lithospheric extension induced mantle upwelling with decompression partial melting and the generation of voluminous tholeiitic basalts at several extrusive centers in the Oregon–Washington coastal ranges. There is a southward enrichment trend in the basalt chemistry that could be the signature of a mantle plume, possibly the Yellowstone hotspot, which has a reconstructed position just off the Oregon Coast during the early Eocene. However, the tectonic environment of oblique convergence and the subduction of the Kula–Farallon ridge could have been sufficient to induce large-scale rifting along the continental margin. Radiometric dating, which is consistent with fossil assemblage ages, indicates that Crescent volcanism continued from about 57 Ma to at least 45 Ma in several separate, but partly overlapping basins of deposition. With the data available, we cannot establish whether rifting and extrusion occurred on an oceanic plate or in the forearc, but we prefer the latter hypothesis because there is no geological or geophysical evidence that the subduction zone stepped outward after an accretion event.

As the rate of extension and volcanic activity diminished, thermal subsidence and sediment deposition became dominant. Along the northern Olympic coast the Tofino–Fuca basin was tectonically deepened, probably by thrusting of the Crescent volcanic rocks beneath Vancouver Island, during deposition of the Aldwell through the Makah (ca. 45–23 Ma). Finally, from about 23 Ma to 20 Ma, deposition of the Pysht, Clallam, and Sooke Formation sediments rapidly filled the remnant of the basin. On the southern margin of the Crescent "terrane", shallow-marine deposition continued into the Pliocene (Montesano Formation), and sediment composition records the Miocene uplift of first the Cascades Range and then the Olympic subduction complex.

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track of plate motions, the Kula–Farallon–North American triple junction, and the Yellowstone hotspot.

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Cooling tower of an unfinished nuclear power plant at Satsop in Grays Harbor County. Recent advances in our understanding of the tectonics and earthquake hazards of the Cascadia Subduction Zone were due at least partially to investigations prompted by (and funded because of) safety concerns at this facility. Photo by Timothy J. Walsh, 1988.

Late Cenozoic Structure and Stratigraphy of South-Central Washington

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ABSTRACT

The structural framework of the Columbia Basin began developing before Columbia River Basalt Group (CRBG) volcanism. Prior to 17.5 Ma, the eastern part of the basin was a relatively stable area, with a basement of Paleozoic and older crystalline rock. The western part was an area of subsidence in which large volumes of sediment and volcanic rocks accumulated. The boundary between the two parts is a suture zone between the stable craton and accreted terranes.

Concurrent with eruption of the CRBG, anticlinal ridges of the Yakima Fold Belt (YFB) were growing under north-south compression. Topographic expression of these features, however, was later masked by the large volume of CRBG basalt flowing west from fissures in the eastern Columbia Basin. The folds continued to develop after cessation of volcanism (ca. 6 Ma), leading to as much as 1,000 m of structural relief in the past 10 million years. Rates of subsidence and fold growth in the Columbia Basin decreased from the Miocene to Holocene. Shortening across the entire YFB probably does not exceed 25 km.

Post-CRBG evolution of the Columbia Basin is recorded principally in folding and faulting in the YFB and sediments deposited in the basins. The accompanying tectonism resulted in lateral migration of major depositional systems into subsiding structural lows.

Although known late Cenozoic faults are on anticlinal ridges, earthquake focal mechanisms and contemporary strain measurements indicate most stress release is occurring in the synclinal areas under north-south compression. There is no obvious correlation between focal mechanisms for earthquakes whose foci are in the CRBG and the location of known faults.

High *in situ* stress values help to explain the occurrence of microseismicity in the Columbia Basin but not the pattern. Microseismicity appears to occur in unaltered fresh basalt. Faulted basalt associated with the YFB is highly brecciated and commonly altered to clay. The high stress, abundance of ground water in confined aquifers of the CRBG, and altered basalt in fault zones suggest that the frontal faults on the anticlinal ridges probably have some aseismic deformation.

INTRODUCTION

The Columbia Basin is an intermontane basin between the Cascade Range and the Rocky Mountains that is filled by Cenozoic volcanic rocks and sediments. This basin forms the northern part of both the Columbia Plateau physiographic province (Fenneman, 1931) and the Columbia River flood-basalt province (Reidel and Hooper, 1989).

Within the Columbia Basin are four structural subdivisions or subprovinces (Fig. 1). The Yakima Fold Belt (YFB) is a series of anticlinal ridges and synclinal valleys in the western and central parts of the basin. Structural trends range from northwest to northeast but are predominantly east-west. The Palouse Slope forms the eastern subprovince of the basin; it shows little deformation, with only a few faults and low-amplitude, long-wavelength folds on an otherwise gently westward-dipping paleoslope (Swanson and others, 1980). A third subprovince, the Blue Mountains, forms the southern border of the Columbia Basin. It is a northeast-trending anticlinorium that extends 250 km from the Oregon Cascades to the eastern part of the Columbia Basin. The easternmost part of the basin constitutes the fourth subprovince and will not be discussed in this paper.

In the central and western parts of the Columbia Basin, the Columbia River Basalt Group (CRBG) overlies Tertiary continental sedimentary rocks and is overlain by younger Tertiary and Quaternary fluvial and glaciofluvial deposits (Campbell, 1989; Reidel and others, 1989b; Smith and others, 1989; U.S. Department of Energy, 1988). In the eastern part, a thin (<100 m) sedimentary unit separates the basalt and underlying crystalline basement and a thin (<50 m) veneer of eolian sediments overlies the basalt (Reidel and others, 1989b).

In this paper, we discuss the Late Cenozoic structure and stratigraphy of the Columbia Basin. We focus largely on the period following eruption of the last voluminous flood-basalt flows of the Saddle Mountains Basalt, those of the Elephant Mountain Member (10.5 Ma). However, we trace the development of the Columbia Basin from before the CRBG eruptions to the present because tectonic events that occurred during this time are closely related.

STRATIGRAPHY

The generalized stratigraphy of the Columbia Basin is shown in Figure 2 and summarized in Table 1. Most of the rocks exposed in the basin are the CRBG, intercalated sedimentary rocks of the Ellensburg Formation, and younger sedimentary rocks that include the Ringold Formation, Snipes Mountain conglomerate, Thorp Gravel (not shown

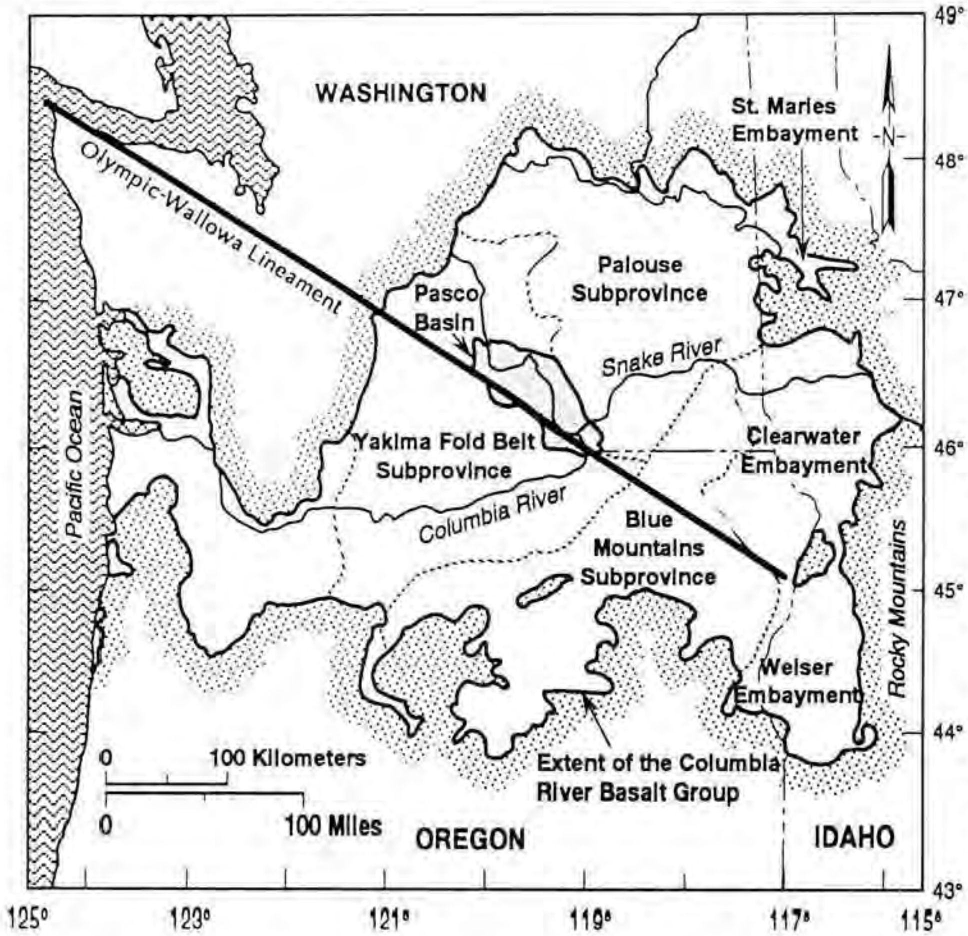


Figure 1. The Columbia Basin. Shown are the areal extent of the Columbia River Basalt Group, the major structural-tectonic subprovinces, the Pasco Basin, and the Olympic–Wallowa lineament.

on Fig. 2), the Hanford formation, and other localized strata. In this report, the Hanford formation includes all deposits of cataclysmic Pleistocene floods, including those from glacial Lake Missoula and the Columbia River system.

Stratigraphy Older than the Columbia River Basalt Group

Rocks older than the CRBG are exposed along the margin of the Columbia Basin. Their stratigraphy is complex and varies widely in both age and lithology (Campbell, 1989). Along the northwest margin, a series of sedimentary basins formed in early Tertiary time (Tabor and others, 1984; Campbell, 1989). These basins are now separated by tectonic "blocks" or uplifts that expose pre-Tertiary rocks and have a northwest-trending structural grain.

Along the northeast and east margins of the Columbia Basin, the CRBG laps onto Paleozoic rocks and Precambrian metasedimentary rocks interspersed with crystalline rocks. These include: Proterozoic metasedimentary rocks of the Windermere Group and Belt Supergroup, miogeosynclinal lower Paleozoic shallow marine rocks, rocks associated with the Kootenay arc, granitic rocks of the Idaho Batholith, and other Jurassic and Cretaceous intrusions (Stoffel and others, 1991). The structural grain of these rocks is north to northeast.

To the south and southwest, lower to middle Tertiary (Paleogene) volcanic rocks and related volcaniclastic rocks directly underlie the CRBG. The rocks include tuffs, lahars, and tuffaceous sedimentary rocks interbedded with rhyolite, andesite, and basalt flows and breccias; these are primarily assigned to the Clarno and John Day Formations. Older (Permian–Cretaceous) volcaniclastic sedimentary rocks and metasedimentary rocks of accreted intra-arc and volcanic-arc origin are exposed along the southeast margin of the CRBG (Walker and MacLeod, 1991).

To the west, younger volcanic rocks erupted from the High Cascades cover the CRBG and obscure the older rocks. Rare inliers of older Paleozoic rocks, such as the Rimrock inlier, are exposed as erosional windows in younger volcanic rocks.

Sedimentary rocks related to those along the northwest margin are thought to be present under the interior of the Columbia Basin. However, units exposed along the north, east, and south margins of the CRBG probably contributed to the sedimentary package because they were extensively eroded by west-flowing rivers (Fecht and others, 1987) and their deposits accumulated in the subsiding western part of the basin.

Columbia River Basalt Group and Ellensburg Formation

The CRBG consists of 174,000 km^3 of tholeiitic flood-basalt flows that were erupted between 17.5 and 6 Ma. It now covers approximately 164,000 km^2 of eastern Washington and Oregon and western Idaho (Tolan and others, 1989). Eruptive units have volumes as great as 5,000 km^3 (Reidel and others, 1989a; Reidel and Tolan, 1992). These flows are the structural framework of the Columbia Basin, and their distribution pattern reflects the tectonic history of the area (Reidel and others, 1989b).

Intercalated with, and in some places overlying, the CRBG are epiclastic and volcaniclastic sedimentary rocks of the Ellensburg Formation (Waters, 1961; Swanson and others, 1979a; Smith, 1988). Most volcaniclastic material occurs in the western basin; in the central and eastern basin, epiclastic deposits of the ancestral Clearwater and Columbia Rivers form the dominant lithologies (Fecht and others, 1982, 1987).

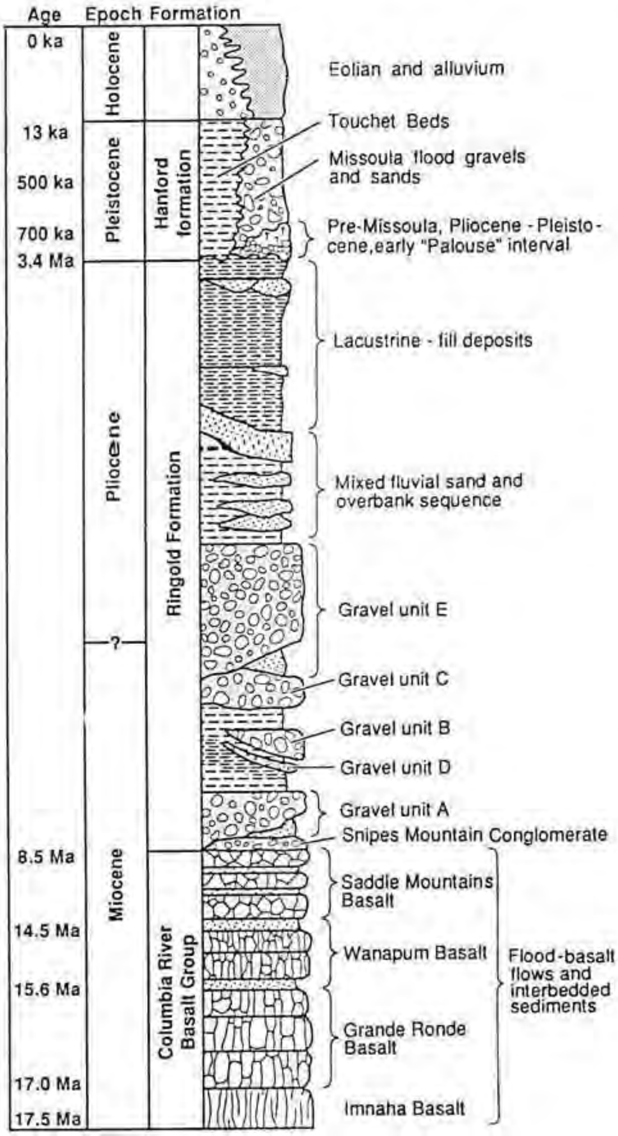


Figure 2. Diagrammatic representation of the Columbia River Basalt Group and younger sedimentary units of the central Columbia Basin. Emphasis is on the post-basalt rocks. Not to scale.

Post-Columbia River Basalt Stratigraphy

Most post-CRBG sediments are confined to the synclinal valleys of the YFB. The sedimentary record is incomplete, but it is a direct reflection of the structural development of the Columbia Basin (Fecht and others, 1987). The upper Miocene to middle Pliocene record of the Columbia River system in the Columbia Basin is represented by the upper Ellensburg Formation, Ringold Formation, and Snipes Mountain conglomerate. The Thorp Gravel (≈3.7 Ma) river terrace deposits record the post-CRBG history of the upper Yakima River (Waitt, 1979; Campbell, 1983). Except for local deposits (for example, the "Pliocene–Pleistocene unit" and the "early Palouse soil" [U.S. Department of Energy, 1988]), there is a hiatus in the stratigraphic record between the end of the Ringold (3.4 Ma) and Thorp Gravel (≈3.7 Ma) deposition and the Pleistocene (1.6 Ma) deposits. Pleistocene to Holocene sediments overlying the CRBG include: flood gravels and slackwater sediments of the Hanford formation; terrace gravels of the Columbia, Snake, and Yakima Rivers; and, in eastern Washington, eolian deposits including the Palouse Formation (Keroher and others, 1966).

GEOLOGIC STRUCTURES OF THE COLUMBIA BASIN

Structural Setting

The major structural features that underlie the Columbia Basin are exposed along its west and north margins (Fig. 3). Major structures along the western margin have been discussed by Campbell (1988, 1989) and Tabor and others (1984); those along the north margin are discussed in the compilation by Stoffel and others (1991). All these features are older than the CRBG, and it is not clear how many of them extend under the CRBG. Those structural features that we consider important to understanding the Columbia Basin and are known to extend into or cross the Columbia Basin interior are summarized below.

The Olympic–Wallowa Lineament

The Olympic–Wallowa lineament (OWL) is a major topographic feature in Washington and Oregon that crosscuts the Columbia Basin (Raisz, 1945; Fig. 1). This feature parallels prebasalt structural trends along the northwest margin of the Columbia Basin, but it has not been linked to any individual structure (Campbell, 1989; Reidel and Campbell, 1989). Within the YFB, the OWL includes a zone of Miocene and post-Miocene deformation along Manastash Ridge and apparent bending of Umtanum Ridge, Yakima Ridge, and Rattlesnake Ridge (the ridge immediately south of Yakima Ridge) (Fig. 3).

The portion of the OWL that crosses the Columbia Basin is called the Cle Elum–Wallula deformed zone (CLEW; Kienle and others, 1979). It is a 10-km-wide, moderately diffuse zone of anticlines that have a N50°W orientation (Fig. 3). As defined by G. A. Davis (*in* Washington Public Power Supply System, 1981a), the CLEW consists of three structural parts: (1) a broad zone of deflected or anomalous fold and fault trends extending south from the northwest part of Manastash Ridge (in the Cascade foothills near the edge of the CRBG) to Rattlesnake Mountain; (2) a narrow belt of topographically aligned domes and doubly plunging anticlines extending from Rattlesnake Mountain to Wallula Gap, where the OWL crosses the Columbia River (a zone commonly called the Rattlesnake–Wallula alignment or RAW); and (3) the Wallula fault zone, extending from Wallula Gap southeast to the Blue Mountains.

Northwest of the CRBG margin, numerous northwest- and north-trending faults and shear zones of the Straight Creek fault system lie subparallel to the OWL (Tabor and others, 1984). The Snoqualmie batholith intrudes these faults but is not cut by them, indicating that any movement

Table 1. Stratigraphic units and characteristics. Palouse Formation, Latah Formation, and other loess deposits not included

Unit	Age	Thickness	Distribution	Lithology	Stratigraphic trends	Tectonic implications
Columbia River Basalt Group	17.5 Ma–6 Ma	as much as 4 km in central Columbia Basin	Extent defines the Columbia Plateau	Tholeiitic flood-basalt flows	Older, more voluminous units cover entire area; younger, thinner units occur in eastern and central Columbia Basin	Records stability in the eastern Columbia Basin, subsidence in the central and western basin, and Miocene uplift of anticlinal ridges
Upper Ellensburg Formation (Fecht and others, 1987; Smith, 1988)	10 Ma–4.72 ±0.28 Ma	as much as 350 m in Yakima Basin	Nile, Selah, Yakima, Kittitas, and Satus and Toppenish basins; also in Goldendale area	Basalt side-stream gravels, lahars, and volcaniclastic sediments derived from Cascade Range, siliciclastics deposited by Columbia River	Volcaniclastic sediments most common in west; siliciclastic sediments in east; gravels record channel system positions	Shifting channel deposits reflect displacement of river courses caused by ridge uplift and basin subsidence, especially Yakima River and Columbia River southwest of Pasco Basin
Snipes Mountain conglomerate (Schminke, 1964; Fecht and others, 1987; Smith, 1988)	<8.5 Ma–8.5 Ma	30–150 m	Lower Yakima valley and across western Horse Heaven Hills; also eastern Toppenish basin	Quartzose gravel and siliciclastic sands; interbedded volcaniclastic sediments common in eastern Toppenish basin	Linear channel tracts from Sunnyside Gap up Moxee and Yakima valleys and then across Horse Heaven Hills to Goldendale area	Records Columbia River course prior to diversion into Pasco Basin
Ringold Formation (Fecht and others, 1987; Lindsey, 1991)	<8.5 Ma–>3.4 Ma	as much as 185 m	Pasco Basin, north side of Saddle Mountains, and Walla Walla basin	Fluvial gravels and sands, overbank deposits, lacustrine deposits, and alluvial fan deposits	Gravelly alluvial tracts mixed with basin-wide overbank systems dominate lower part of section; sharply overlain by a sandy alluvial system that, in turn, grades up-section into cyclic lacustrine deposits	Records initial post-CRBG Columbia River deposits in Pasco Basin; shifting channel courses reflect syndepositional uplift on basin margins; large lakes reflect regional changes in river gradients
Lake deposits overlying Snipes Mountain conglomerate (Smith, 1988)	<4.7 ±0.3 Ma	more than 30 m	Lower Yakima valley near Sunnyside and Granger and on Ahtanum Ridge	Laminated silts and fine-grained sandstone	Laterally correlative with upper Ringold lake deposits(?)	Large lakes form as a result of regional gradient changes
Thorp Gravel (Waitt, 1979; Bentley and Campbell, 1983; Campbell, 1983; Fecht and others, 1987; Smith, 1988)	<3.64 ±0.74 Ma–3.7 ±0.2 Ma	as much as 200 m	Kittitas and Selah basins, and south into Yakima basin as far as Toppenish	Basaltic side-stream gravels and polymictic mainstream gravels deposited as an alluvial wedge off the Cascades	Old terraces of the Yakima River	Record uplift and erosion of Cascades and Yakima folds.
Pliocene and Pleistocene strata (U.S. Dept. of Energy, 1988; Baker and others, 1991)	<3.5 Ma–~1 Ma	as much as 10 m	Regional distribution in basins and on uplifts	Pedogenic carbonates, basaltic alluvium, eolian deposits, multilithologic quartzose gravels	Unconformably overlies middle Pliocene and older (>3.5 Ma) strata; discontinuous horizons in and around basins and uplifted on ridges	Deposited after post-3.5 Ma base-level change that led to regional incision of main rivers; records more recent uplift of anticlinal ridges
Hanford formation (Fecht and others, 1987; Baker and others, 1991)	<1 Ma–~12 ka	as much as 70 m	Pasco Basin, Toppenish, Yakima, and Walla Walla basins, and flood deposits throughout the Columbia Basin	Pebble to boulder gravel, sand, and laminated silts deposited by cataclysmic flood waters released from glacial Lake Missoula	Common throughout region, mostly below elevations of approximately 1200 ft	Strata locally offset by faults recording Pleistocene deformation
Quaternary alluvium (Baker and others, 1991)	<2 Ma	0 to 75 m	Regional	Locally derived alluvial and colluvial deposits	Laterally discontinuous strata common on basin margins and on uplifted ridges	Folded and faulted deposits record neotectonic deformation in region

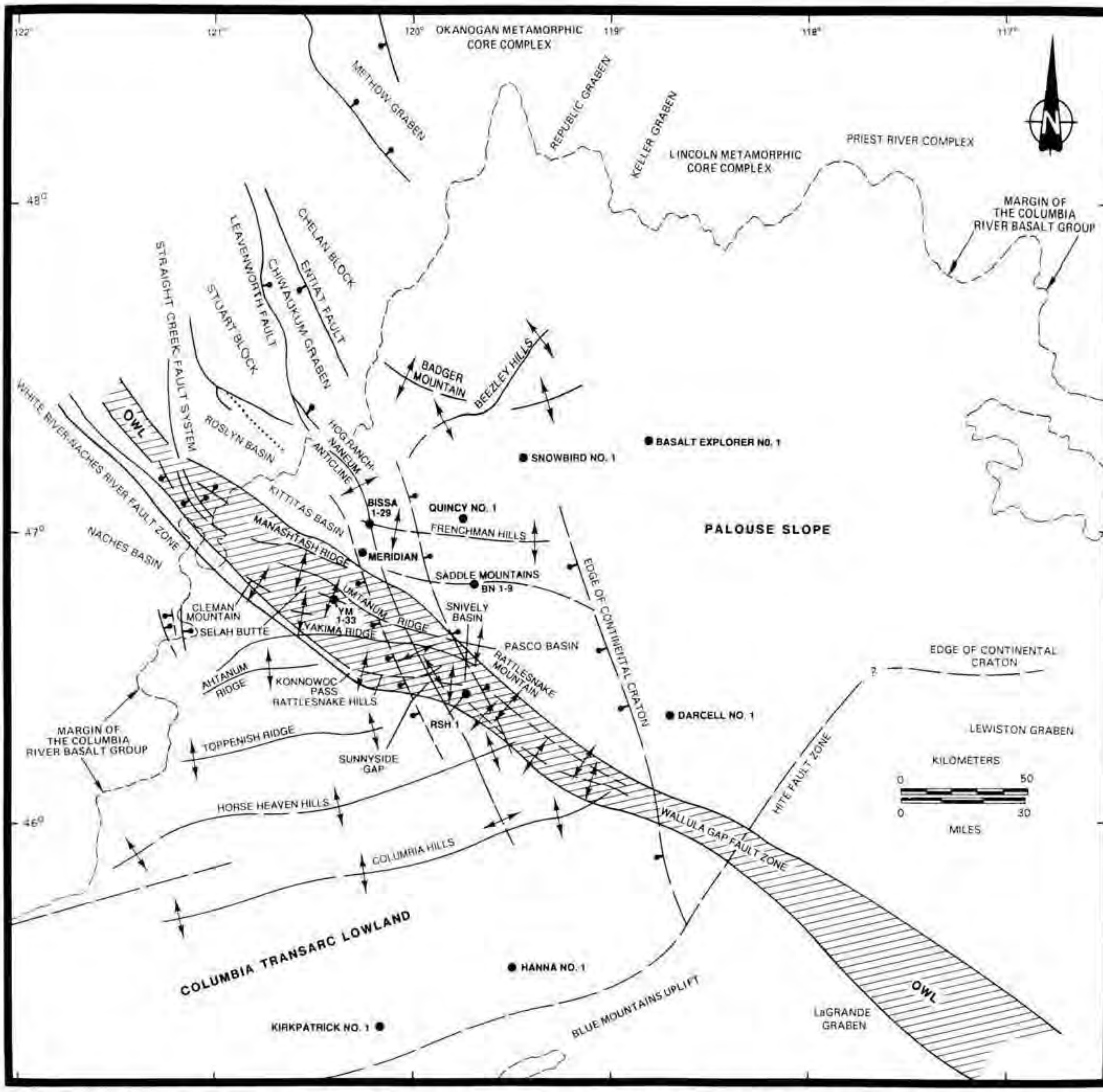


Figure 3. Major geologic features in the Columbia Basin. Shown are the Hog Ranch–Naneum Ridge anticline, the Olympic–Wallowa lineament (OWL), and major surface structures. Also shown are the inferred locations of major subbasalt structural features and the proposed edge of the continental craton.

along the OWL at the western margin of the Columbia Basin must be older than the batholith, 17 to 19.7 Ma (Frizzell and others, 1984).

The structural significance of the OWL has been called into question by two recent geophysical studies. Neither a seismic profiling survey by Jarchow (1991) nor a gravity survey by Saltus (1991) could find any obvious geophysical signature of the OWL below the CRBG.

Hog Ranch–Naneum Ridge Anticline

The Hog Ranch–Naneum Ridge (HR–NR) anticline is a broad south-trending anticline in the CRBG that crosses the YFB at a nearly right angle. The anticline begins at the north basalt margin on Naneum Ridge (at the south end of the Stuart block and southwest of the city of Wenatchee), trends southeast for about 12 km and then turns south toward Prosser, WA, where it separates the Toppenish basin to the west from the Pasco Basin. This south-plunging structure passes through five Yakima folds and the OWL. A gravity gradient and a series of gravity highs delineate part of the anticline in the subsurface. The southern extension of the anticline appears to be a Bouguer gravity high near the Washington–Oregon border southeast of Prosser.

The HR–NR anticline was active in middle to late Miocene time, as demonstrated by thinning of basalt flows across it (Reidel and others, 1989b). However, the east-

Table 2. Characteristics of anticlinal ridges; $\bar{x}$ = mean; σ = 1 standard deviation; R = range; NK = not known (best approximation). * structural relief on youngest basalt

Anticline	Length/ width	Relief* (maximum)	Trend	Number of segments	Segment length	Vergence	Amount of shortening	Geometry
Beezley Hills–Beezley Coulee	160 km/ 5 km	430 m	230°	4	$\bar{x}$=40 σ=20 R=30–70	S	NK (<2 km)	Asymmetrical, monoclinal
Badger Hills–Moses Stool	50 km/ 5–15 km	300 m	220°	1	---	N	NK (<2 km)	Asymmetrical, monoclinal
Frenchman Hills	100 km/ 5–10 km	200 m	90°–100°	7	$\bar{x}$=14 σ=10 R=7–35	N	NK (<2 km)	Asymmetrical, gentle to open
Saddle Mountains	110 km/ 5–10 km	550 m	90°–115°	6	$\bar{x}$=14 σ=10 R=5–20	N	>3 km	Asymmetrical, gentle to open, box fold
Manastash Ridge–Thrall structure	55 km/ 5–10 km	370 m	120°	4	$\bar{x}$=12 σ=2.5 R=10–15	N	NK (>3 km)	Asymmetrical, gentle to open
Umtanum Ridge	110 km/ 3–10 km	520 m	90°–130°	9	$\bar{x}$=11 σ=4.2 R=5–17	N	1–3 km	Asymmetrical, tight to open, en echelon segments east end
Cleman Mountain	35 km/ 8 km	950 m	130°	2	$\bar{x}$=18 σ=8 R=13–23	S	NK (>1 km)	Asymmetrical
Yakima Ridge	100 km/ 5–10 km	550 m	135°–225°	12	$\bar{x}$=12 σ=8 R=5–30	N	NK (>3 km)	Asymmetrical, gentle to open, en echelon segments, box fold segments
Rattlesnake Mountain and "rattles"	85 km/ 5–20 km	800 m	310°	11	$\bar{x}$=9 σ=6 R=5–25	N	NK (>3 km)	Asymmetrical, tight to open, faulted out hinge, doubly plunging
Rattlesnake–Ahtanum Ridge	100 km/ 5–8 km	610 m	238°–315°	11	$\bar{x}$=9 σ=4 R=5–18	N	NK (>1 km)	Asymmetrical, gentle to open
Toppenish Ridge	85 km/ 4–8 km	500 m	118°–258°	5	$\bar{x}$=17 σ=7 R=10–28	N	NK (>1 km)	Asymmetrical, tight to open
Snipes Mountain	13 km/ 1 km	150 m	110°	3	13 km	S	NK (<1 km)	Asymmetrical, tight to open
Horse Heaven Hills	185 km/ 5–30 km E; 2–7 km W	335–1100 m	115°–255°	21	$\bar{x}$=17 σ=5 R=5–20	N	>2 km (0.67–1.25 km)	Asymmetrical, tight to open, en echelon subsidiary crest folds, box folds
Columbia Hills	170 km/ 5–10 km	250–365 m	255°	10	$\bar{x}$=15 σ=6 R=6–23	S	NK (>2 km)	Asymmetrical, tight to open doubly plunging, en echelon subsidiary crest folds, box folds

trending Yakima folds show no apparent offset by the cross structure (Campbell, 1989; Tabor and others, 1982; Kienle and others, 1979; Reidel and others, 1989b), nor is the HR–NR anticline offset where the OWL–CLEW crosses it. Growth of the HR–NR anticline continued from the Miocene to the present time and is now delineated by the highest structural points along the ridges that cross it.

White River–Naches River Fault Zone

The White River–Naches River Fault Zone (WR–NRFZ) (Fig. 3), a major fault zone that extends 90 km from Naches to Enumclaw, WA, separates two domains of dissimilar structure, stratigraphy, and topography (Campbell, 1988, 1989). To the northeast, structures strike N60°W; to the southwest, structures in pre-Tertiary rocks trend N5°E to N20°W. The WR–NRFZ probably extends under the basalt at least as far as Konnowoc Pass (Fig. 3) and either parallels or crosscuts the HR–NR anticline. The WR–NRFZ is the major structure trending into the Columbia Basin that can be demonstrated to be a fundamental structural boundary in rock below the CRBG.

Leavenworth Fault Zone

The Leavenworth fault is part of a fault zone that consists of a series of northwest-trending high-angle faults and associated tight folds that marks the southwest side of the Chiwaukum graben (Fig. 3). At the basalt margin, the fault passes southeast under the CRBG in alignment with the HR–NR anticline. The Leavenworth fault is assumed to continue under the basalt along the HR–NR anticline and to have been a factor in its development (Campbell, 1989). Saltus (1991), however, suggests it does not continue underneath the basalt.

Table 3. Characteristics of major faults; NK = not known (best approximation)

Fault zone	Length	Trend	Horizontal offset	Vertical offset	Sense of movement; fault-plane dip	Age of last movement
CLEW	290 km	310°	0–4 km	0–800 m	mainly reverse	Quaternary; see Table 4
RAW (includes Wallula Fault Zone)	125 km	310°	0–4 km	0–800 m	mainly reverse, some possible strip-slip	Quaternary; see Table 4
Hite Fault System	135 km	330°–335°	NK	NK (0–>300 m)	vertical, en echelon, and strike-slip	Holocene (1936 Milton-Freewater earthquake); see Table 4
Frenchman Hills	100+ km	90°–100°	>300 m	~200 m	reverse, thrust; >45°S	>500,000 yr
Saddle Mountains	100 km	90°–115°	>2.5 km	600 m	reverse; >60°S	<3.4 Ma, Pleistocene–Holocene?
Manastash–Hansen Creek	70 km	120°	<1 km	~300 m	reverse, thrust	>1 Ma to 3.4 Ma
Umtanum	110 km	90°–130°	>300 m	1500 m	reverse, thrust; 30–70°S	<13,000 yr
Cleman Mountain	20? km	130°	NK	~900 m	reverse, thrust; N	Unknown
Yakima Ridge	120+ km	225°–135°	NK	~500 m	reverse, thrust; S, locally N	<1 Ma
Rattlesnake–Ahtanum Ridge	100 km	238°–315°	NK	~800 m	reverse, thrust; S	>13,000 yr
Toppenish Ridge	65–90 km	118°–258°	NK	~500 m	reverse, thrust; S	Holocene; see Campbell and Bentley (1981)
Horse Heaven Hills	200+ km	255°–115°	NK	~335–1100 m	reverse, thrust; S	Unknown
Columbia Hills	160 km	255°	NK (1 km)	~365 m	reverse, thrust; 70°N	Unknown
Northwest-trending faults	40–120 km	320°	<100 m	<100 m	strike-slip, vertical (dip reversal)	Holocene

Other Structures Marginal to the Columbia Basin

Other major structures including the Entiat fault, Methow, Republic, and Keller grabens, and numerous faults associated with the Kootenay Arc appear to die out or their magnitude is significantly reduced before reaching the basalt margin. Therefore, we do not consider them significant tectonic elements in the Columbia Basin.

Northwest-Trending Wrench Faults

A series of northwest-trending, dextral strike-slip wrench faults is present in the CRBG west of 120° longitude and the HR–NR anticline (Newcomb, 1969, 1970; Shannon & Wilson, Inc., 1973; Bentley and others, 1980; Swanson and others, 1979a, 1981; Anderson, 1987). These are conjugate and en-echelon faults and genetically related en-echelon folds that have a mean strike of N40°W. Many can be traced for more than 100 km but do not extend beyond the CRBG. These wrench faults cross and offset several Yakima folds, but they do not appear to have more than 100 m total displacement (Anderson, 1987).

The Yakima Folds

The YFB subprovince covers about 14,000 km^2 of the western Columbia Basin (Fig. 1) and formed as basalt flows and intercalated sediments were folded and faulted under north-south-directed compression. The reader is referred to Tables 2 and 3 and the recent compilation of structural features for the Columbia Basin by Tolan and Reidel (1989), which updates plate II-20 of Myers, Price and others (1979).

Most of the present structural relief in the Columbia Basin has developed since about 10.5 Ma when the last massive outpouring of lava, the Elephant Mountain Member of the Saddle Mountains Basalt (Fig. 2), buried much of the central Columbia Basin. The main deformation is concentrated in the YFB; there is only minor deformation on the Palouse Slope.

The YFB consists of anticlinal ridges separated by synclinal valleys. The folds are typically segmented, and most have north vergence. However, some anticlines, such as Columbia Hills, Cleman Mountain, and a few segments of other ridges, have a south vergence. Fold length ranges from 1 to more than 100 km; fold wavelengths range from several kilometers to as much as 20 km (Table 2). The folds are segmented by crosscutting faults and folds (Reidel, 1984; Reidel and others, 1989b). Structural relief is typically less than 600 m but varies along the length of the fold. The greatest structural relief along the Frenchman Hills, the Saddle Mountains, Umtanum Ridge, and Yakima Ridge occurs where they intersect the north-trending HR–NR anticline (Fig. 3).

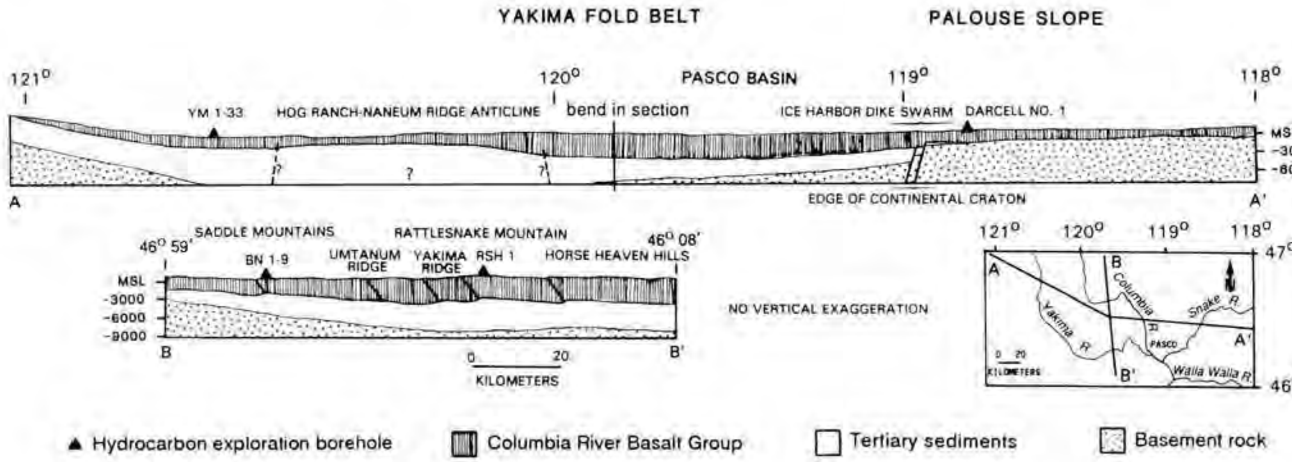


Figure 4. North-south and generally east-west cross sections through the central Columbia basin (modified from Reidel and others, 1989b).

Anticlines of the Yakima Fold Belt have various trends, ranging from N50°W through east-west to N50°E. Abrupt changes in trend occur along individual anticlines at segment boundaries, giving these anticlines a disjointed appearance along their length. This is most apparent where anticlines cross the zone of the CLEW. On a regional scale, parts of the fold belt typically have dominant directional trends. The anticlines lying within the CLEW zone typically trend N50°W (Fig. 3). However, anticlines southwest of the CLEW have a dominant N50°E trend (Swanson and others, 1979b; Reidel and others, 1989b; Tolan and Reidel, 1989) (Fig. 3), and the trend of anticlines northeast of the CLEW, such as the Saddle Mountains and Frenchman Hills, is dominantly east-west. Anticlines that extend across the entire fold belt, such as Umtanum and Yakima Ridges, have varied trends that reflect the dominant trends of the areas they cross.

Although the faults are rarely exposed, nearly all the steep forelimbs of the asymmetrical anticlines are faulted (Fig. 4). These frontal fault zones typically consist of imbricated thrusts (Bentley, 1977; Goff, 1981; R. D. Bentley *in* Swanson and others, 1979b; Hagood, 1986; Reidel, 1984, 1988; Anderson, 1987) that are emergent at ground surface. The tops of lava flows at ground surface become the planes of low-angle thrust faulting. Where erosion provides deeper exposures, these frontal faults are shown to be steep reverse faults: the fault in the Columbia water gap in the Frenchman Hills dips 45 degrees south (Grolier and Bingham, 1971), and the fault in the Columbia Hills at Rock Creek, WA, dips 50–70 degrees north (Swanson and others, 1979b).

Hydrocarbon exploration boreholes provide direct evidence for the dips of these frontal faults. Reidel and others (1989b) have shown that the Saddle Mountains fault must dip more than 60 degrees where the Shell-ARCO BN 1-9 borehole was drilled (Fig. 3). Drilling through the Umtanum fault near Priest Rapids Dam suggests that this fault dips at least 30 degrees (Puget Sound Power and Light Co., 1981) but perhaps as steeply as 60 degrees southward under the ridge (Price, 1982; Price and Watkinson, 1989).

Total shortening across the YFB is less than 25 km (Reidel and others, 1989b), or about 5 percent of the fold belt width. Crustal shortening is difficult to estimate, however. Typically, shortening caused by folding is 1 to 1.5 km, but the amount of shortening caused by faulting is generally unknown. Estimates for shortening due to faulting alone range from several hundreds of meters to as much as 4 km (Table 3).

Typical synclines are structurally low areas formed between the gently dipping limb of one anticline and the steeply dipping limb of another. The steep limb was thrust up on the gently dipping limb of the neighboring anticline to form the syncline. Few synclines were formed by simple synclinal folding of the basalt.

THE PRE-MIDDLE MIOCENE COLUMBIA BASIN

The Columbia Basin as we now see it reflects structural elements that formed prior to flood-basalt volcanism. This section of our paper provides an interpretation of the structural setting of the Columbia Basin prior to the eruption of the CRBG. Our interpretation is based on data from the last two decades of geophysical surveys and from deep hydrocarbon exploration boreholes in the Columbia Basin, both combined with our field observations.

The YFB and Palouse Slope overlie the two major pre-CRBG structural features of the Columbia Basin (Reidel and others, 1989b). The YFB lies on a large pre-basalt basin filled with as much as 7,000 m of continental sediments. The Palouse Slope subprovince overlies a crystalline basement high; only a thin (<100 m) discontinuous sediment package separates the basalt and basement.

The crystalline basement underlying the Palouse Slope (Fig. 4) has been penetrated by two boreholes (the Darcell and Basalt Explorer, Fig. 3). This basement is essentially metasedimentary rock composed of quartz, feldspar, and mica. It resembles the Addy Quartzite and certain Precambrian Belt Supergroup rocks. We interpret this basement to be part of the old continental craton that has remained fairly stable except along its west and south margins. Mohl and Theissen (1985) extended the southern boundary by extrapolating gravity data from Lewiston, ID, as far west as Pomeroy, WA (Mohl, 1985). The rock penetrated by the Darcell borehole (Fig. 3) indicates that the craton must also extend west and south from Pomeroy. We suggest that the trace of the southern cratonic margin is reflected at the surface by the Hite fault zone (Fig. 3), the major fault system that forms the western margin of the Blue Mountains between Pomeroy and Pendleton, OR, and has been the locus of many historic earthquakes (U.S. Department of Energy, 1988).

The YFB is underlain by a thick sequence of sediments that was first recognized in deep hydrocarbon exploration boreholes (Campbell and Banning, 1985). On the basis of seismic refraction survey data, Catchings and Mooney (1988) interpret this as a failed rift basin. Other studies using additional geophysical data sets (Rohay and Malone, 1983; Rohay and others, 1985; Glover, 1985; Zervas and Crosson, 1986), however, question this interpretation. These studies do not "see" any deep crust or mantle evidence for a failed rift basin.

The pre-CRBG basin above the crystalline basement is filled with deposits consisting primarily of Paleocene and Eocene continental sediments and some Oligocene volcaniclastic sediments (Campbell and Banning, 1985; Campbell, 1989). Reidel and others (1989b) interpreted the suture zone between the pre-CRBG basin and craton to coincide with the Ice Harbor dike swarm (Fig. 4). They also interpreted the location of the dike swarm to have been controlled by this fundamental crustal boundary. Reidel and others (1992) suggested that this boundary is the suture zone between the continental craton and accreted terranes. Furthermore, they suggested that the suture is oriented approximately N10°W and is marked at the surface by the Ice Harbor dikes and the boundary between the Saddle Gap and Eagle Lakes segments of the Saddle Mountains (Reidel, 1984, 1988). This location is also marked by the prominent aeromagnetic anomaly shown by Swanson and others (1979c) for the Ice Harbor dike swarm and that extends as far south as the east side of Wallula Gap. This boundary between craton and basin may cross the CLEW and intersect the structurally similar Hite fault boundary near Pendleton, OR.

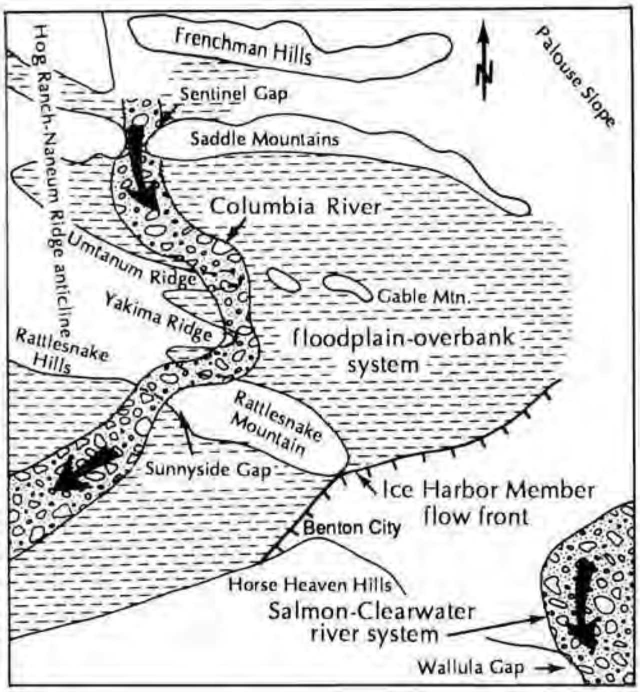
The HR–NR anticline parallels the eastern margin of the pre-CRBG basin and appears to divide it into two parts. Campbell and Banning (1985) have interpreted pre-CRBG rocks under the HR–NR anticline as part of a horst, but results from a recent seismic profile (Jarchow, 1991) suggest that the HR–NR anticline may not involve basement. We suggest that the Pasco Basin is underlain by a northwest-trending graben between the HR–NR anticline and the suture zone. The western graben fault may only involve the basalt and sediment and not the basement rock. Tertiary sedimentary and volcaniclastic rocks west of the HR–NR anticline extend beyond the margin of the CRBG and form part of the Cascade Range. The western basin margin is not faulted, as implied by the rift model of Catchings and Mooney (1988); the pre-CRBG sedimentary rocks continue across the present Cascade Range and were arched upward with the Cascade Range to form the present western boundary of the Columbia Basin. This model is supported by geophysical studies (Rohay and Malone, 1983; Rohay and others, 1985; Glover, 1985; Zervas and Crosson, 1986) suggesting that the Columbia Basin is not a failed rift basin but perhaps simply a back-arc basin.

The Blue Mountains uplift forms the southern boundary of the pre-CRBG basin, although the boundary is not precisely located. The boundary may lie near the present Blue Mountains crest because a thick sequence of sedimentary rock lies north of the crest below the basalt near the Washington–Oregon border (Fox and Reidel, 1987). We interpret this boundary to coincide with Beeson and others' (1989) east-trending Columbia Transarc lowland (Fig. 3) that forms the low area extending along the Washington–Oregon border through the Columbia Gorge. The Columbia Transarc lowland and the pre-CRBG basin combine to produce an apparent northeast-trending trough through the Columbia Basin. The thickest pre-CRBG sediment package lies along this trend; in addition, more CRBG flows are found along the trough than elsewhere in the basin. The lowland was the main pathway for CRBG lavas flowing between the vent area and western Oregon and Washington (Reidel and Tolan, 1992).

THE MIDDLE MIOCENE COLUMBIA BASIN

Borehole, geophysical, and stratigraphic data (Bergstrom and others, 1987; Catchings and Mooney, 1988; Reidel and others, 1989b) indicate that the CRBG thins onto the Palouse Slope and thickens into the YFB (Fig. 4). The CRBG ranges from 500 to 1,500 m thick on the Palouse Slope, but it abruptly thickens to as much as 4,000 m in the Pasco Basin area (Reidel and others, 1982, 1989b). Regional thickness patterns for both the CRBG and underlying Tertiary sedimentary rocks indicate that the pre-CRBG basin was subsiding relative to the Blue Mountains and Palouse Slope from at least the Paleocene or Eocene though the Miocene. By far the most significant regional tectonic activity was continued subsidence in the basin. The subaerial nature of the CRBG indicates that subsidence continued as long as basalt was being erupted and that basalt accumulation kept pace with subsidence (Reidel and others, 1982, 1989a, 1989b). Subsidence rates from 17 to 15.6 Ma were approximately 1 cm/yr initially and decreased to 3×10^{-3} cm/yr in the late Miocene (Reidel and others, 1989b).

During the eruption of the CRBG, the anticlinal ridges were topographic highs against which the basalt flows thinned with elevation during emplacement (Reidel, 1984; Reidel and others, 1989b; Anderson, 1987). Flow thickness variations provide a means to estimate fold growth rates (Reidel, 1984). The ridges grew at about 0.25 mm/yr during initial eruption of the CRBG (17–15.6 Ma); the rate de-



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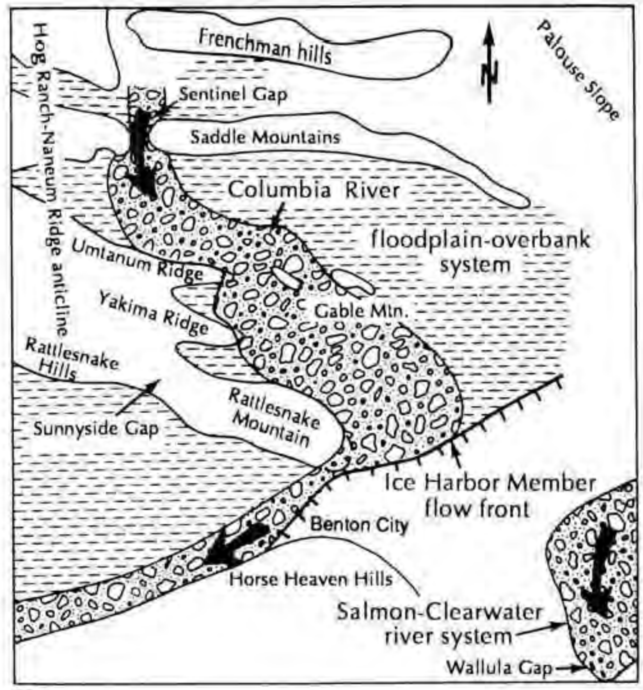

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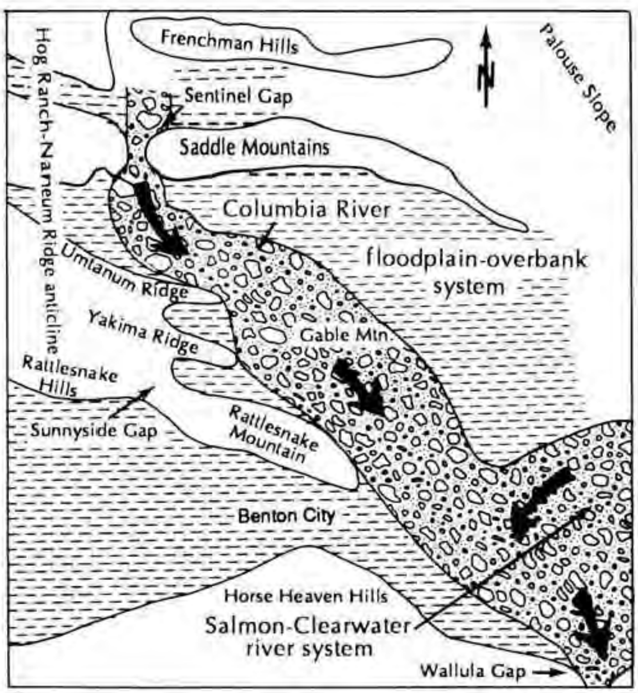

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Figure 5. Generalized evolution of the Columbia River system in the central Columbia Basin during the late Miocene and Pliocene (Modified from Fecht and others, 1987). A, Snipes Mountain time, approximately 10.0 to 8.0 Ma. B, approximately 8.0 to 6.0 Ma. C, about 6.0 Ma.

creased to about 0.04 mm/yr during the waning phases (15.6–10.5 Ma) (Reidel, 1984; Reidel and others, 1989b).

By the end of the massive eruptions of the CRBG (10.5 Ma), most of the Columbia Basin was a shallow, bowl-shaped, nearly featureless plain. The massive eruptions had buried most of the structural and topographic relief. In the western part of the Columbia Basin, standing above the plain were only the anticlinal ridges not inundated by younger flows. Across this plain flowed the ancestral Columbia River and its main tributaries, including the Salmon–Clearwater, Yakima, and Palouse Rivers.

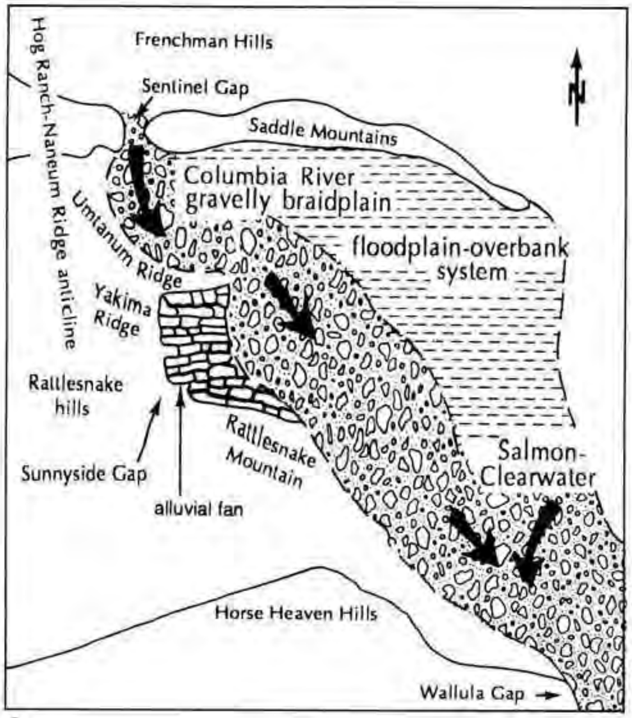
THE LATE MIOCENE TO MIDDLE PLIOCENE COLUMBIA BASIN

Post-CRBG tectonic history of the Columbia Basin is recorded in the Yakima folds and post-CRBG sediments. Alluvial-lacustrine sediments (Table 1) deposited primarily by the Columbia River system show that the Yakima folds were still growing and displacing river channels (Fecht and others, 1987). The structural evolution, as reflected in these sediments, is interpreted from: (1) the lateral distribution of facies, (2) changes in depositional style, and (3) structural deformation.

Lateral Distribution of Facies

Changes in the facies distribution in post-CRBG sediments are one of the best records of the post-CRBG history of the Columbia Basin. Sand and gravel in fluvial channels mark the locations of major drainages. Lesser alluvial fans and side-stream alluvial sequences were deposited adjacent to the main rivers. Ridge uplift and basin subsidence caused lateral shifts in these depositional environments over time (Fecht and others 1987; Smith, 1988; Lindsey, 1991).

During the waning phase of CRBG eruptions (12.5–8.5 Ma), the Columbia River flowed south across the YFB. The post-CRBG pre-Ringold channel (upper Ellensburg Formation and Snipes Mountain conglomerate) of the Columbia River passed across the western Pasco Basin, entering at Sentinel Gap (Reidel, 1984, 1988) and exiting near Sunnyside Gap (Fig. 5A). From there the river flowed



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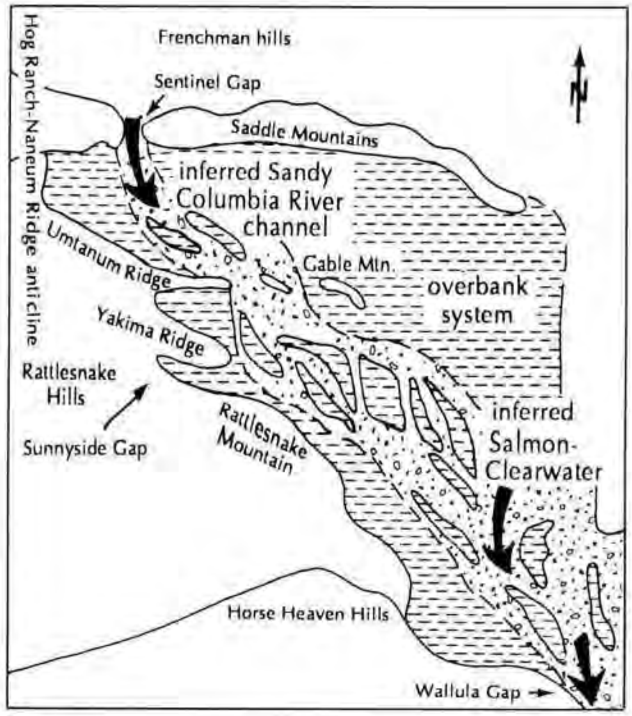
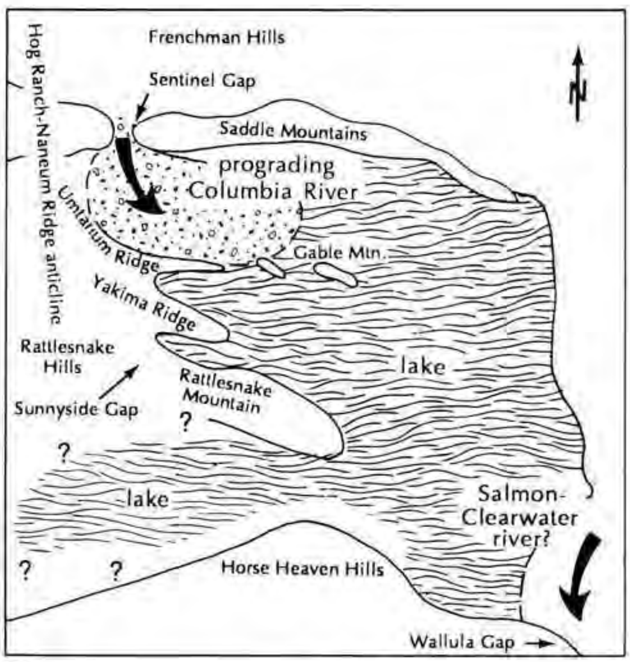


B

Figure 6. Generalized interpretation of the paleogeography of the Pasco Basin during each of the three phases of the Ringold Formation deposition. The first phase (A) ended approximately 6.0 Ma, the second phase (B) approximately 5.0 Ma, and the third phase (C) approximately 3.4 Ma.



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southwest toward Goldendale. At about 8 Ma, the Columbia River began to shift eastward into the central Pasco Basin, occupying a water gap over the eastern end of Rattlesnake Mountain near Benton City (Fecht and others, 1987; Fig. 5B). By middle Ringold time (approximately 6 Ma), the Columbia River had shifted its position again, exiting the Pasco Basin at Wallula Gap (Fig. 5C), as it does now (Fecht and others, 1987).

Concurrent with the eastward shift in the Columbia River, alluvial wedges deposited in the Yakima River drainage system prograded eastward (Smith, 1988). These wedges are seen in the upper Ellensburg Formation and Thorp Gravel.

Changes in Depositional Style

Three major changes in depositional style are found in the Ringold Formation. Each is marked by rapid, apparently basin-wide transitions. The first change occurred at approximately 6 Ma and is seen as a shift from gravelly braid plain and basin-wide paleosol systems of the lower part of the Ringold Formation (Fig. 6A) to the sandy alluvial systems of the middle part (Fig. 6B). The second shift in depositional style, which had occurred by 5.0 Ma, is marked by the replacement of sandy alluvial systems by widespread lacustrine conditions (Fig. 6C). Lacustrine deposits are found throughout the region, including the Yakima Valley (Smith, 1988) and north of the Saddle Mountains (Reidel, 1984, 1988). Lacustrine conditions and Ringold deposition ended with region-wide incision of the Columbia River system beginning approximately 3.4 Ma. Incision resulted in the removal of more than 100 m of Ringold section in the central part of the Pasco Basin. Many of the Pliocene and Pleistocene pedogenic carbonates in sediments overlying the Ringold Formation and found on the anticlinal ridges began to form following initiation of this incision.

These changes in depositional style are inferred to be related to regional factors that led to abrupt changes in gra-

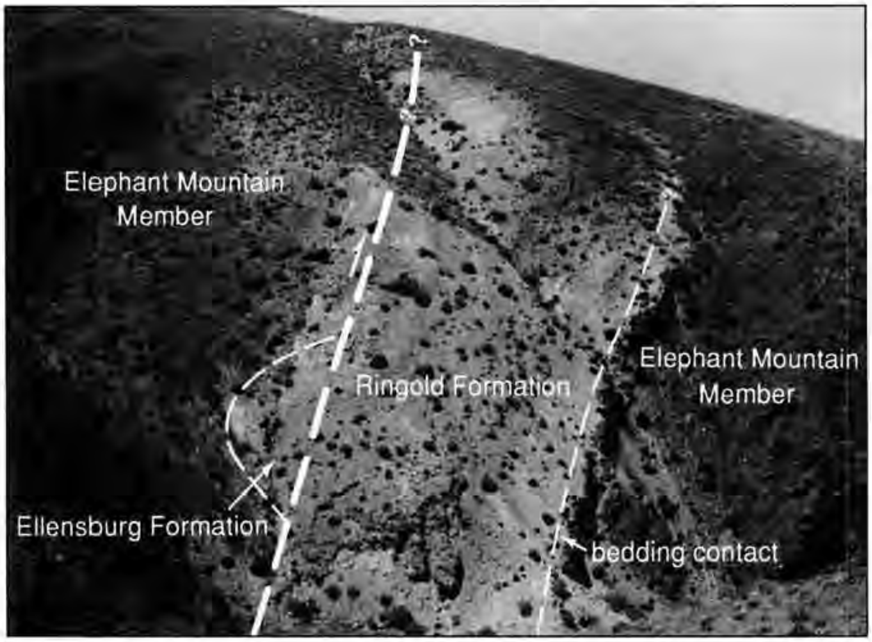

Figure 7. In this view to the west, the Ringold Formation overlies the Elephant Mountain Member of the Saddle Mountains Basalt, CRBG, in the Snively Basin area. This outcropping lies on the north side of the Rattlesnake Mountain structure and is part of the frontal fault zone. The Elephant Mountain Member has been thrust north onto the Ringold Formation.

dient in the Columbia River system. The limited evidence available suggests gradients could have been influenced by: (1) changes in uplift rates where the Columbia River crossed the southern extension of the HR–NR anticline; (2) increased Cascadian volcanism in the Columbia River Gorge between 6 and 3 Ma (Tolan and Beeson, 1984; Tolan and others, 1984); (3) upstream changes in detrital input into the Columbia River system; and (4) headward erosion as base level dropped in the lower Columbia River. It is not yet possible to determine precisely which of these factors (or combinations of factors) influenced the Columbia River.

Distribution of Deformed Post-CRBG Sediments

Uplifted and faulted Ringold and coeval sediments flank most ridges in the central Columbia Basin (Fig. 7; Reidel, 1984, 1988; Reidel and others, 1989b). The elevations at which they are found reflect the amount of structural development on the ridges after the sediments were deposited. Deformed Pliocene and Pleistocene sediments also are found on many ridges.

PLEISTOCENE AND YOUNGER DEFORMATION IN THE COLUMBIA BASIN

Probably all geologic structures of the Columbia Basin had developed their present relief by the end of the Pleistocene. Evidence for continued growth of the YFB is mainly in the frontal fault zones. Although not common, evidence of Pleistocene faulting has been found at many locations across the YFB (Fig. 8; Table 4). Young faults have been described at Toppenish Ridge, at Union Gap in Ahtanum Ridge, on Gable Mountain along Umtanum Ridge, in the Columbia Hills anticline, and along the CLEW (Table 4). Age relations are generally poorly constrained, however, but they suggest that faulting has continued since the last cataclysmic flood (approximately 13,000 yr B.P.). The minimum age of this faulting is not known, but no seismically active faults have yet been identified. Younger glaciofluvial sediments of the Hanford formation locally record some of the youngest deformation in the Columbia Basin.

Toppenish Ridge

Campbell and Bentley (1980, 1981) describe a 0.5- to 2.2-km-wide zone of nearly 100 surface ruptures along a 32-km-long segment of the north flank of Toppenish Ridge (Fig. 9). The scarps are subparallel to the ridge trend and range from 0.1 km to 3 km in length. The lowest scarp (Mill Creek) is a thrust fault (Fig. 10), but all others are high-angle normal faults. Individual ruptures have displacements as great as 4 m; organic material in the ruptures yields ^{14}C ages of 505 ± 160 yr and 620 ± 135 yr (Campbell and Bentley, 1981).

Ahtanum Ridge

A Pleistocene or younger fault was exposed during highway construction near Union Gap on Ahtanum Ridge (Figs. 8 and 11). A high-angle reverse fault offsets Yakima River terrace gravels by at least 7 m, juxtaposing them against the basalt of Ginkgo, Frenchman Springs Member (N. P. Campbell, *in* Washington Public Power Supply System, 1981a). A questionable U-Th caliche age of 30,000 yr was obtained for the terrace gravels (Campbell, 1983). The fault gouge appears to be capped by undeformed 13,000-yr-old slackwater flood sediments, but exposures 1 km to the east reveal faulted slackwater sediments (N. P. Campbell, unpub. data, 1992).

Umtanum Ridge–Gable Mountain

Glaciofluvial deposits 13,000 years old are offset as much as 6.5 cm by the central Gable Mountain fault (Puget Sound Power and Light Co., 1981) at the east end of Umtanum Ridge (Fig. 8). This tear fault has a component of reverse movement; it shows increasing offset in progressively older units.

Columbia Hills

On the western side of the Columbia Basin, Anderson and Tolan (1986) report a Holocene fault cutting Pleistocene slackwater flood sediments on the south side of the Columbia Hills anticline near Goldendale.

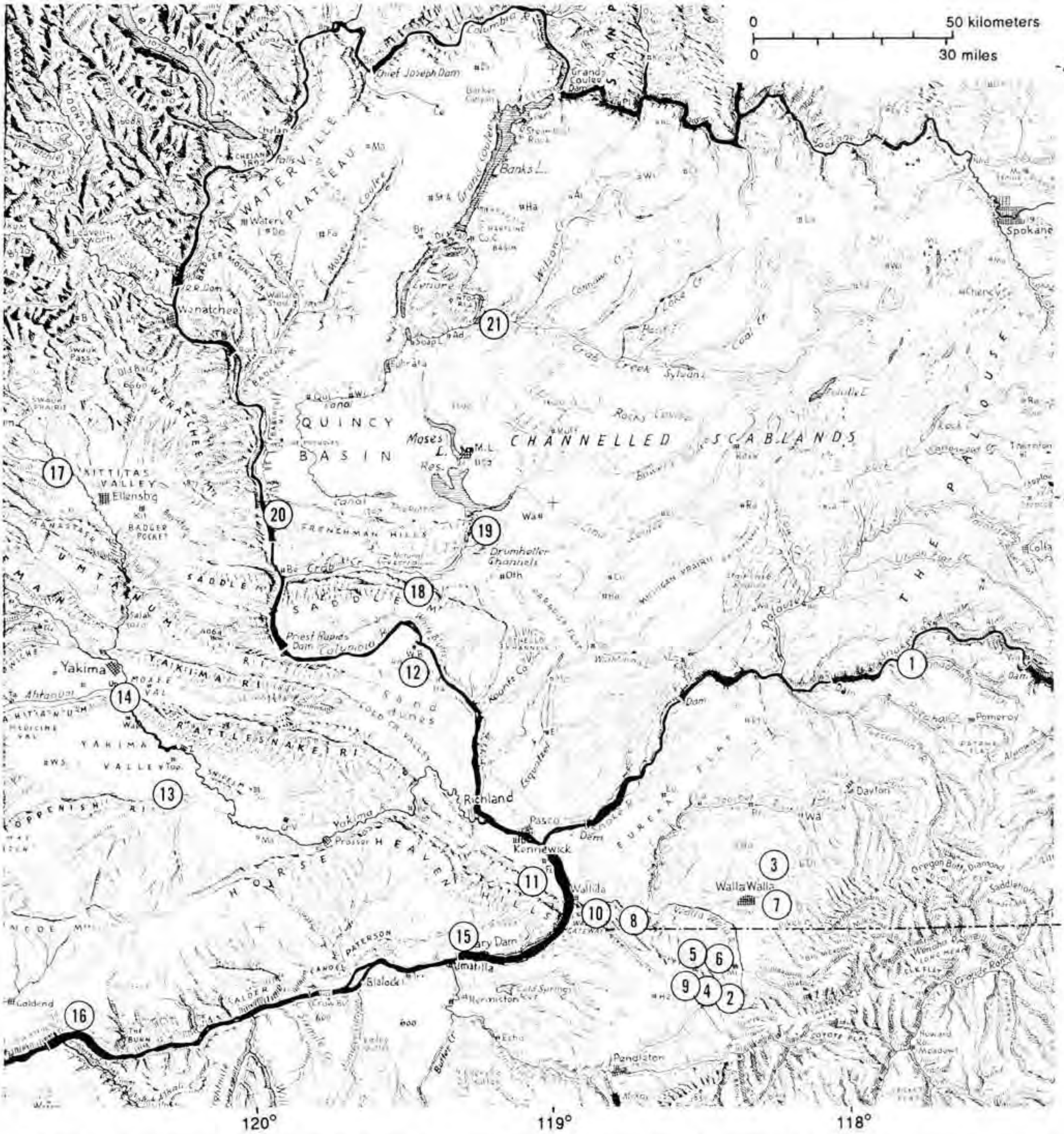

Figure 8. Location of Pleistocene–Holocene faults and possible Pleistocene–Holocene faults in the Columbia Basin. See Tables 4 and 5 for descriptions of the faults.

CLEW

Quaternary deformation has been reported at 15 localities along the CLEW (Washington Public Power Supply System, 1981a; U.S. Department of Energy, 1988; Tolan and Reidel, 1989) between Kennewick and Milton Freewater, OR (Table 4). The widespread distribution of Quaternary faulting in the Columbia Basin, however, suggests that the CLEW does not have any more or fewer Quaternary faults than other parts of the YFB.

Other Localities

Undocumented and (or) unpublished data for other Quaternary or younger faults include the following (Table 5, Fig. 8): Manastash Ridge (Geomatrix Consultants, Inc., 1988), Boylston Mountains (R. D. Bentley, Central Wash. Univ., oral commun., 1983), Cleman Mountain (Geomatrix Consultants, Inc., 1988; N. P. Campbell, unpub. data), Yakima Ridge (R. D. Bentley, oral commun., 1990), Medicine Valley (Geomatrix Consultants, Inc., 1988), Hog

Table 4. Pleistocene–Holocene faults in the Columbia Basin; NK = not known; see Fig. 8 for locations

Fault	Primary structural feature	Age of last movement	Sense of movement	Amount of movement	Location	Source of information
1. Central Ferry	Palouse Slope	Pleistocene?	sinistral oblique-slip	1 to 1.5 m	center sec. 22, T12N, R40E	Foundation Sciences, Inc. (1980, p. 25)
2. Thorn Hollow	Hite Fault System	early Holocene?	strike-slip	not determined	SW¼NE¼ sec. 2, T4N, R35E	Kienle and others (1979, p. 28)
3. Buroker	Wallula Fault System	Pleistocene to early Holocene?	thrust fault with component of sinistral strike-slip	>1 m	sec. 31, T7N, R37E	Foundation Sciences, Inc. (1980, p. 41-44)
4. Little Dry Creek	Wallula Fault System	Pleistocene?	normal	0.5 m	NE¼ sec. 11, T4N, R35E	Kienle and others (1979, p. 33-34)
5. Barrett	Wallula Fault System	late Pleistocene to Holocene?	dextral oblique-slip	varied, 2 to 50 cm	SW¼SE¼ sec. 25, T6N, R34E	Kienle and others (1979, p. 35)
6. Milton-Freewater	Wallula Fault System	Holocene (1936)	dextral strike-slip	ground breakage not located	SE¼ sec. 18, T5N, R35E	Kienle and others (1979, p. 36)
7. Promontory Point	Wallula Fault System	Pleistocene?	normal	NK	sec. 10, T6N, R37E	Kienle and others (1979, p. 40); Foundation Sciences, Inc. (1980, p. 40-41)
8. Wallula (near Warm Springs Canyon)	Wallula Fault System	Pleistocene?	strike-slip or oblique-slip	NK	sec. 12, T6N, R32E	Farooqui (1979, p. 8-9)
9. Umapine	Wallula Fault System	Pleistocene? to Holocene	oblique-slip	NK	sec. 15, T5N, R34E	Mann and Lewis (1991)
10. Wallula (near Vansycle Canyon)	Wallula Fault System	early Holocene	not determined	NK	sec. 3, T6N, R32E	Glass (1977, p. 2R K8-K9); Farooqui (1979, p. 10-11)
11. Finley Quarry	Rattlesnake–Wallula lineament	Pleistocene?	reverse	NK	sec. 3, T7N, R30E	Farooqui and Thoms (1980, p. 4-8)
12. Central Gable Mountain	Gable Mountain anticline (Yakima fold)	late Pleistocene?	reverse	~6 cm	sec. 19, T13N, R27E	Puget Sound Power and Light Company (1981, p. 34-45)
13. Mill Creek thrust fault and numerous unnamed faults	Toppenish Ridge (Yakima fold)	Holocene	both normal and reverse	as much as 4 m	area between lat. 46°15′–46°19′N, long. 120°22′–122°40′W	Campbell and Bentley (1981, p. 519-524)
14. Union Gap	Ahtanum Ridge (Yakima fold)	Pleistocene?	reverse	~7 m	T12N, R19E	Washington Public Power Supply System (1981b, p. 2.5K-53); Geomatrix Consultants, Inc. (1988, p. 47-54)

Ranch–Naneum Ridge anticline (R. D. Bentley, oral commun., 1986), Tamarack Springs (Campbell, 1983), Frenchman Hills and Smyrna Bench, Saddle Mountains (Geomatrix Consultants, Inc., 1990; West and Shaffer, 1989; Shaffer and West, 1989; S. P. Reidel, unpub. data), Wenas Valley (West, 1987), Kittitas Valley (Geomatrix Consultants, Inc., 1989), and Ahtanum Ridge (N. P. Campbell, unpub. data; Geomatrix Consultants, Inc., 1989).

There appears to be no pattern of faulting in the YFB that would suggest that Pleistocene–Holocene faulting is more concentrated in one part of the basin than in another. Rather, the distribution seems to suggest that the entire fold belt has continued to develop in a pattern similar to that of the Pliocene and Miocene.

CONTEMPORARY STRESS AND STRAIN

Regional Stress Indicators

In situ stress for the Columbia Basin has been determined from geodetic surveys and earthquake focal mechanism solutions. Geodetic surveys (Prescott and Savage, 1984) across the Pasco Basin suggest north-south shortening at a

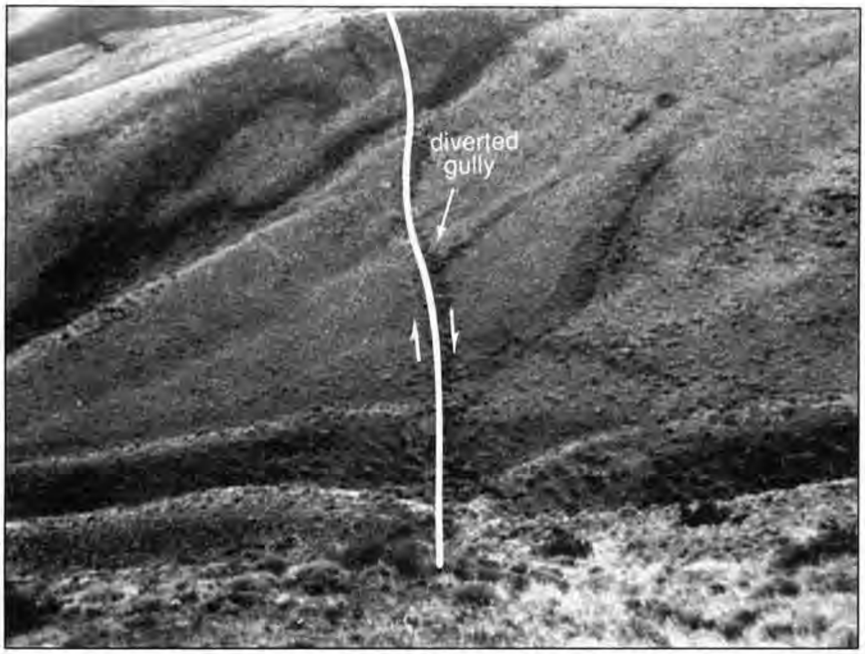

Figure 9. Holocene fault on Toppenish Ridge near Yakima, WA. View southeast along the north flank of Toppenish Ridge at the fault scarp. The fault scarp offsets the CRBG, alluvium, and loess. Note the diverted gully.

rate of -0.27 ± 0.22 microstrain/yr. The principal strain directions are N3°W $\pm$ 34° at 0.06 ± 0.013 microstrain/yr and N87°E $\pm$ 34° at -0.024 ± 0.014 microstrain/yr). This rate is not statistically significant at the 95 percent confidence level, however, and the measurements are within the error limits of the recording instruments.

Focal mechanism solutions are more definitive, and they indicate that the maximum principal stress is generally north-south and the minimum principal stress is near vertical (U.S. Department of Energy, 1988). Seismicity occurs in three stratigraphic zones: the CRBG, the sub-basalt sedimentary rocks, and the crystalline basement. Most of the seismicity is concentrated in the CRBG and the crystalline basement. Almost no seismic events have been associated with known faults; seismicity tends to be concentrated in the synclinal areas (Fig. 12; U.S. Department of Energy, 1988).

The only recognized association between seismicity and faults is on the Smyrna monocline on the north flank of the Saddle Mountains (Reidel, unpub. data). This is a segment boundary between the Smyrna Bench and Saddle Gap segments of Reidel (1984, 1988). Seismicity occurs north of the Saddle Mountains (Fig. 12) on the west side of a suspected deep fault marking the boundary and on the south side of the Saddle Mountains on the east side of the boundary. A time analysis of the microseismicity shows that this deep suspected fault controls the pattern of earthquake occurrence.

Geodetic measurements and, in particular, the pattern of seismicity indicate that the maximum stress in the Columbia Basin is horizontal, north-south compression (Fig. 13). The intermediate stress is horizontal and east-west, and the minimum stress is vertical. These data are consistent with geologic evidence (Reidel and others, 1989b) suggesting that the Columbia Basin has been under north-south compression since at least the Miocene and that the same stress pattern continues today.

Contemporary Stress in the Cold Creek Syncline

Core disking and spalling in core holes drilled in the Cold Creek syncline, central Columbia Basin, indicate high *in situ* stress (U.S. Department of Energy, 1988). Core disking (Fig. 14) occurs when drill core fractures into thin disks during drilling. Borehole spalling or

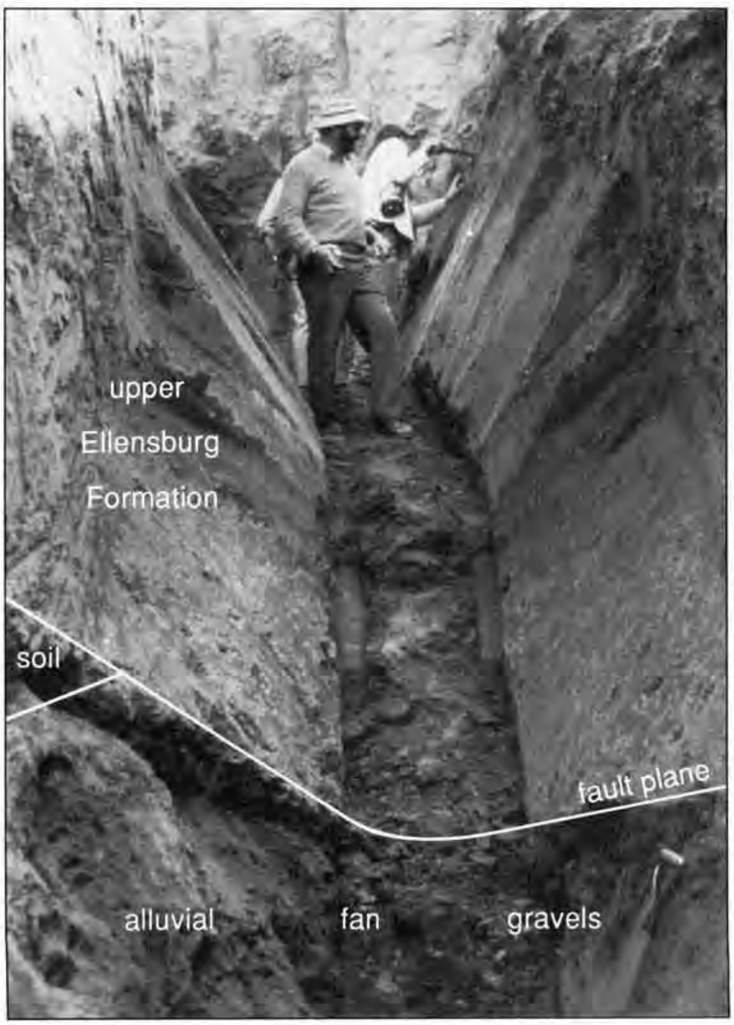

Figure 10. Trench through the Mill Creek thrust fault at the base of the north limb of Toppenish Ridge. This fault places pumicite of the upper Ellensburg Formation over fan gravels and a wedge of soil (on the left side of the photograph).

Table 5. Localities of suspected Pleistocene–Holocene faults in the Columbia Basin; NK = not known; see Fig. 8 for locations

Fault	Primary structural feature	Age of last movement	Sense of movement	Amount of movement	Location	Source of information
15. Unnamed	Service anticline	Late Pleistocene?	strike-slip	NK	SE¼SW¼ sec. 28, T6N, R28E	Foundation Sciences, Inc. (1980, p. 48-49)
16. Luna Butte	Columbia Hills (Yakima fold)	Early Holocene	dextral strike-slip	NK	sec. 8, T3N, R18E	Anderson and Tolan (1986, p. 82)
17. Kittitas Valley (3 faults)	Kittitas Basin (Yakima fold)	Pleistocene?	normal	2 m?	Area between lat. 47°00′–47°10′N, long. 120°25′–120°45′W	Waitt (1979)
18. Smyrna Bench	Saddle Mountains (Yakima fold)	Late Pleistocene–Holocene?	normal-reverse	6 m?	T15–16N, R25–27E	Geomatrix Consultants, Inc. (1990); West and Shaffer (1989)
19. Frenchman Hills (2 segments)	Frenchman Hills (Yakima fold)	Holocene	reverse	2 m?	T17–18N, R27–29E	Geomatrix Consultants, Inc. (1990); Shaffer and West (1989)
20. West Canal	Frenchman Hills (Yakima fold)	Pleistocene	reverse	1–3 m	T18N R23E	Grolier and Bingham (1971); Geomatrix Consultants, Inc. (1990)
21. Pinto	Pinto Ridge	Pleistocene	NK	NK	lat. 47°30′N, long. 119°15′W	Geomatrix Consultants, Inc. (1990)

breakout is also an indicator of high deviatoric stress and suggests that the *in situ* stress is not distributed lithostatically. Spalling occurs in the direction of least horizontal compression. The consistent east-west (Fig. 15) orientation of borehole spalling also indicates that the maximum horizontal compression is oriented generally north-south.

Stress Magnitude at Depth

Hydraulic fracturing is currently the most widely used method to directly determine the magnitude of *in situ* stress. Hydraulic fracturing tests were conducted in boreholes at about 1 km depth in the upper part of the Grande Ronde Basalt in the Cold Creek syncline (U.S. Department of Energy, 1988). The results indicated that the maximum horizontal stress ranges from 52.6 to 67.4 MPa (7,630–9,780 lbf/in^2) and the minimum horizontal stress ranges from 30.3 to 35.7 MPa (4,400–5,180 lbf/in^2). The ratio of average horizontal stress [(σH + σh)/2] to the vertical stress (σv) ranges from 1.41 to 2.14, with a mean value of 1.77 ± 0.20. This ratio is close to the higher end of known stress conditions at comparable depths at other locations (Fig. 16 on p. 178). The mean orientation of induced fractures, and thus the direction of the maximum horizontal stress, is consistent with north-south compression (Paillet and Kim, 1987).

Figure 11. Roadcut along Interstate Highway 82 at Union Gap in which a Holocene fault is exposed. Frenchman Springs Member, Saddle Mountains Basalt, CRBG, is thrust south over terrace gravel of the Yakima River. Vertical reinforcing bar is for a retaining wall along the highway.

In Situ Stress, Faults, and Microseismicity

Given the high *in situ* stress conditions in the central Columbia Basin, Kim and others (1986) suggested that movement is possible on an east-striking reverse fault that dips 60 to 65 degrees and has an effective friction angle of 33 degrees or less along the fault plane. A comparison of shear strengths (Byerlee, 1978) to test results on the properties of CRBG joint surfaces (U.S. Department of Energy, 1988) suggests that slip could occur on a fault plane in the present stress field.

The regional pattern of Pleistocene and Holocene faulting indicates that the hypothesis of Kim and others (1986) is

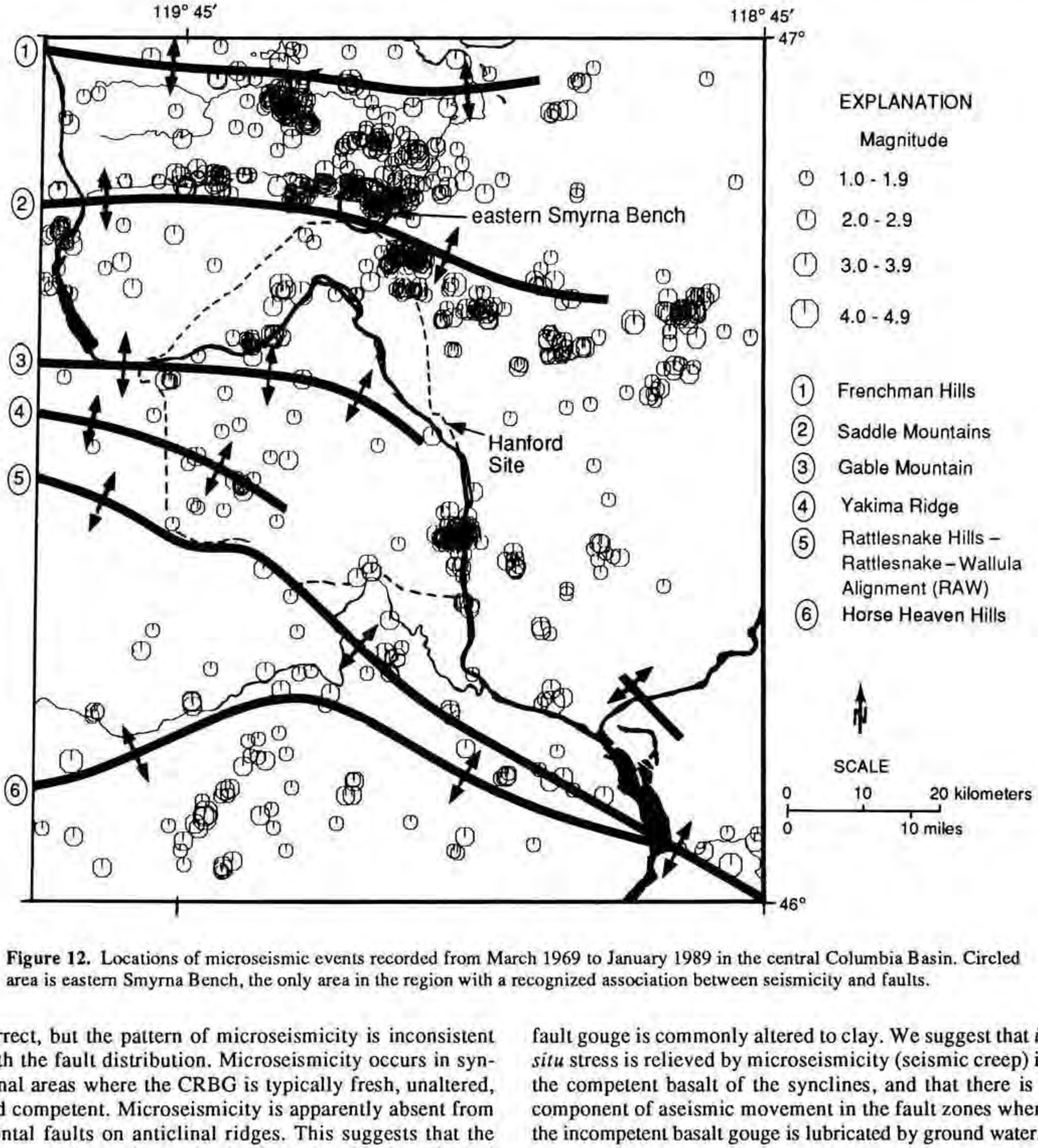

Figure 12. Locations of microseismic events recorded from March 1969 to January 1989 in the central Columbia Basin. Circled area is eastern Smyrna Bench, the only area in the region with a recognized association between seismicity and faults.

correct, but the pattern of microseismicity is inconsistent with the fault distribution. Microseismicity occurs in synclinal areas where the CRBG is typically fresh, unaltered, and competent. Microseismicity is apparently absent from frontal faults on anticlinal ridges. This suggests that the ridges are deforming aseismically or that they have locked up and are not moving.

It is difficult to imagine why the strong, competent rock in the synclines would continually exhibit widespread microseismic activity while the incompetent, weak, altered basalts in the presence of abundant ground water in the CRBG aquifers would not move and apparently be locked up. The regional high *in situ* stress must be acting on the anticlinal ridges. Some insight into this problem comes from examining the frontal fault zones on the anticlinal ridges. The faults are marked by highly brecciated basalt; fault gouge is commonly altered to clay. We suggest that *in situ* stress is relieved by microseismicity (seismic creep) in the competent basalt of the synclines, and that there is a component of aseismic movement in the fault zones where the incompetent basalt gouge is lubricated by ground water.

SUMMARY AND CONCLUSIONS

The Columbia Basin is the product of deformation that began early in the Tertiary, prior to the eruption of the CRBG, and continues today. The Columbia Basin has two fundamental parts important to the post-CRBG geologic history of the Columbia Basin, the Palouse Slope and the YFB. The Palouse Slope is a stable, undeformed area overlying the old continental craton. The YFB overlies a large prebasalt basin that has been subsiding since the early Tertiary. The basin is divided by the HR–NR anticline. The edge of the

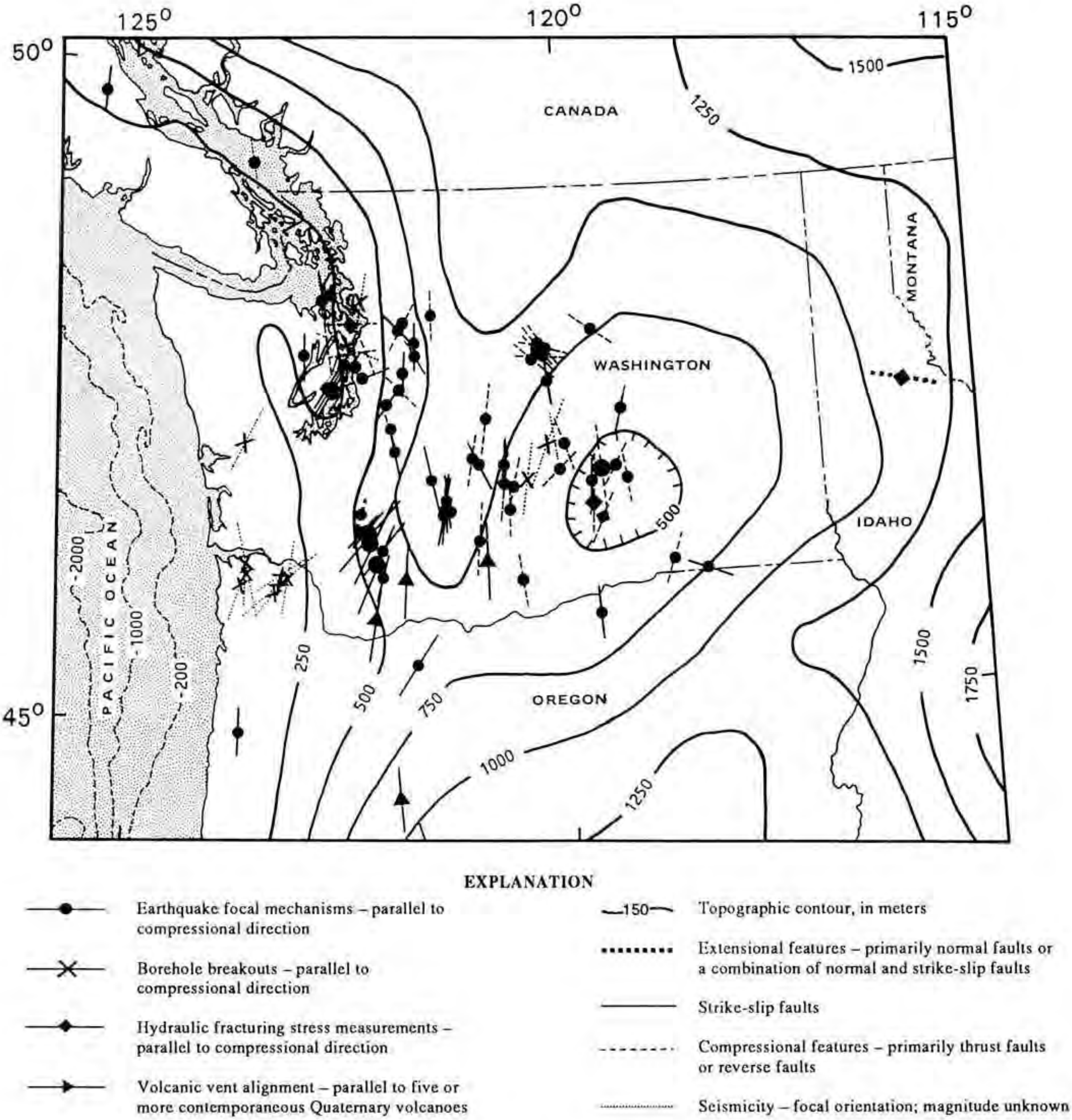

Figure 13. Stress orientation in the Pacific Northwest (modified from Zoback and others, 1992).

old continental craton is a suture zone that lies at the junction of the two structural subprovinces and is presently marked by the Ice Harbor dike swarm of the CRBG.

The pattern of deformation in the Columbia Basin has been dominated by north-south compression, and the YFB is the principal product of this stress. This deformation has also controlled the location of the Columbia River system since the late Miocene, as well as the depositional pattern of the post-basalt sediments.

The rate of deformation in the Columbia Basin has declined since the mid-Miocene, that is, the rates of basin subsidence and ridge growth have both declined. The present rate of ridge growth is estimated at 0.04 mm/yr, and subsidence in the basin is estimated to be 3×10^{-3} mm/yr.

Microseismicity, high *in situ* stress conditions, and areas of Pleistocene and Holocene faulting indicate that the basin is still experiencing north-south compression. Although known late Cenozoic faults are found on anticlinal

Figure 14. Disking in core, borehole DC-6, Hanford Site.

ridges, earthquake focal mechanisms and strain measurements suggest that most present-day stress release is occurring in the synclinal areas. No earthquake events have been shown to be related to known faults. The high *in situ* stress in the Cold Creek syncline explains the microseismicity in that area, but the absence of microseismicity associated with the anticlinal ridges may result from a component of aseismic slip on weakened fault zones lubricated with ground water.

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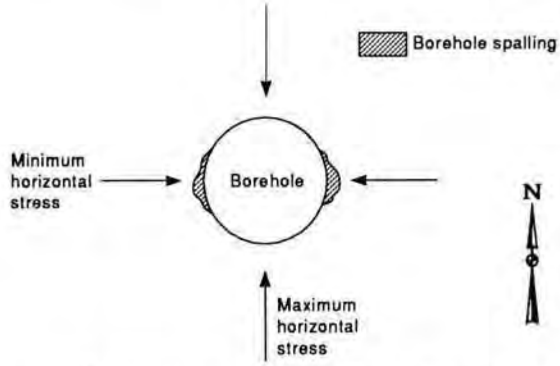

Figure 15. Generalized pattern of borehole spalling in the Cold Creek syncline. Redrawn from U.S. Department of Energy (1988).

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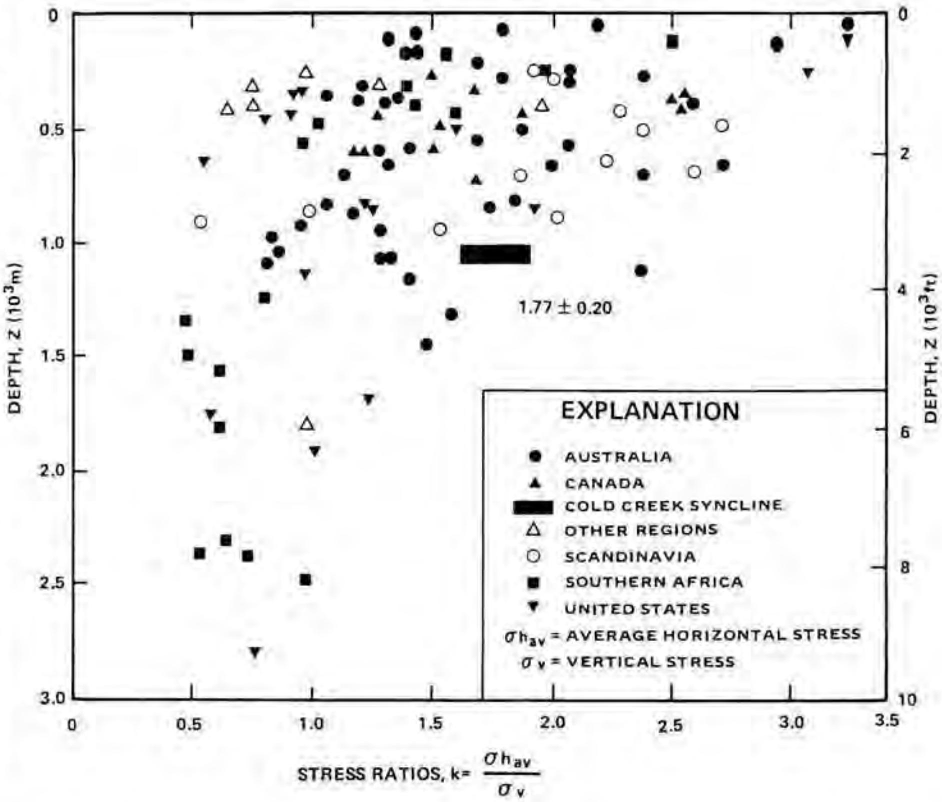

Figure 16. Worldwide horizontal/vertical stress ratio plotted versus depth (from U.S. Department of Energy, 1988).

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Regional Sedimentation of Late Quaternary Loess on the Columbia Plateau: Sediment Source Areas and Loess Distribution Patterns

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ABSTRACT

Recent development of a stratigraphic framework of buried soils, loess strata, and correlated volcanic tephra layers in late Quaternary loess in the Palouse and Channeled Scabland areas makes it possible for the first time to assess regional patterns of loess distribution and source areas and materials for loess on the Columbia Plateau. Stratigraphic descriptions were made of, and samples were collected from, two informally named loess units, L1 and L2, at 95 sites. We used geostatistical methods to produce contour maps of thickness and geometric mean particle size of L1 and L2.

L1 consists of loess that overlies Mount St. Helens tephra set S (about 13,000 radiocarbon [^{14}C] years B.P.). L1 loess is more than 400 centimeters thick to the northeast of the Walla Walla Valley and the Pasco Basin, and it thins progressively with distance to the northeast. The 200-centimeter isopach has a bilobate pattern. One lobe extends 110 kilometers to the north-northeast and coincides approximately with the Cheney–Palouse Scabland; a second, broader lobe extends 70 kilometers to the east-northeast on both sides of the Snake River into the main Palouse. L1 then thins to 100–150 centimeters around the margin of the Columbia Plateau. Geometric mean particle size of L1 loess has a complex pattern across the plateau that does not coincide with that of thickness. It is as large as 60 micrometers in thin L1 loess around the Quincy Basin and the lower Cheney–Palouse Scabland where localized bodies of sandy flood sediments are abundant. Mean size is only 30 micrometers in the thickest L1 loess, which is close to abundant deposits of silty slackwater sediments. The contrasting patterns of thickness and mean size strongly suggest that the source of L1 was multiple bodies of upper Wisconsin sediments deposited by floods that originated from glacial Lake Missoula. The patterns further suggest that dust-transporting winds during the late Wisconsin and Holocene were southwesterly and westerly as they are today.

L2 consists of loess below the Mount St. Helens S tephra layer and above the buried Devils Canyon Soil formed in still older loess. A proxy marker near the base of L2 loess is the Mount St. Helens set C tephra layer (about 36,000 to more than 42,000 ^{14}C years B.P.). L2 has a regional thickness pattern much like that of L1 but is thicker: it is more than 750 centimeters thick to the northeast of the Walla Walla Valley and thins progressively to the northeast. The 200-centimeter isopach extends north-northeast 120 kilometers along the scabland and northeast 60 kilometers into the Palouse. The geometric mean particle size of L2 exceeds 50 micrometers only near the confluence of several flood coulees in the south-central scabland. It decreases gradually in a northeasterly direction to a minimum of 14 micrometers along the north and east margins of the plateau. The similar overall patterns of thickness and grain size of the two loess units are consistent with independent evidence that an episode of giant floods during middle or early Wisconsin time brought flood slackwater sediments to the Pasco Basin and Walla Walla Valley, providing the source of sediment for the L2 loess.

INTRODUCTION

Loess deposits of the Palouse and Channeled Scabland areas are a prominent geologic feature in eastern and central Washington (Fig. 1). The geomorphologist Kirk Bryan was perhaps the first to recognize that an intriguing history of Pleistocene time on the Columbia Plateau might be contained in the windblown silts (Bryan, 1927).

We now know that regional Pleistocene history includes the largest flood flows documented on Earth in the glacial Lake Missoula–Channeled Scabland system (Fig. 1). The loess indeed contains important information about Pleistocene history: for example, loess deposits adjacent to flood coulees contain sedimentologic and stratigraphic evidence that several, perhaps many, episodes of cataclysmic floods (jökulhlaups) occurred in the Channeled Scabland in pre-Wisconsin time (Patton and Baker, 1978; McDonald and Busacca, 1988). Bryan was correct in suggesting that the loess is ancient: recent evidence supports a record as long as 2 million years (Busacca, 1991). Recent studies have included examination of distal tephra layers interbedded in the loess (Foley, 1982; Nelstead, 1988; Busacca and others, 1992), paleosols (Foley, 1982; McDonald, 1987; Busacca, 1989; Feldman, 1989; McDonald and Busacca, 1990), regional stratigraphy (McDonald and Busacca, 1992), sedimentology (Ludwig, 1987; McDonald, 1987), and chronology (Kukla and Opdyke, 1980; Foley, 1982; Berger and Busacca, 1991).

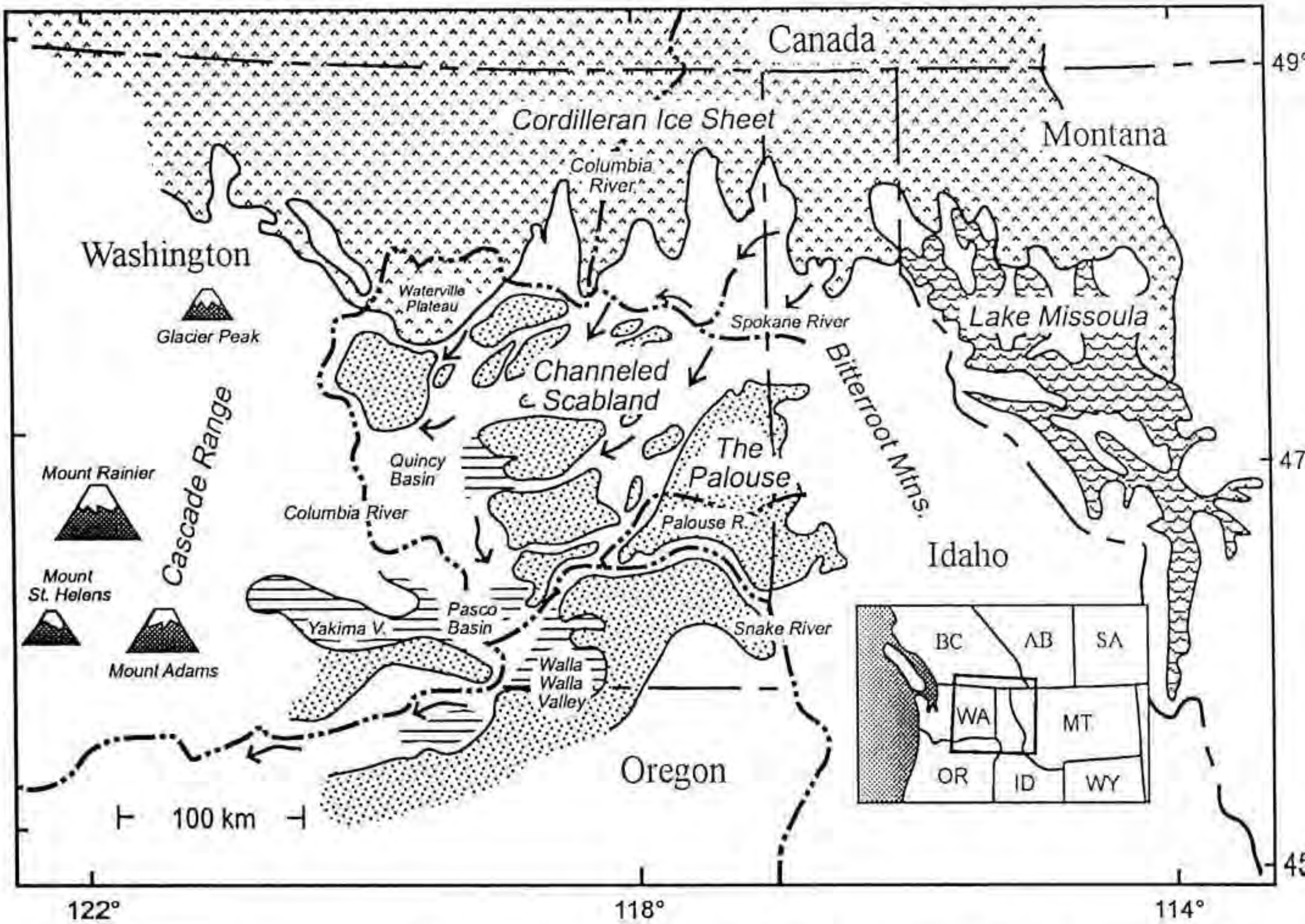


Figure 1. Location map of the Pacific Northwest showing the glacial Lake Missoula–Channeled Scabland system and the Cordilleran ice sheet at its late Wisconsin maximum, loess deposits (stipples), main areas of silty slackwater sediments from cataclysmic floods (horizontal lines), and nearby volcanoes of the Cascade Range. Arrows indicate generalized flood flow directions.

From the early recognition of these sediments as loess (Salisbury, 1901; Calkins, 1905; Bryan, 1927), questions have arisen as to the source area(s), source material(s), and spatial distribution patterns of the loess. Bryan (1927) recognized a paradox: The quartzo-feldspathic mineralogy of the loess suggests a source in the Rocky Mountains to the north and east of the Palouse, yet the present-day surface winds are predominantly southerly and westerly (Lewis, 1960; Philips, 1970).

The loess may have been derived from at least four potential sources that have grossly similar mineralogy: (1) alluvium of the Ringold Formation (Miocene and early Pliocene; Culver, 1937; Newcomb, 1958), which crops out in the Pasco and Quincy Basins; (2) outwash plains at the front of the Cordilleran ice sheet on the northern Columbia Plateau (Hobbs, 1947); (3) Cordilleran ice-sheet outwash in the Columbia River valley (Waitt, 1983); and (4) the Touchet Beds (Flint, 1938; slackwater sediments) left by cataclysmic floods in the Walla Walla and Yakima Valleys and the Pasco and Quincy Basins (Fig. 1).

Dispersal of suspended sediment by wind from source areas is well understood from classic studies of loess in mid-continental North America over the last 100 years (reviewed in Ruhe, 1983). Late Wisconsin loess in the Midwest is thinner with increasing distance from major river valleys that carried glacial outwash (Ruhe, 1983). A second major characteristic of midwestern loess is a fractionation to finer particle sizes as loess thins downwind.

Similar patterns have been suggested for the Palouse loess on the Columbia Plateau (Rieger, 1952; Lewis, 1960; Ludwig, 1987; Busacca, 1991), although until now there were insufficient stratigraphic and age controls and sampling for detailed analysis. We report here initial results of a project we undertook to fill this gap in our knowledge. We hoped to document the patterns of loess that resulted from transport of eolian sediment on the Columbia Plateau during the late Pleistocene: the thickness and grain size of individual strata, where they originated, and the source sediments from which they were derived.

STRATIGRAPHY OF PALOUSE LOESS AND PALEOSOLS

Loess on the Columbia Plateau ranges from about 0.1 to 75 m thick. Tens of buried ancient soils (paleosols) are interstratified in thick loess (Busacca, 1989, 1991). The total stratigraphic record may span as much as 2 million years

(Busacca, 1991). Zones of relatively unaltered loess probably represent periods of rapid eolian sedimentation, whereas paleosols apparently represent periods of slower eolian sedimentation when landscapes were fairly stable and little erosion and deposition were taking place (McDonald and Busacca, 1990).

On the basis of recent work on the upper interval of the loess, which is exposed in many roadcuts on the Columbia Plateau, we have developed a regional stratigraphy and chronology for loess, paleosols, and tephras back to about 85,000 thermoluminescence years B.P. (TL yr B.P.) (Berger and Busacca, 1991; McDonald and Busacca, 1992; Busacca and others, 1992). Much of this stratigraphy is shown in Figure 2; this is the part that is pertinent to our discussion. At least six distal tephra layers from Cascade Range volcanoes have been recognized and correlated (Fig. 2; Busacca and others, 1992). The more widespread and important of these in the present context are the Mount St. Helens set S tephra (ca. 13,000 ^{14}C yr B.P.; 19,000 TL yr B.P.) and the Mount St. Helens set C tephra (ca. 36,000–42,000 ^{14}C yr B.P.; 46,000-57,000 TL yr B.P.; Mullineaux, 1986; Busacca and others, 1992; Berger and Busacca, 1991), each of which have been recognized at more than 30 sites.

Downward from the modern surface soil, the vertical sequence of buried pedostratigraphic units also has been defined (McDonald and Busacca, 1992) (Fig. 2): Sand Hills Coulee Soil, Washtucna Soil (which is a single soil in sites distal from the sediment source, a double soil in proximal sites; McDonald and Busacca, 1990), Old Maid Coulee Soil, and Devils Canyon Soil. Each buried soil is recognizable by the combination of its stratigraphic position, soil morphology, and relation to distal tephra layers.

METHODS

For this study of loess distribution patterns, we focused our sampling on two stratigraphic intervals that are widely exposed in roadcuts across the Columbia Plateau. We informally named the first loess layer L1. It is bounded by the pre-farming land surface and the approximate position of the Mount St. Helens set S tephra (Fig. 2). This is the loessal sediment deposited largely after the late Wisconsin episode of cataclysmic flooding in the Channeled Scabland. It contains the modern surface soil and the weakly developed Sand Hills Coulee Soil (Fig. 2). The lower boundary of L1 in most exposures is marked by the Mount St. Helens set S tephra, but where the tephra was not found, the base of L1 is defined instead by (a) the base of the lowermost calcic (Bk) horizon of the Sand Hills Coulee Soil in L1, (b) the top of the underlying Washtucna Soil, or (c) the top of underlying late Wisconsin flood sediments or top of a flood-cut unconformity, if either is present instead of (b).

We informally named the next loess layer L2. It lies generally beneath the Mount St. Helens set S tephra. Where the Mount St. Helens set S tephra was not found, we defined the top of L2 as the top of the Washtucna Soil. The base of L2 in most exposures is just below the Mount St. Helens set C tephra (Fig. 2). The base is more precisely

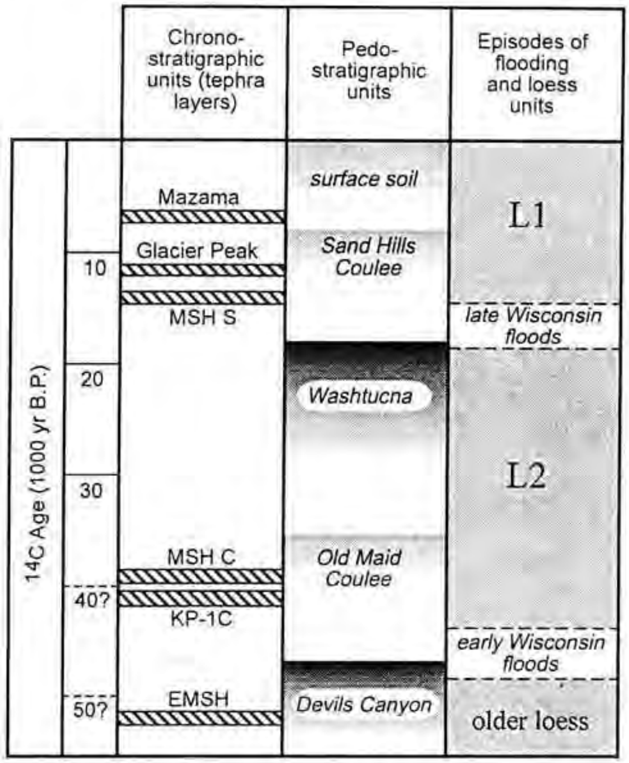


Figure 2. Composite diagram of late Quaternary chronostratigraphic marker tephra layers and pedostratigraphic units in Palouse loess (Busacca and others, 1992; McDonald and Busacca, 1992), informal loess units L1 and L2 (see text), and episodes of cataclysmic flooding. MSH S designates distal tephra layers in loess correlated with the Mount St. Helens set S eruptive sequence and MSH C with the Mount St. Helens set C eruptive sequence; KP-1C is a mafic-rich distal tephra layer of unknown source, and EMSH is a probable pre-set C Mount St. Helens eruption (Busacca and others, 1992). Lower boundary of loess unit L1 and upper and lower boundaries of L2 are defined on basis of multiple criteria; see text for explanation.

defined and recognized either as (a) the top of the underlying Devils Canyon Soil or (b) the top of underlying early Wisconsin cataclysmic flood sediments or of a flood-cut unconformity, where present. Broadly defined, L2 is the loess that was deposited after an early Wisconsin episode of cataclysmic flooding but before the well-known late Wisconsin episode. The weakly developed Old Maid Coulee Soil occurs near the middle of the unit.

Stratigraphic and soil descriptions were made from study of sediments exposed in trenches excavated in 95 roadcuts across the Columbia Plateau (Fig. 3). Sites were selected as much as possible to represent gentle south-facing present-day and paleoslopes, so that depositional position on the landscape would be nearly constant at all sites. The depth of exposure and stratigraphy varied at each site. For this reason we were able to obtain fewer measurements of layer thickness and particle size for L2 than for L1. The actual number of data points used in each regional analysis was: 95 for L1 layer thickness; 38 for L2 layer thickness;

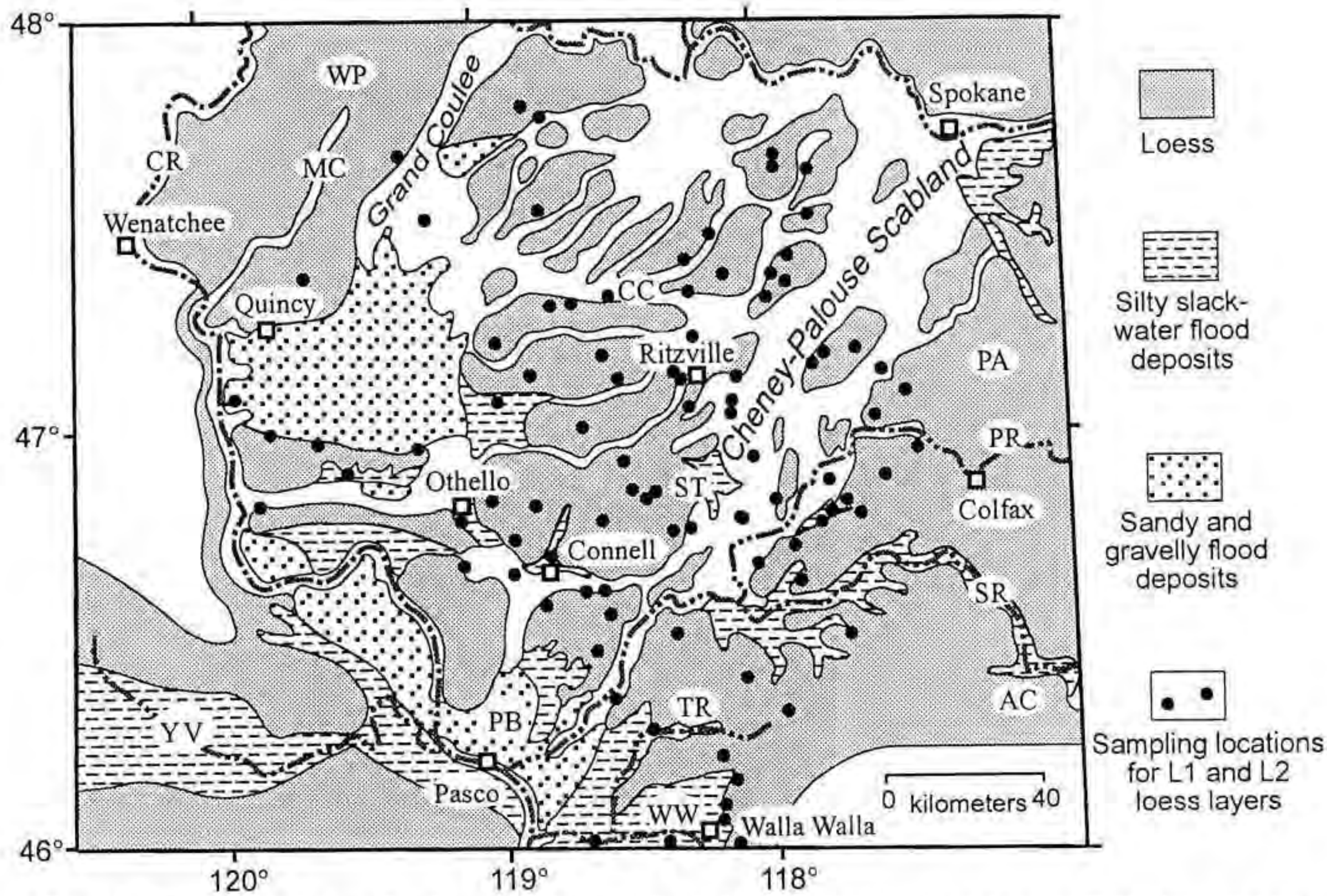


Figure 3. Surficial geologic map of the Channeled Scabland, modified from Bretz (1928). Black dots are sampling locations used to generate contour maps for L1 and L2 loess in Figure 4. AC, Alpowa Creek; CC, Crab Creek Coulee; CR, Columbia River; MC, Moses Coulee; PA, the Palouse; PB, Pasco Basin; PR, Palouse River; SR, Snake River; ST, Staircase Rapids; TR, Touchet River; WP, Waterville Plateau; WW, Walla Walla Valley; YV, Yakima Valley. Dash-dot line is modern course of Columbia River; dash-double dot lines are modern courses of Palouse, Snake, Spokane, Touchet, Walla Walla, and Yakima Rivers. Open squares show selected towns across the area.

44 for L1 geometric mean particle size; and 29 for L2 geometric mean particle size.

From five to ten depth-increment samples were collected from each of the L1 and L2 intervals at each site. Particle-size distribution was measured by the pipette-and-sieve method of Gee and Bauder (1986) after removal of $CaCO_3$ cements. Weighted means were calculated for each measured particle-size class (2,000–1,000,1,000–500, 500–250, 250–100, 100–50, 50–20, 20–2, <2 μm) from the increment samples. Geometric mean particle size of L1 and L2 was then calculated using the method of moments (Boggs, 1987). We used Golden Graphics *Surfer*® Software (version 3.00, 1987) to calculate semivariance and plot semivariograms for layer thickness and mean grain size of L1 and L2; the semivariogram models were then used in kriging (interpolation). Regional contour maps (Fig. 4) were generated from the kriged data sets.

RESULTS AND INTERPRETATIONS

For a number of years, sandy and silty sediments deposited by cataclysmic floods (Fig. 2) have been considered probable sources of late Wisconsin and Holocene loess on the Columbia Plateau, but inferences were based on analysis of fairly few samples for which there was little stratigraphic control. Our regional analysis of loess layer thickness and geometric mean particle size is based on many measurements of loess units L1 and L2. These units can be recognized across the plateau because they have consistent relations to buried marker soils and tephras. L1 is the loess unit that was deposited after the onset of late Wisconsin flooding perhaps 17,000 years ago. It buried the moderately developed Washtucna Soil (Fig. 2) that had been formed in L2 loess. Comparison of the slight development of the lower Holocene Sand Hills Coulee Soil and the surface soil with that of the Washtucna Soil suggests that deposition rates of L1 loess have remained high from the late Wisconsin through the Holocene.

L1 Loess

The results of our analyses of L1, presented in Figures 4a and 4b, show clear regional trends in layer thickness and mean particle size. L1 loess is more than 400 cm thick in only two areas of the entire Columbia Plateau. One lies directly to the northeast of the Pasco Basin, and a second area lies to the northeast of the Walla Walla Valley (Fig. 4a). A distinct bilobate pattern extending to the northeast of the Pasco Basin and Walla Walla Valley is seen in the 200-cm isopach of L1 loess. One lobe parallels the Cheney–Palouse Scabland and the second trends to the northeast into the Palouse area on both sides of the Snake River. A large lobe of loess more than 100 cm thick with a peak thickness of 200 cm occurs east of the Quincy Basin and the town of Othello. L1 loess remains thicker than 100 cm into the area of Spokane and northwest of Grand Coulee on the Waterville Plateau (Fig. 4a).

The thickest accumulations of upper Wisconsin loess in central North America have been found immediately to the east of flood plains of south-flowing rivers such as the Mississippi, Missouri, and Illinois, which carried sediment-charged glacial meltwater from the Laurentide ice sheet (Ruhe, 1983). Prevailing winds through the region are westerly or northwesterly. The upper Wisconsin loess decreases in thickness and mean particle size with increasing distance from source flood plains. These trends can be modeled mathematically (for example, Frazee and others, 1970; Handy, 1976; Ruhe, 1983). Loess thickness trends of L1 on the Columbia Plateau clearly indicate that the dominant loess source areas during latest Wisconsin and Holocene time have been the low-lying structural valleys of the southern part of the plateau.

Which sediments were the dominant contributors of dust that became L1 loess? The Pasco Basin contains deposits of both upper Wisconsin cataclysmic flood sediments and of the older Ringold Formation, so both are potential loess sources there. The Walla Walla Valley, in contrast, contains only one of these, the flood slackwater sediments. Thus, deflation of slackwater sediments must have been the source of the thick L1 loess that lies to the northeast of the Walla Walla Valley. In the Pasco Basin, perhaps both slackwater and Ringold sediments contributed to the thick L1 loess. This interpretation is consistent with observations that the mineralogy of L1 loess is similar to that of the slackwater Touchet Bed sediments (H. W. Smith, emeritus professor, Washington State Univ., oral commun., 1986), and that late Wisconsin and Holocene loess can be traced directly to eroding outcrops of slackwater sediments in the Walla Walla area (Fryxell and Cook, 1964).

The loess to the northeast of the Walla Walla Valley and Pasco Basin forms a fantastic topography of linear hills with their axes aligned to the north-northeast and northeast. Lewis (1960) explained this as a primary depositional feature resulting from long-continued accumulation of proximal-source loess in the lee of minor irregularities in the underlying basalt bedrock. Similar features, called gredas, have been described in loess derived from the Danube and other major rivers of Romania, Bulgaria, Hungary, Moldavia, and the Ukraine (Rozycki, 1968). There, these aligned forms show excellent conformity with prevailing dry-season winds and are best developed in deep, sandy loess along the escarpments of river valleys, especially where the valleys widen, such as at the mouths of tributaries (Rozycki, 1968).

Winds on the Columbia Plateau are most intense in spring and fall today and are dominantly southerly and southwesterly in those seasons in the southern part of the plateau (Philips, 1970). The lobate shape of the 200-cm isopach of L1 to the northeast of the Walla Walla and Pasco Basin areas (Fig. 4a) and the strongly linear hill shapes there indicate that, in general, the dust-transporting winds during the accumulation of the L1 layer blew from the same directions as they do today.

The isopach map (Fig. 4a) gives scant support to the hypothesis that the front of the Cordilleran ice sheet itself or glacial-alluvial sediments in the canyon of the Columbia River were major sources of L1 loess. The area of L1 loess more than 100 cm thick northwest of the Grand Coulee suggests that outwash carried down Moses Coulee from the Okanogan glacial lobe of the Cordilleran ice sheet on the Waterville Plateau (Figs. 1 and 2) may have been a local source of L1 loess.

In the idealized situation where a river valley trends perpendicular to prevailing wind and thereby serves as a line source of eolian sediment (such as the lower Mississippi River valley), loess thickness decreases logarithmically and monotonically from the source. The complexity of the thickness pattern in central Washington may be due in part to secondary or intermediate sediment sources along the path of dust transport. For example, the long arm of thick L1 loess that extends northeastward along the Cheney–Palouse Scabland, including one measurement of more than 250 cm of L1 loess at its northern end, probably accumulated both from long-distance transport of silt from the Pasco area and from local deflation of flood sediments within the scabland itself. Similarly, local areas of thicker L1 loess, such as those east of Othello and the Quincy Basin, are immediately downwind of smaller deposits of silty and sandy flood sediments (Fig. 4a). Another potential source of sediment is in the lower reach of the Snake River, and the bow to the northeast in the 250- and 200-cm isopachs there may reflect a secondary contribution from that canyon and its tributary valleys, which also contain slackwater sediments.

One surprising result of our measurements is that the extensive slackwater sediments in the Yakima Valley do not appear to have generated a corresponding lobe of thick L1 loess. We have very few measurements from this area (Fig. 3) because of lack of exposures, but we have seen no evidence for L1 loess thicker than about 100 cm east or northeast of the Yakima Valley. Even in the Kittitas Valley, the next valley north, the thickest accumulation of probable L1 loess is less than 40 cm (Herman Gentry, Soil Conservation Service, oral commun., 1993). A key reason may be that basaltic anticlinal ridges on all sides of the valley rise as much as 600 m above the valley floor. These ridges leave only narrow topographic outlets at the upper and lower ends of the valley. The surrounding hills may have collected dust on their steep upland flanks, effectively blocking loess from moving onto the plateau farther to the east, northeast, or north. We speculate that loess accumulating on the steep slopes may have been continuously removed by sheet erosion processes and carried into the Yakima River fluvial system. Observations of soil stratigraphy on the upland areas support this hypothesis.

The contour map of geometric mean particle size of L1 loess (Fig. 4b) provides further insight into the character and location of the source sediments that were deflated to form L1 loess. The areas of *coarsest* L1 loess do not coincide with the areas of *thickest* L1 loess (compare Figs. 4a and 4b). Two areas of very coarse loess (>50 µm mean particle size) are centered on the Quincy Basin and the lower Cheney–Palouse Scabland. The coarse L1 loess at the southern end of the Cheney–Palouse Scabland lies just to the east of a deposit of sandy flood sediments at the head of Staircase Rapids (Bretz, 1928; Fig. 4b). In the Quincy Basin, sandy flood sediments were reworked in Holocene time to form an extensive sand dune field. Coarse (but thin; Fig. 4a) L1 loess encircles and trends to the east from the dune field, suggesting that very fine sand and coarse silt were winnowed out of the flood deposits during dune formation and were transported relatively short distances to accumulate as coarse-textured L1 loess. Rieger (1952) and Ludwig (1987) previously reported that the Palouse loess fines with distance easterly from the Quincy Basin.

Small areas of coarse loess are associated with Crab Creek Coulee and other coulees near Ritzville. Mean size of L1 is progressively smaller to the east and northeast across the plateau, but the fining trend is interrupted by localized areas of coarse loess such as in Crab Creek Coulee. These areas highlight the important contributions made to the L1 loess system by locally reworked flood sediments in coulees. Our reconstruction of regional patterns of mean particle size in Figure 4b differs from the finding of Lewis (1960) that median grain size of loess decreases systematically with distance from Walla Walla to Spokane, perhaps because his sampling transect did not cross the bull's eye of coarse loess in the Cheney–Palouse Scabland.

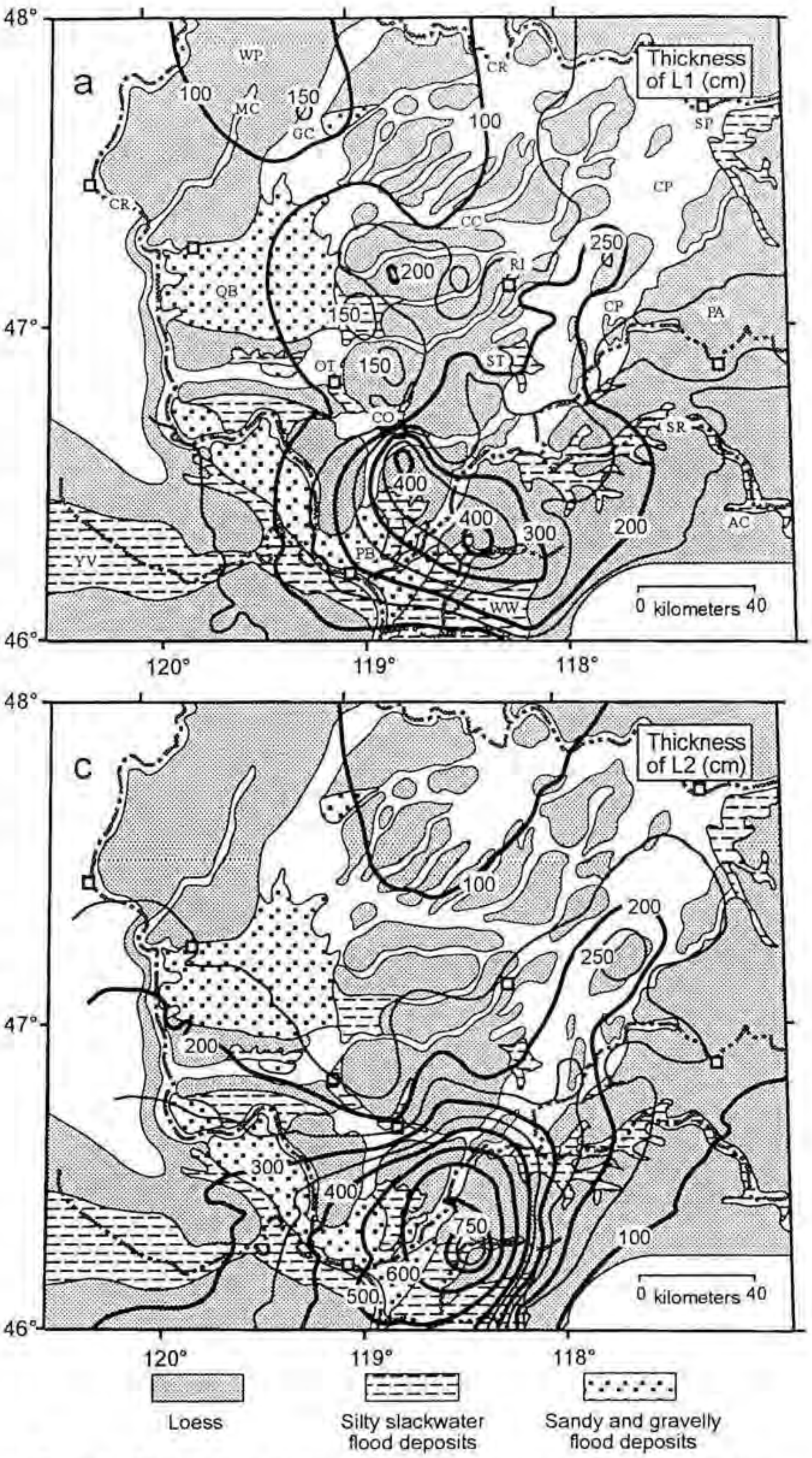
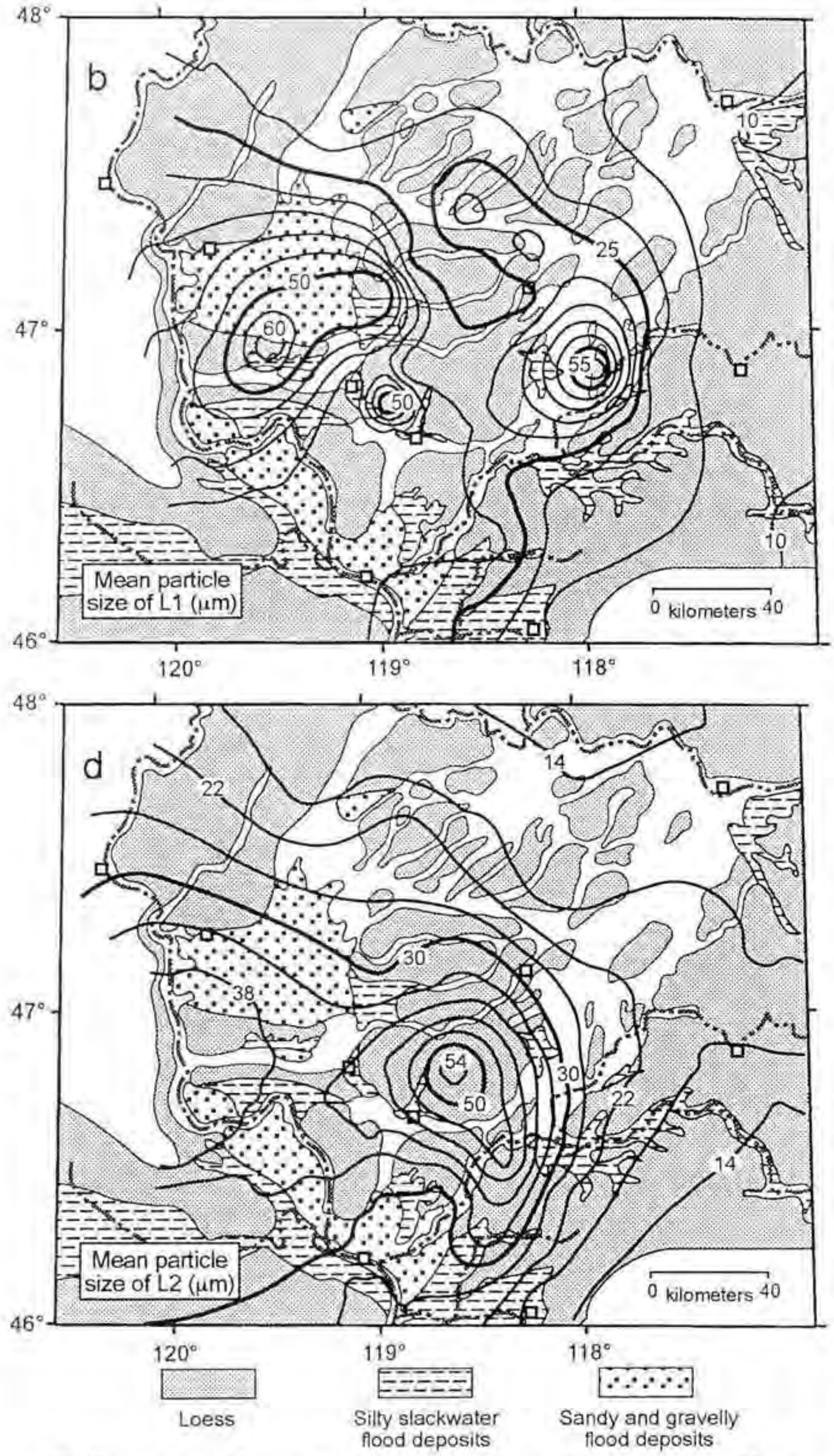


Figure 4. Contour maps of loess thickness and mean particle size. **4a**, Contour map of thickness of L1 loess; bold contours are 100-cm intervals, thin contours are 50-cm intervals. **4c**, Contour map of thickness of L2 loess; bold contours are 100-cm intervals; thin contours are 50-cm intervals. See facing page for remainder of this figure. All geographic and cultural features mentioned in text are shown in Figure 3. AC, Alpowa Creek; CC, Crab Creek; CO, Connell; CP, Cheney–Palouse Scabland; CR, Columbia River; GC, Grand Coulee; MC, Moses Coulee; OT, Othello; PA, the Palouse; PB, Pasco Basin; QB, Quincy basin; RI, Ritzville; SP Spokane; SR, Snake River; ST, Staircase Rapids; WP, Waterville Plateau; WW, Walla Walla; YV, Yakima Valley.



4b, Contour map of mean particle size of L1 loess; bold contours are 25-μm intervals; thin contours are 5-μm intervals. **4d**, Contour map of man particle size of L2 loess; bold contours are 20-μm intervals; thin contours are 4-μm intervals. See facing page for remainder of this figure.

The areas of thickest L1 loess near Pasco and Walla Walla in the lower end of the Columbia Plateau scabland system have a geometric mean size of only about 25 to 30 μm (Fig. 4b): floods apparently sorted the suspended fluvial sediments as they were carried in flood coulees to the west and southwest across the plateau, leaving sandy and gravelly materials in the center of the plateau and Quincy Basin while carrying generally more silty sediments to the Pasco Basin and Walla Walla Valley. In the latter areas, they generated the thick deposits of finer textured L1 loess. The progressive easterly decrease in mean size of L1 loess on the south side of the Snake River further confirms up-valley dust-transporting winds along the Snake River.

L2 Loess

Our analyses of loess layer L2, summarized in Figures 4c and 4d, also demonstrate clear regional trends in layer thickness and mean particle size. L2 loess is a stratigraphic unit deposited from perhaps 50,000 or 55,000 years ago to about 25,000 years ago. On the basis of its consistent relations to a series of paleoflood indicators, including an underlying regional angular unconformity (not shown in Fig. 2), a variety of Touchet Bed-like sediments, and other features, we had postulated as early as 1987 that its deposition was initiated by a middle or early Wisconsin episode of scabland flooding (McDonald, 1987; McDonald and Busacca, 1988, 1990). If the L2 and L1 loess layers had essentially the same origin, we would expect their regional thickness and particle-size patterns to be very similar.

The regional thickness pattern of L2 loess is strikingly similar to that of L1 loess over much of the Columbia Plateau (compare Figs. 4a and 4c). L2 loess attains maximum thicknesses of more than 750 cm to the northeast of the Walla Walla Valley and about 600 cm northeast of the Pasco Basin. The long arm of L2 loess more than 200 cm thick along the Cheney–Palouse Scabland and the bowing of thickness contours to the northeast of Walla Walla and into the main Palouse area are also similar to those of L1. These patterns support the interpretation that the main source areas of L2 loess were the structural valleys of the southern plateau and that the dominant wind direction was southwesterly during L2 time. There is no recognizable thickening of L2 to the east of the Quincy Basin as there is for L1 loess.

Cataclysmic flooding during early Wisconsin time seems the only plausible means of bringing vast quantities of quartzo-feldspathic sediments into the Walla Walla Val-

Figure 5. View of roadcut Clyde-2 (McDonald and Busacca, 1990) where the greatest measured cumulative thickness of L1 and L2 occurs. The distance from the top of the exposure to the roadway is approximately 15 m. The Washtucna Soil at the top of L2 is represented by the two white petrocalcic (caliche) horizons. Mount St. Helens set C tephra (not visible) was found near the base of L2 about 2 m above the roadway.

ley to serve as the source of L2 loess. We do not believe that "normal" glacial outwash sediment, if it had been transported in an unblocked Columbia River system, would have been deposited in the Walla Walla Valley, or even if this were so, not in the quantity required to produce the observed volume or geographic pattern of L2 loess. There is evidence of older slackwater sediments in the Walla Walla and Snake River systems (and indeed across the Columbia Plateau; McDonald and Busacca, 1988) and of their age. Probable deposits of reworked lower Wisconsin slackwater sediments that are covered by younger sediments have been reported from the Walla Walla area (Vrooman and Spencer, 1990), and remnants of slackwater sediments at the confluence of the Snake River and Alpowa Creek more than 150 km upstream from the Columbia (Fig. 3) contain the middle or lower Wisconsin Mount St. Helens set C tephra (Busacca, unpub. data). The sediments at Alpowa Creek that contain the Mount St. Helens set C tephra are overlain by local basaltic colluvium and in turn by upper Wisconsin flood sediments that contain the Mount St. Helens set S tephra (Foley, 1976). The fact that lower Wisconsin slackwater sediments are scarce and difficult to identify in the Walla Walla Valley today, despite our assertion that they were once extensive enough to have spawned the L2 loess, probably results from extensive erosion by wind and water in the millennia following their emplacement, further erosion by late Wisconsin floods, and finally, deep burial by upper Wisconsin flood sediments.

The regional pattern of geometric mean particle size of L2 loess is quite different from that of L1 loess (compare Figs. 4b and 4d). The coarsest L2 loess, with a mean size of about 50 μm, is centered north of Connell in an area of moderately thick L2 loess. This coarse loess is not near any recognized body of lower Wisconsin sandy flood sediment, but Connell is at the confluence of several major flood channels, including Washtucna and Esquatzel Coulees, that may have been loci of early Wisconsin flooding.

Mean particle size of L2 decreases regularly with distance to the northeast across the plateau (Fig. 4d). The absence of coarse L2 loess in the Quincy Basin area contrasts strongly with this pattern for L1 loess (compare Figs. 4b and 4d). It suggests that the northwestern part of the Channeled Scabland, including the Quincy Basin, did not receive sediment during the early Wisconsin episode of flooding. This is consistent with the belief of some researchers that the Grand Coulee may not have been formed until the late Wisconsin (Waitt and Thorson, 1983; Baker and Bunker, 1985). At the least, it appears to have been inactive. A less extensive ice advance than that of the late Wisconsin may not have blocked the Columbia River at the valley of the Okanogan River. Without this blockage, much of the water from outburst floods could travel down the Columbia River to the Pasco Basin without spilling onto the northwestern part of the Channeled Scabland.

CONCLUSIONS AND DISCUSSION

Extensive sampling of two major loess depositional units spanning roughly the last 50,000 years has allowed the first detailed reconstructions of regional patterns of loess accumulation on the Columbia Plateau. The reconstructions are based on stratigraphic intervals that were defined with respect to marker paleosols and tephras. The greatest cumulative thickness of L1 and L2 is nearly 1,200 cm in an area north of the Walla Walla Valley (Fig. 5); both units thin with increasing distance to the northeast to less than 200 cm on the northern and northeastern margins of the plateau (McDonald and Busacca, 1990, 1992). These thickness patterns clearly demonstrate that the Pasco Basin and Walla Walla Valley were the major points of origin of the sediment that contributed to L1 and L2 loess. We believe that immense quantities slackwater sediments were left in these basins by cataclysmic floods during early as well as late Wisconsin time and that these sediments were the source of L2 and L1 loess, respectively. Smaller bodies of flood sediments in scabland coulees and the Quincy Basin also have been important secondary sources of sediment that locally thicken or coarsen L1 and L2 loess. Spatial patterns of thickness and mean particle size do not support the hypotheses that the Ringold Formation, the front of the Cordilleran ice sheet, or alluvium in the Columbia valley upstream of the Pasco Basin were sources of more than minor amounts of loess during the last 50,000 years.

In late Wisconsin and Holocene time, the coarsest loess was derived from sandy flood sediments in the Quincy Basin and in northern and eastern coulees of the scabland system, whereas in early and middle Wisconsin time coarse loess seems to have been derived from coulees in the south-central part of the scabland system. The differences in spatial patterns of thickness and particle size associated with individual loess depositional units may be unique to loess of the Palouse region. Clearly the anastomosing channel systems of the Channeled Scabland functioned in a very different way than did river valleys such as the Mississippi, which served as fixed line sources of sediment for dust transport and loess accumulation (for example, Frazee and others, 1970).

Further analysis of this data set may provide more refined reconstructions of prevailing wind directions though late Pleistocene time. For example, it is not yet clear why loess apparently was not generated at the front of the Cordilleran ice sheet (except perhaps at the Waterville Plateau) when a split jet stream and glacial anticyclonic winds (Manabe and Broccoli, 1985; Anderson and others, 1988) might have carried dust to the south. Perhaps this occurred, but the effect was short lived relative to the time scale of our loess units. More measurements of loess thickness and grain size along the perimeter of the Columbia Plateau (Fig. 3) clearly would be helpful in defining localized patterns from secondary sources. It would perhaps also be fruitful to subdivide units L1 and L2 into smaller stratigraphic intervals to analyze trends in loess distribution over shorter increments of time.

A hypothesis has arisen from our observations of the consistent stratigraphic relations between major buried soils and episodes of flooding: When and whether soils developed in the loess appears to have been dominantly a function of loess deposition rate because that regulated the time available for soil formation on each volume of the accumulating sediment, and perhaps only secondarily a function of climate or vegetation (McDonald and Busacca, 1990). At least for the last two episodes of scabland flooding, several lines of evidence, including the information presented here, indicate that high rates of loess deposition were initiated by the introduction of sediments from cataclysmic floods into the path of prevailing winds. High deposition rates probably continued for thousands of years until the sediment source was substantially exhausted or physically stabilized. Minor soils such as the Old Maid Coulee in the lower part of L2 seem to indicate that there were brief fluctuations in deposition rate. Only during the waning stage of deposition, however, did soil development become dominant over depositional processes. For example, the Washtucna Soil at the top of L2 apparently did not begin forming until more than 20,000 years after the early Wisconsin episode of flooding. The relatively insignificant development of the surface soil compared to the Washtucna or Devils Canyon Soils may indicate that loess deposition rate today is only beginning to slow (excluding the probable impact of farming in the last 100 years), more than 13,000 years after the end of the late Wisconsin floods.

Our work begins to resolve questions of loess source only for the two most recent of dozens of loess units in a geologic system that may have been operating for as much as 2 million years. The origin of the loess in deeper, older units awaits further work. There have been at least seven episodes of cataclysmic floods, however, back to the middle Pleistocene (McDonald and Busacca, 1988), and the zone of maximum *total* loess thickness extends to the northeast of the Walla Walla Valley and into the main Palouse (Busacca and McDonald, unpub. data; Ringe, 1970). These facts, taken together with the results presented here, suggest that slackwater sediments from cataclysmic floods served as the source for much of the Palouse loess through the Pleistocene.

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Chronology of Pre-Late Wisconsin Pleistocene Sediments in the Puget Lowland, Washington

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ABSTRACT

The Quaternary history of the Puget Lowland is marked by three early Pleistocene to late Pliocene glaciations (the Orting, Stuck, and Salmon Springs, separated by interglacial periods represented by the Alderton and Puyallup Formations) and two pre-late Wisconsin glaciations (the Double Bluff and Possession, separated by an interglacial period represented by the Whidbey Formation). Laser-argon, fission-track, thermoluminescence (TL), amino-acid, and paleomagnetic dating techniques now allow the beginning of a firm chronology for sediments of these glacial and interglacial periods.

The Orting Drift is reversely magnetized and lies beneath sediments laser-argon dated at about 1.6 Ma. Considering that the beginning of the Matuyama Reversed Polarity Chron occurred at 2.4 Ma, the age of the Orting Drift is considered to be between 1.6 and 2.4 Ma. The Alderton Formation is reversely magnetized, and laser-argon dating of volcanic ash interbedded with mudflows and fluvial and lacustrine sediments gave an age of about 1.6 Ma. Drift of the Stuck Glaciation has so far yielded no datable material. The Puyallup Formation is reversely magnetized, and laser-argon dates of 1.69 ± 0.11 Ma on plagioclase and 1.64 ± 0.13 Ma from hornblende have been obtained from pumice in fluvial sediments. The Lake Tapps tephra, interbedded between two drifts of the Salmon Springs Glaciation, has been fission-track dated at 1.06 ± 0.11 Ma and associated silt is reversely magnetized.

Clay beneath Double Bluff Drift at its type section on Whidbey Island was dated at 289 ± 74 ka by TL analysis, and Double Bluff glaciomarine drift there was dated at 177 ± 38 ka. Clay beneath Double Bluff till near Pebble Beach on Camano Island was TL-dated at 291 ± 86 ka, and laminated (varved?) clay overlying Double Bluff till at Lagoon Point on Whidbey Island was dated at 320 ± 100 ka. These ages correspond reasonably well with amino-acid ages from mollusk shells in Double Bluff glaciomarine drift that range from 111 to 178 ka at the type locality and 150 to 250 ka elsewhere in the central Puget Lowland. Four TL dates from clay in Whidbey Formation interglacial fluvial sediments range from 102 ± 38 to 151 ± 43 ka: 102 ± 38 ka at Lagoon Point; 106 ± 17 ka at Blowers Bluff; 142 ± 10 ka north of West Beach; and 151 ± 43 ka at Point Wilson. These ages compare favorably with amino-acid ages of 97 ± 35 ka, 96 ± 35 ka, and 107 ± 9 ka from shells in marine sediments correlated with the Whidbey Formation on Whidbey Island.

Amino-acid analyses of marine shells in glaciomarine drift at three localities suggest a mean age of 80 ± 22 ka for the Possession Glaciation.

INTRODUCTION

Building the stratigraphic framework of Pleistocene sediments in the Puget Lowland (Fig. 1) began at the turn of the century with the investigations of Willis (1898) and Bretz (1913). Willis recognized drift of two glaciations, the Vashon and Admiralty, separated by sediments of the Puyallup Interglaciation. Hansen and Mackin (1949) demonstrated the occurrence of more than one pre-Vashon glaciation when they recognized interglacial sediments between two tills beneath Vashon till on southern Whidbey Island. Deposits of four glaciations were mapped in the southern Puget Lowland on the basis of their relative stratigraphic positions by Crandell and others (1958), and deposits believed to belong to the two youngest of these glaciations were subsequently mapped throughout the southern and central Puget Lowland.

The stratigraphic framework of Crandell and others (1958) was used throughout the lowland until the mid-1960s when it was redefined and expanded to include several new glacial and interglacial units by Armstrong and others (1965) and, soon after, by Easterbrook and others (1967). Although no ages were available for any of the pre-Vashon units, correlations throughout the region were made on the basis of stratigraphic position and character of the sediments. This stratigraphic scheme prevailed until 1979 when the first set of fission-track dates on tephra (Easterbrook and Briggs, 1979) confirmed what earlier paleomagnetic analyses at Auburn (Othberg, 1973) and at the Salmon Springs type locality (Easterbrook and Othberg, 1976) had suggested—that the Salmon Springs Glaciation was 10 times older than previously thought. Fission-track ages of nearly 1 Ma from the Lake Tapps tephra (Easterbrook and others, 1981), which is interbedded with Salmon Springs Drift, meant that most of the widespread correlations of pre-Vashon deposits with the Salmon Springs Drift throughout the Puget Lowland were almost certainly invalid and that the regional chronology of pre-Fraser deposits was badly in need of revision.

In the following decade, paleomagnetic data and fission-track and amino-acid ages provided a firmer base for

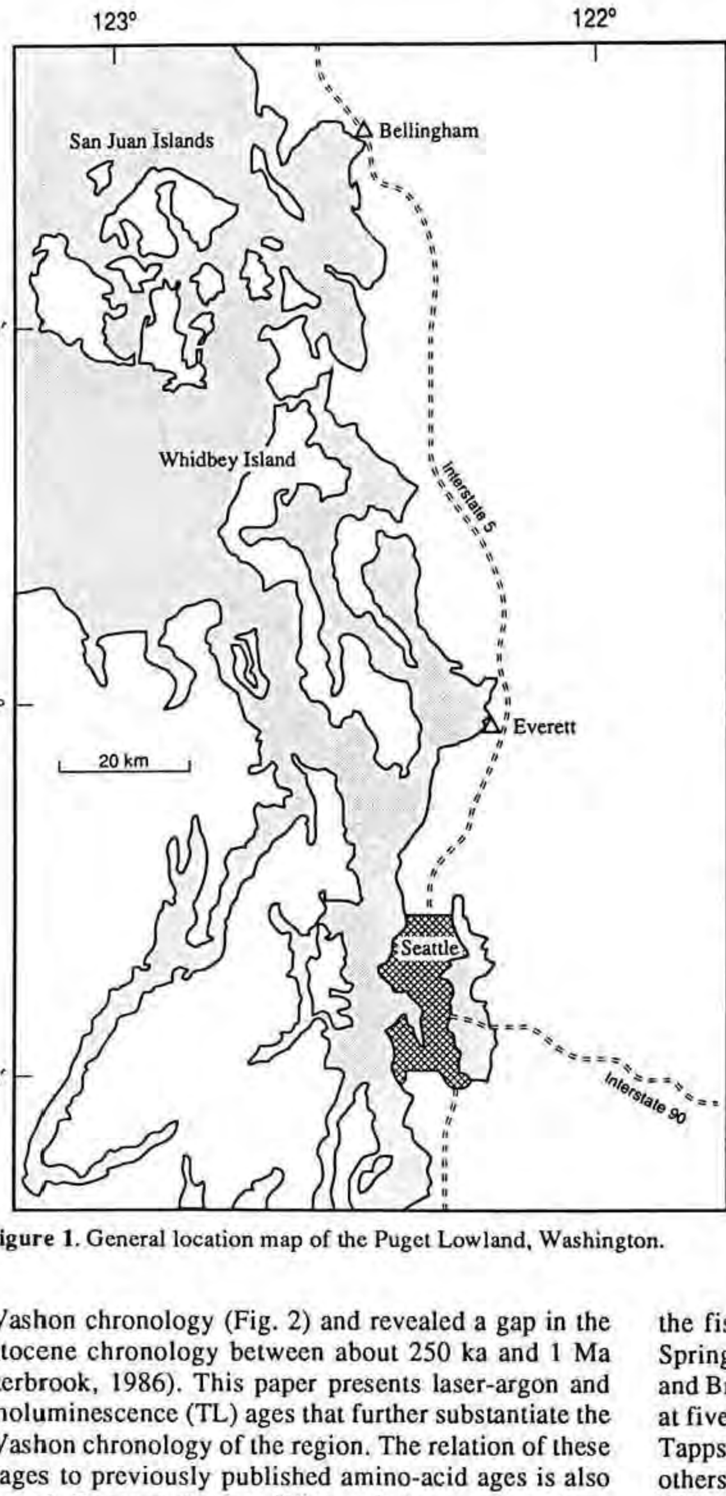

Figure 1. General location map of the Puget Lowland, Washington.

pre-Vashon chronology (Fig. 2) and revealed a gap in the Pleistocene chronology between about 250 ka and 1 Ma (Easterbrook, 1986). This paper presents laser-argon and thermoluminescence (TL) ages that further substantiate the pre-Vashon chronology of the region. The relation of these new ages to previously published amino-acid ages is also discussed. Because the late Wisconsin has recently been discussed in some detail (Easterbrook, 1992), its chronology is not discussed in this paper.

EARLY PLEISTOCENE (0.8–1.7+ Ma) STRATIGRAPHY AND CHRONOLOGY

Introduction

Type localities for the Orting Drift, Alderton Formation, Stuck Drift, Puyallup Formation, and Salmon Springs Drift were defined by Crandell and others (1958) near the southernmost extension of the Cordilleran ice sheet (Fig. 3). When they were defined, no finite ages were known for any of the sediments, and the maximum extent of pre-Wisconsin glaciations was unknown because older drift units have been buried by younger deposits. Virtually all data about early Pleistocene glaciations of the region come from deposits at the southern extent of the Puget lobe. No early Pleistocene deposits of the Juan de Fuca lobe or in the northern or central Puget Lowland have been documented, but they may exist below sea level.

The original definitions of the pre-Vashon glacial and interglacial units in the southern Puget Lowland were based entirely on stratigraphic evidence, primarily stratigraphic position and the provenance of sediments. Glacial deposits were distinguished from nonglacial sediments on the basis of their lithology, which was used to differentiate sediments of the Cordilleran ice sheet from deposits derived from the Cascade Range to the east. Sediments related to Cordilleran ice were characterized by gneiss, schist, quartzite, and distinctive heavy minerals derived from granitic and metamorphic terranes of British Columbia. In contrast, the central Cascade Range east of the southern Puget Lowland typically shed andesite, basalt, diorite, and granodiorite into the adjacent lowland. Fluvial sand and volcanic mudflows derived from Mount Rainier hypersthene-hornblende andesite were deposited in the lowland only when Cordilleran ice was not present.

Without finite dates from pre-Vashon sediments, correlations of the units defined by Crandell and others (1958) were made throughout the Puget Lowland on the basis of stratigraphic position until the 1970s when reversely magnetized silt was discovered near Auburn (Othberg, 1973) and at the Salmon Springs type locality (Easterbrook and Othberg, 1976; Easterbrook and Briggs, 1979; Easterbrook and others, 1981). Shortly thereafter, the fission-track age of tephra interbedded with Salmon Springs Drift was determined to be 0.84 Ma (Easterbrook and Briggs, 1979). Subsequently, this tephra was identified at five localities in the Puget Lowland and named the Lake Tapps tephra (Easterbrook and others, 1981; Westgate and others, 1987). Additional fission-track ages of 0.87 ± 0.27, 0.84 ± 0.21 and 0.89 ± 0.29 Ma were obtained from zircons in Lake Tapps tephra by N. Naeser, and a plateau-annealing fission-track age of 1.0 Ma was measured (Westgate and others, 1987). Reversed magnetic polarities were measured in silt overlying and underlying the Lake Tapps tephra at all five localities, and the tephras were correlated geochemically (Westgate and others, 1987).

These fission-track ages and the reversed paleomagnetism of Salmon Springs Drift invalidated most pre-Vashon correlations elsewhere in the Puget Lowland and

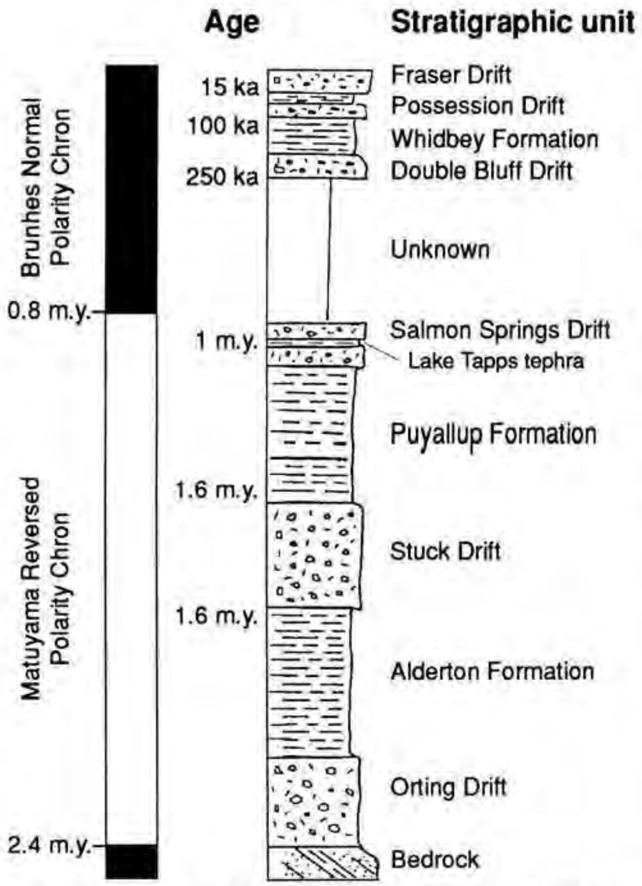

Figure 2. Composite stratigraphic section and ages of sediments in the Puget Lowland.

raised the possibility that even the age and stratigraphic position of pre-Salmon Springs sediments in the Puyallup River valley might be incorrect. However, measurements of the remanent magnetism of the Puyallup Formation, Stuck Drift, Alderton Formation, and Orting Drift at their type localities established that they are all reversely magnetized (Easterbrook and others, 1988), that they were thus most likely deposited during the Matuyama Reversed Polarity Chron between 0.8 and 2.4 m.y. ago.

Orting Drift

The Orting gravel, exposed near the town of Orting in the Puyallup valley (Fig. 3), was named by Willis (1898). Later, Crandell and others (1958) identified till sheets and gravel lenses within the unit and redefined it as Orting Drift. Orting Drift was considered to be the oldest (exposed) Pleistocene glacial deposit in the Puget Lowland on the basis of its intense weathering.

The drift consists of 60–80 m of basal gravel, deposited by streams flowing westward from the central Cascade Range, and overlying till and gravel that was deposited by the Puget lobe. It lies on Tertiary rocks and is unconformably overlain by interglacial sand and mudflows of the Alderton Formation. The till contains 10–15 percent clasts of northern-provenance lithologies and 5–15 percent garnet in the heavy mineral fraction (Crandell, 1963). Because the stratified sand and gravel derived from the central Cascade Range lacks Mount Rainier lithologies, the Orting Drift may predate the volcano.

Paleomagnetism

No sediments suitable for paleomagnetic analysis have been found at the type locality of Orting Drift. However, 34 samples from three sections nearby were reversely magnetized, with a mean declination of 164° and a mean inclination of –31° (Easterbrook and others, 1988).

Age

No age determinations have yet been obtained from the Orting Drift. However, Orting Drift lies stratigraphically beneath the Lake Tapps tephra, fission-track dated at 1 Ma, and beneath sediments laser-argon dated at 1.6 Ma (Easterbrook and others, 1981, 1992; Westgate and others, 1987). Because the drift is both reversely magnetized and older than 1.6 Ma, it was probably deposited during the Matuyama Reversed Polarity Chron and thus younger than 2.4 Ma (the beginning of the Matuyama).

Orting Drift is much more deeply weathered than other pre-Salmon Springs sediments, suggesting that a fairly long period of surface exposure of the drift occurred prior to burial by the Alderton Formation. This leads to the interesting possibility that the Orting may be close to 2 m.y. old and may correlate with other 2- to 2.5-m.y.-old glaciations recognized elsewhere in the world (Boellstorff, 1978; Grube and others, 1986; Zagwijn, 1986).

Alderton Formation

The Alderton Formation was named by Crandell and others (1958) for interbedded mudflows, lahars, alluvium, lake sediments, peat, and volcanic ash exposed along the sides of the Puyallup valley near the town of Alderton (Fig. 3). Mudflows in the Alderton sediments range from 1 to 8 m thick and consist of angular to subrounded pebbles and boulders in a sandy clay matrix. Clasts are dominantly hornblende-hypersthene andesites derived from an ancestral Mount Rainier.

The dominance of Mount Rainier source material and absence of lithologies typical of northern glacial provenance suggest that the Alderton sediments are interglacial (Crandell and others, 1958). Confirmation of an interglacial climate comes from pollen in peat in the lower part of the formation. Engelmann spruce pollen in the lower part of a peat bed represents a Canadian vegetation zone (now at elevations of 1,060 to 1,970 m) typical of a climate somewhat cooler than at present. Large amounts of Douglas-fir and alder pollen higher in this peat are characteristic of the humid transition zone (now at elevations of 0 to 1,060 m) typical of a climate similar to that of the present (Leopold and Crandell, 1958). Pollen from other nearby peat and silt beds suggests deposition under climatic conditions ranging from cool and moist to warmer and drier than the present (Heusser, 1977).

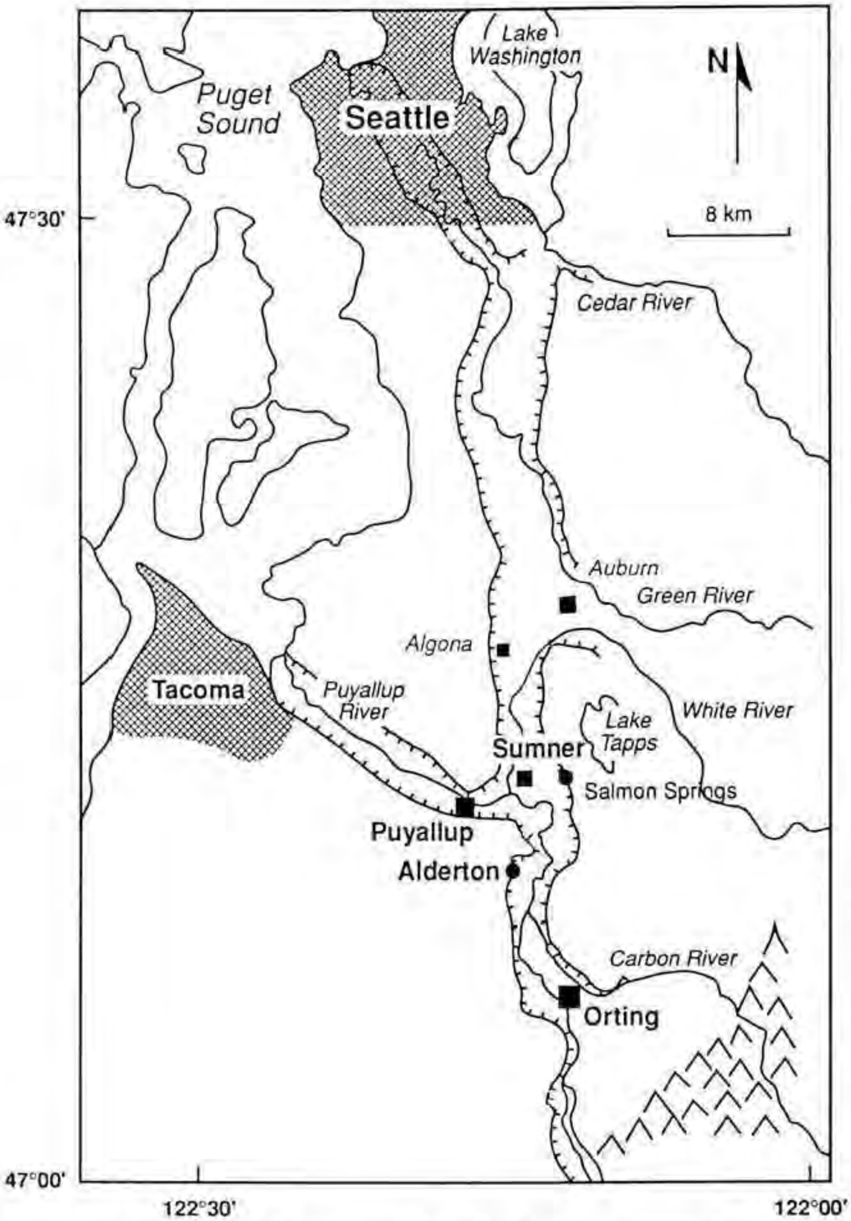

Figure 3. Early Pleistocene sites in the Puyallup River valley.

Paleomagnetism

The remanent magnetism of silt in the Alderton at its type locality is reversed; the average declination is 176°, and the average inclination is 50° (Roland, 1984; Easterbrook and others, 1988), which places its age within the Matuyama Reversed Polarity Chron. The reversed polarity has been overprinted with a viscous normal remanent magnetism that can be removed with high levels of demagnetization. Paleomagnetic measurements on Alderton samples from an exposure 645 m from the type section show an average declination of 170° and inclination of -65° at the 500 oersted (oe) level of demagnetization.

The reversed paleomagnetism of the Alderton Formation and its stratigraphic position below the 1-m.y.-old Lake Tapps tephra places its age within the Matuyama Reversed Polarity Chron, thus placing the age of the Alderton between 1.0 and 2.4 Ma.

Laser-argon age

Volcanic ash interbedded with silt and sand at the type locality of the Alderton Formation occurs less than a meter beneath Stuck Drift (Figs. 4 and 5). Three of five plagioclase crystals that were laser-argon dated by R. Walter (Easterbrook and others, 1992) were clearly detrital because their ages range from 2 to 25 Ma, but two yielded a mean age of about 1.6 Ma. Laser-argon ages are being measured on additional samples to better constrain the age of this tephra and tephras at other sections in the Puyallup valley.

Stuck Drift

Stuck Drift was defined on the basis of oxidized till overlying the Alderton Formation southwest of the town of Alderton (Fig. 3) (Crandell and others, 1958). Stuck Drift at the type locality consists of oxidized till; 10–15 percent of the clasts were derived from a northern provenance, indicating deposition by the Cordilleran ice sheet. The till is overlain and underlain by oxidized sand and gravel. Elsewhere in the Puyallup valley, Stuck Drift consists of two tills separated by 15–50 m of fluvial and lacustrine silt, sand, and gravel (Crandell, 1963; Mullineaux, 1970). The lithology of clasts in this alluvium indicates that they were derived from the Cascade Range and that the Cordilleran ice sheet had retreated far enough north to allow west-flowing drainage from the Cascades to be re-established. Because exposures of Stuck Drift are limited to the Puyallup valley and are not continuous, the two Stuck tills separated by nonglacial deposits could represent two glaciations or a minor fluctuation of the ice margin. Stuck deposits at their type locality are too coarse for paleomagnetic measurement, but sediments south of Sumner, interpreted as Stuck by Crandell (1963), consist of interbedded glaciolacustrine sand and silt. These sediments are overlain by sand and silt of the Puyallup Formation and are underlain by oxidized fluvial gravel. Samples from sediments transitional between Stuck gravel and the Puyallup Formation are reversely magnetized. The mean magnetic declination measured at the 700 oe demagnetization level from samples in the lower part of the sediments transitional to the Puyallup Formation is 183°, and the mean inclination is −19°. Silt in the upper part of these sediments is also reversely magnetized, with a mean declination of 126° and a mean inclination of −51° (Easterbrook and others, 1988).

Figure 4. Stuck Drift overlying silt and volcanic ash laser-argon dated at 1.6 Ma at the type locality of the Alderton Formation.

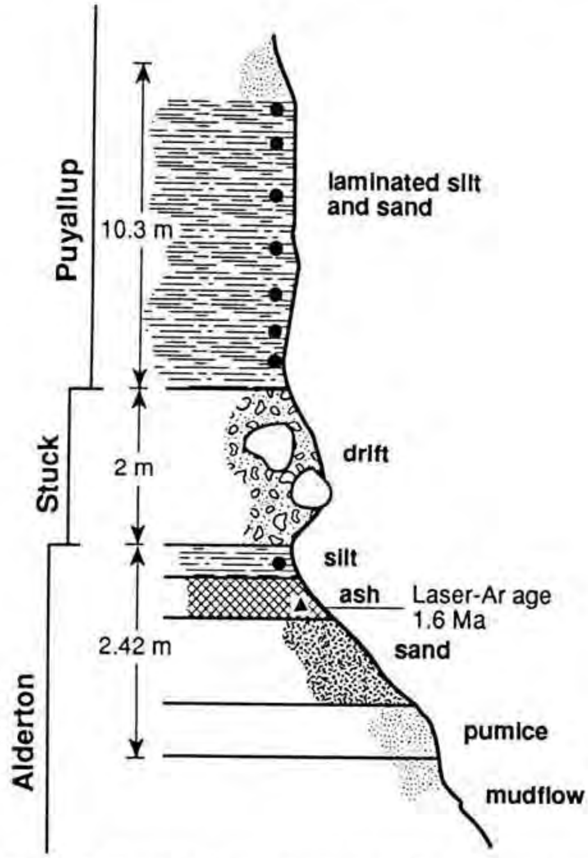

Figure 5. Stratigraphic section at the Alderton Formation type locality. Black dots represent reversely magnetized samples. The laser-argon date of 1.6 Ma was obtained from the ash beneath the Stuck Drift.

On the basis of its stratigraphic position beneath the Salmon Springs Drift (1 Ma) and its reversed remanent magnetism, the Stuck Drift is considered to have been deposited during the Matuyama Reversed Polarity Chron (Easterbrook and others, 1988). This is consistent with a preliminary laser-argon age of 1.6 Ma from the underlying Alderton Formation and 1.64–1.69 Ma from the overlying Puyallup Formation. The overlapping ages may be due to (a) the preliminary nature of the 1.6-Ma age of the Alderton Formation (based on measurements of two grains) or perhaps (b) reworking of the 1.64–1.69 Ma pumice in the Puyallup Formation. Based on the data presently available, the age of the Stuck Glaciation is considered to be close to 1.6 Ma.

Puyallup Formation

Nonglacial sediments in the Puyallup River valley were named Puyallup Sand by Willis (1898), but the term Puyallup was later expanded by Crandell and others (1958) to include lacustrine silt and clay, peat, fluvial sand and gravel, and mudflows derived mostly from ancestral Mount Rainier and renamed the Puyallup Formation. As redefined, the Puyallup Formation includes, but is not limited to, the Puyallup Sand of Willis (1898). It is underlain by Stuck Drift and overlain by Salmon Springs Drift. A well-developed weathering profile characterized by the formation of kaolin at the top of the Puyallup Formation was recognized by Crandell and others (1958), who interpreted it to record a long period of weathering prior to deposition of the overlying Salmon Springs Drift.

At its type locality, the Puyallup Formation consists of 41 m of laminated lacustrine silt and sand, fluvial sand and gravel, as well as mudflows, interbedded in places with tephra and peat. Ninety-five percent of the clasts in the mudflows were derived from Mount Rainier, as were many of the gray, unoxidized sand beds in the unit. Some of the fluvial sand and gravel contains a mixture of Mount Rainier and central Cascade mountains lithologies. Puyallup sediments, like those of the Alderton Formation, were deposited by streams flowing from Mount Rainier and the nearby Cascade Range across the southeastern Puget Lowland.

Pollen analyses of peat beds in the lower part of the Puyallup Formation suggest early postglacial forest growth typical of the upper Canadian vegetation zone, followed by a warmer climate sustaining forest growth of the middle

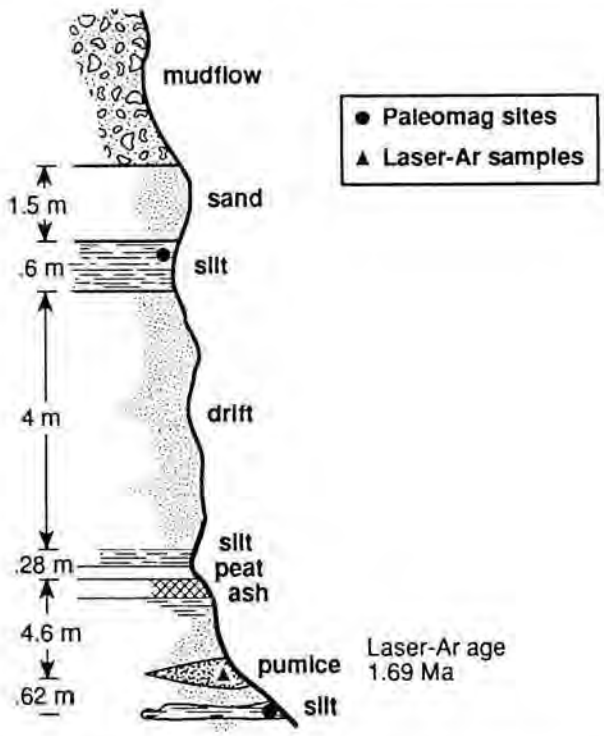

Figure 6. Stratigraphic section at the Corless gravel pit south of the Stuck type locality near Sumner. Laser-argon ages of 1.69 ± 0.11 Ma and 1.64 ± 0.13 Ma were measured on pumice clasts in fluvial sand near the base of the section.

Canadian vegetation zone (Crandell, 1963). Pollen samples from nearby peat beds higher in the section suggest a progressive climatic warming to conditions similar to those of today, followed by a cooling trend. However, the pollen record is incomplete because poor exposures prevent lateral correlation of sampled units and because an unknown amount time is represented by weathering to form the paleosol at the top of the Puyallup Formation.

Paleomagnetism

Paleomagnetic measurements from part of Crandell's (1963) type section of the Puyallup Formation and from nearby exposures indicate reversed magnetic polarity with a complex post-depositional magnetic overprint (Easterbrook and others, 1988). Most of that section is now covered with vegetation, but samples were taken from the 3 m of laminated, pink-gray silt and fine sand that remains exposed. Consistent, stable, single-component paleomagnetic directions could not be satisfactorily isolated during step demagnetization of samples from the type locality because of a strong, high-coercivity, normal-polarity magnetic overprint that masked the primary magnetization (Easterbrook and others, 1988). However, sediments from a laterally equivalent section in the Corless gravel pit along the wall of the Puyallup valley near Sumner, less than 200 m south of the type locality, were found to be reversely magnetized with a mean magnetic declination 183° and a mean inclination of –42° at the 700 oe demagnetization level. The reversed magnetic polarity, together with its stratigraphic position beneath the Lake Tapps tephra, places the Puyallup Formation within the Matuyama Reversed Polarity Chron.

Laser-argon age

Pumice collected from the Corless gravel pit was laser-argon dated by R. Walter (Easterbrook and others, 1992). The pumice occurs as baseball-size detrital clasts in fluvial sand beneath wood-bearing volcanic ash interbedded with reversely magnetized silt (Fig. 6); this silt is correlated with the Puyallup Formation by Roland (1984). Because the pumice clasts are clearly detrital, the possibility that they have been reworked from Alderton sediments cannot be discounted. However, the pumice clasts are concentrated in a single sand lens, and their surfaces show no evidence of having once been enclosed in a different sediment.

Laser-argon dating of four plagioclase crystals from the pumice gave a mean age of 1.69 ± 0.11 Ma, and five hornblende crystals yielded a mean age of 1.64 ± 0.13 Ma (Easterbrook and others, 1992). These ages are consistent with the age implied by the reversed magnetic polarity measured in the Puyallup Formation, but overlap the 1.6 Ma preliminary laser-argon date of the Alderton Formation. Additional laser-argon measurements of the Alderton tephra should increase the accuracy of the 1.6 Ma age and may resolve the apparent overlap. However, considering the relatively short duration of the late Wisconsin Fraser Glaciation, the Stuck Glaciation could also have been relatively short-lived, and the Alderton and Puyallup Formation could be close to the same age.

Salmon Springs Drift

The Salmon Springs Drift was defined by Crandell and others (1958) on the basis of till and outwash gravel in the Puyallup valley east of Sumner. It includes a lower glacial outwash unit with interbedded lenses of till; this is overlain by about 1 m of lacustrine silt, the Lake Tapps tephra (Easterbrook and others, 1981; Westgate and others, 1987), and peat, which is in turn overlain by outwash sand and gravel (Fig. 7).

Contacts between the tephra, silt, and peat are gradational, indicating that deposition of these sediments was continuous. Pollen from the ash-silt-peat sequence indicates a tundra vegetation changing progressively to a fir forest before burial by upper Salmon Springs gravel (Leopold and Crandell, 1958; Crandell and others, 1958; Crandell, 1963; Heusser, 1977).

The Lake Tapps tephra is a calc-alkaline, dacitic to rhyolitic volcanic ash containing phenocrysts of plagioclase, hornblende, hypersthene, biotite, magnetite, ilmenite, apatite, and zircon (Easterbrook and others, 1981; Westgate and others, 1987). Glass in the tephra consists of angular, chunky shards with few vesicles and pumiceous shards with lineated tubular vesicles. Detailed geochemistry of the tephra at five localities by Westgate (Tables 1 and 2)

(Westgate and others, 1987) demonstrates that all are the same tephra, making it a highly useful marker bed for correlation.

Paleomagnetism

Measurements of the remanent magnetism of the silt bed directly above Lake Tapps tephra at the Salmon Springs type locality showed both reversed and normal polarities. However, paleomagnetic profiles of silt interbedded with the Lake Tapps tephra at four localities in the Puyallup valley and one on the Olympic Peninsula all show stable, reversed polarities (Easterbrook and others, 1981, 1988; Westgate and others, 1987; Easterbrook, 1986). The reversed magnetic polarity of sediments above and beneath the Lake Tapps tephra indicate that the tephra was deposited during the Matuyama Reversed Polarity Chron.

Figure 7. Silt bed containing the Lake Tapps tephra at the Salmon Springs type locality. Outwash gravel underlies the ash in the lower part of the photo, and peat overlies the silt bed in the upper left corner of the photo.

Age

The Lake Tapps tephra has been positively identified at four additional localities (Tables 1 and 2) (Westgate and others, 1987). Six fission-track dates from four of these sites, including the type locality, are shown in Table 3.

The fission-track ages on glass at Auburn and Algona are younger than the zircon ages, presumably because of track annealing in the glass. The fission-track age of 1.06 ± 0.11 Ma on glass at Algona, corrected for partial track fading (Westgate and others, 1987), is thought to be the most accurate of the dates. Thus, the age of the Lake Tapps tephra, and hence the Salmon Springs Drift, is now considered to be close to 1.0 Ma (0.95–1.17 Ma).

Although Salmon Springs Drift has been widely mapped in the southern and central Puget Lowland on the basis of sections containing glacial deposits beneath Vashon Drift, the only places where it has been positively identified by radiometric dating and paleomagnetic analysis are in the Puyallup valley and on the Olympic Peninsula. Without such data, the Salmon Springs cannot be definitely recognized. Most of the drift previously mapped in the central and southern Puget Lowland as Salmon Springs Drift is probably the much younger Double Bluff Drift or Possession Drift.

MIDDLE PLEISTOCENE (800–300 ka) STRATIGRAPHY AND CHRONOLOGY

The "middle Pleistocene" has been defined as the period between 300 ka and 800 ka (Richmond and Fullerton, 1986). No sediments of this age have been recognized in the Puget Lowland. They may be present below sea level in the Puget Lowland, but are not accessible and have not been unequivocally identified in cores. Pre-Vashon glacial deposits have been found in sea-cliff sections and in wells, but without positive age determination they cannot be definitely correlated with known stratigraphic units. Pre-Double Bluff drift occurs in the southern lowland, but, except for exposures in the Puyallup valley, its age is not known. Consequently, no deposits spanning the time gap between the Double Bluff Drift and the Salmon Springs Drift have been recognized.

LATE PLEISTOCENE (300–10 ka) STRATIGRAPHY AND CHRONOLOGY

Introduction

The "late Pleistocene" has been defined as the interval between 10 ka and 300 ka (Richmond and Fullerton, 1986). Pleistocene sediments whose ages range between 100 ka and 300 ka are widespread in the central Puget Lowland,

Table 1. Major-element microprobe analyses of glass shards in Lake Tapps tephra (weight percent) (Westgate and others, 1987)

	Salmon Springs UT88	Peasley Canyon UT400	Frigid Creek UT55	Frigid Creek UT57	Frigid Creek UT58
SiO_2	78.6	78.1	78.5	78.2	78.2
TiO_2	0.20	0.18	0.18	0.22	0.19
Al_2O_3	12.3	12.9	12.5	12.6	12.8
FeO	0.69	0.73	0.71	0.87	0.81
CaO	0.81	0.83	0.80	0.76	0.79
Na_2O	3.6	3.5	3.5	3.5	3.3
K_2O	3.7	3.7	3.7	3.7	3.8
H_2O	7.5	6.4	7.3	7.1	6.7

Table 2. Neutron activation analyses of trace and rare earth elements in Lake Tapps tephra (ppm) (Westgate and others, 1987)

	Salmon Springs UT88	Auburn UT52	Algona UT462	Peasley Canyon UT400	Frigid Creek UT55	Frigid Creek UT56	Frigid Creek UT57	Frigid Creek UT58
La	19.8	19.5	20.5	18.8	19.2	19.8	19.5	19.8
Ce	38.1	35.9	39.1	35.6	37.5	37.6	38.1	38.0
Nd	8.8	9.0	10.6	9.2	11.3	11.7	11.0	11.0
Sm	2.08	1.89	2.09	2.50	1.89	1.94	1.94	2.04
Eu	0.57	0.57	0.53	0.67	0.57	0.56	0.53	0.61
Tb	0.29	0.28	0.28	0.34	0.25	0.25	0.22	0.29
b	1.62	1.47	1.56	1.26	1.60	1.50	1.57	1.83
Lu	0.26	0.22	0.25	0.26	0.23	0.23	0.24	0.23
U	3.88	3.53	3.60	3.68	3.58	3.76	3.70	3.70
Th	9.08	8.06	8.60	8.62	8.28	8.59	8.48	8.63
Ba	1105	1120	1130	920	1115	1180	1175	1120
Rb	66	67	74	54	73	71	75	74
Cs	0.82	1.01	1.01	0.78	0.91	1.02	0.96	1.10
Sc	2.10	1.30	1.38	3.51	1.42	1.45	1.75	1.86
Hf	3.50	2.97	3.19	3.60	3.03	2.91	2.89	3.11
Ta	0.95	0.86	0.89	0.95	0.84	0.89	0.86	0.89
Mo	4.65	4.96	5.35	2.89	5.73	5.65	5.28	5.81
As	5.47	4.77	5.26	5.18	4.54	4.87	5.39	4.56
Zn	43	37	35	37	35	40	50	40
Co	2.09	1.38	1.32	4.31	1.37	1.53	2.11	2.76

but they have not been unequivocally identified and dated elsewhere in the lowland.

The late Pleistocene stratigraphy and chronology of the Puget Lowland may be considered in terms of the pre-late Wisconsin interval and the late Wisconsin. Because the late Wisconsin has recently been discussed in considerable detail (Easterbrook, 1986, 1992), the reader is referred to these publications for information and data, and discussion here is limited to the pre-late Wisconsin interval. The chronology of pre-late Wisconsin sediments is based on amino-acid and TL dating. Details concerning amino-acid analyses are given in Blunt and others (1987). The type sections of pre-late Wisconsin late Pleistocene sediments are located on southern Whidbey Island (Fig. 8).

Double Bluff Drift

The Double Bluff Drift (Easterbrook and others, 1967; Easterbrook, 1968) consists of till, glaciomarine drift, and glaciofluvial and glaciolacustrine sediments lying stratigraphically beneath the Whidbey Formation of the last major interglaciation. Double Bluff Drift crops out at or near sea level in sea-cliff exposures in the central Puget Lowland, but underlying sediments are not exposed above sea level.

Table 3. Fission-track ages of Lake Tapps tephra (Easterbrook and others, 1981; Westgate and others, 1987). m.y., million years; * corrected for partial track fading

Age (m.y.)	Standard deviation (±1)	Locality	Material dated	Sample no.
1.06	0.11	Algona	Glass*	UT462
0.65	0.08	Algona	Glass	UT462
0.84	0.22	Salmon Springs	Zircon	UT88
0.90	0.15	Frigid Creek	Glass	UT55
0.87	0.30	Auburn	Zircon	UT52
0.66	0.04	Auburn	Glass	UT52

Few exposures of Double Bluff Drift in the central Puget Lowland have been recognized. It crops out in bluffs at or near sea level at Double Bluff, Possession Point, Lagoon Point, and Ebeys Landing on Whidbey Island, at Foulweather Bluff on the Kitsap Peninsula, in bluffs about 1 km west of Point Wilson on the Olympic Peninsula, and in bluffs 1 km southeast of Pebble Beach on Camano Island. Elsewhere, the Double Bluff Drift is not exposed above sea level, is buried beneath younger deposits, or has been eroded away.

Exposures along the sea cliffs at Double Bluff were designated by Easterbrook and others (1967) as the type section of the Double Bluff Drift. About 6 m of gravel is overlain by 3–4 m of sand, silt, and clay and about 12 m of compact gray till and another diamicton interbedded with sand and silt (Figs. 9 and 10). About 9 m of cross-bedded, fairly well sorted sand underlies the lower gravel unit at the extreme northwest end of the bluff. Clasts of pink granite and garnet-kyanite schist in the till and underlying gravel indicate that the source of the drift was British Columbia. On the basis of lithology and texture, the lower gravel is interpreted as proglacial outwash, later overridden by the ice that deposited the till.

Toward the southeast end of Double Bluff, the drift consists of a clayey diamicton interbedded with silt and pebbly silt (Figs. 9 and 11). Marine shells in the diamicton indicate subaqueous deposition as glaciomarine drift from floating ice.

The top of the Double Bluff Drift is exposed in the low bluffs about one-half kilometer east of the southernmost point at Double Bluff. Here, pebbly-silty glaciomarine drift is overlain by about 3 m of oxidized sand and gravel that passes eastward beneath a thick section of peat-bearing sand and silt of the Whidbey Formation.

Paleomagnetism

Paleomagnetic measurements of Double Bluff glaciomarine drift indicate normal declination and inclination, with an average declination of 1° and an average inclination of 49° (Easterbrook, 1976). The till and outwash deposits contain no material suitable for paleomagnetic measurements.

These paleomagnetic data are not as useful for determining age limits as for the early Pleistocene sediments because the Earth's magnetic field has experienced normal polarity for the past ~750–780 ka. However, the paleomagnetic data are useful in demonstrating that normally magnetized sediments cannot correlate with the reversely magnetized deposits of the southern Puget Lowland.

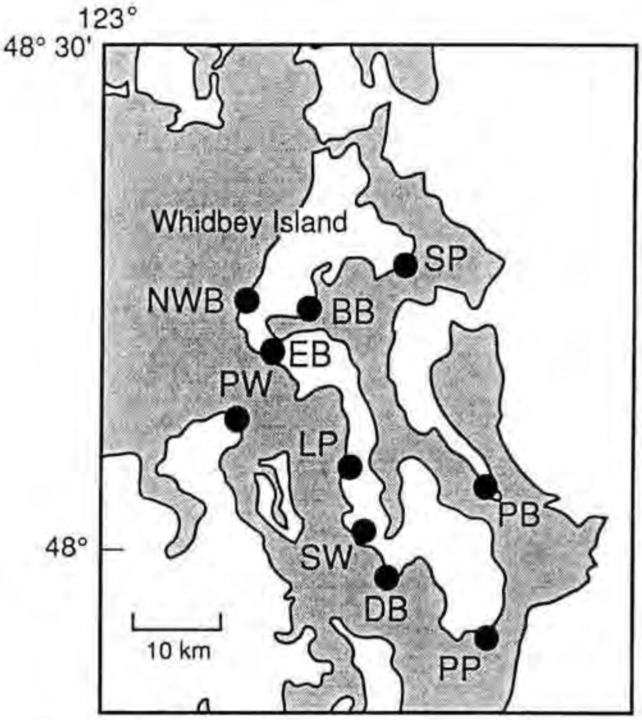

Figure 8. Pre-late Wisconsin sites in the central Puget Lowland. DB, Double Bluff; NWB, North West Bluff; BB, Blowers Bluff; PW, Point Wilson; LP, Lagoon Point; SW, South Whidbey Park; PB, Pebble Beach; PP, Possession Point.

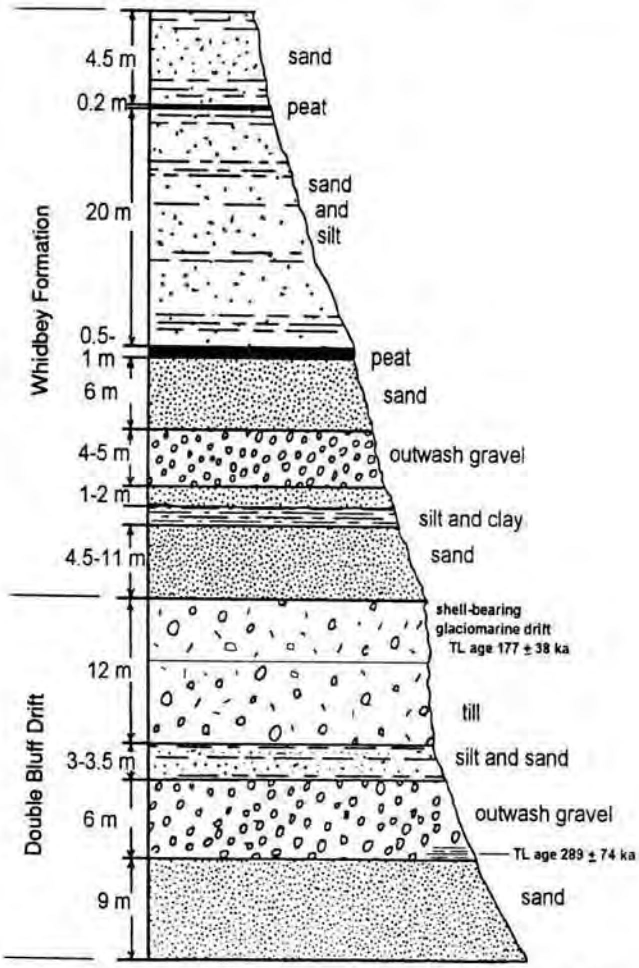

Figure 9. Composite stratigraphic section at the type locality of the Double Bluff Drift. A TL age of 177 ± 38 ka was obtained from glaciomarine drift above the till, and an age of 289 ± 74 ka was obtained from clay in outwash gravel beneath the till.

Amino-acid ages

An age range of 111 ka to 178 ka is suggested by leucine D/L ratios of amino acids in shells from glaciomarine drift at the type locality (Fig. 11) (Blunt and others, 1987). Amino-acid age ranges of 150–250 ka have been determined from shells in Double Bluff glaciomarine drift elsewhere in the Puget Lowland (Easterbrook and Rutter, 1982).

Thermoluminescence ages

TL ages of pre-late Wisconsin sediments in the Puget Lowland have been measured with varying degrees of success (Berger and Easterbrook, 1989; Easterbrook and others, 1992). Although further measurements are desirable to confirm the initial results, TL ages of Double Bluff Drift have been obtained for several localities. TL analysis of clay in glaciomarine drift near the top of the section at the southeast end of the Double Bluff type section (Fig. 11) gave an age of 177 ± 38 ka (Easterbrook and others, 1992); this age is consistent with amino-acid ages from shells in the same part of the glaciomarine drift. At the western edge of the type section, a TL age of 289 ± 74 was obtained from a clay lens at the

Figure 10. Stratigraphic section at the northwest end of the type locality of the Double Bluff Drift (described in Fig. 9). The sample for the 289 ± 74 ka age was collected at the base of the bluffs about 1 m above the beach.

Figure 11. Stratigraphic section at the northwest end of the type section of the Double Bluff Drift. The sample for the 177 ± 38 ka TL age was collected from glaciomarine drift 1 m above beach level. Shells for amino-acid analyses came from the same unit.

Figure 12. Double Bluff till overlain by Whidbey peat-bearing silt and sand and underlain by clay and pumiceous sand at Pebble Beach, Camano Island.

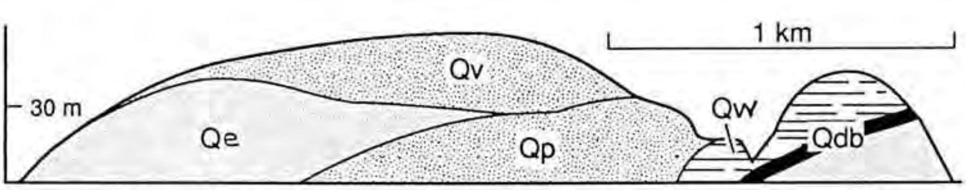

Figure 13. Geologic cross section of Pleistocene sediments at Lagoon Point, Whidbey Island. Qdb, Double Bluff Drift; Qw, Whidbey Formation; Qp, Possession Drift; Qe, Esperance sand; Qv, Vashon till. South is to the right in this section.

base of the bluffs beneath the till-glaciomarine drift unit (Fig. 10). South of Pebble Beach on the southwestern tip of Camano Island, till that is overlain by peat-bearing sand, silt, and clay of the Whidbey Formation (Fig. 12) has been correlated with Double Bluff Drift (Easterbrook, 1968). A TL age of 291 ± 86 ka was obtained from clay a few meters beneath the till (Easterbrook and others, 1992).

At the south end of the Lagoon Point section on Whidbey Island (Fig. 13), fluvial sand and silt with lenses of gravel are overlain by Double Bluff till (Easterbrook, 1968), which in turn is overlain by a thick sequence of north-dipping sediments that includes the Whidbey Formation, Possession Drift, Esperance sand, Vashon till, and Everson glaciomarine drift (Fig. 13), making up one of the most complete stratigraphic sections in the central Puget Lowland. Rhythmically laminated (varved?) clay lying on Double Bluff till (Fig. 14) was TL dated at 320 ± 100 ka (Easterbrook and others, 1992). The laminated clay is directly overlain by pumiceous sand, wood-bearing silt, clay, and peat of the Whidbey Formation.

The TL age of glaciomarine drift at the Double Bluff type locality (177 ± 38 ka) is consistent with the amino-acid age range of 111–178 ka at the same site and with the amino-acid age range of 150–250 ka in Double Bluff glaciomarine drift elsewhere. The 289 ± 74 ka TL age from clay near the base of the Double Bluff type section and the 291 ± 86 ka age from sediments beneath Double Bluff Drift at Pebble Beach overlap the 320 ± 100 ka age at Lagoon Point. When the margin of error of these dates is considered, they do not yet allow an accurate assessment of the age of the Double Bluff Drift. Bearing in mind that some of these ages are from fluvial and lacustrine sediments beneath Double Bluff Drift, some are from within the drift, and that the margin of error is fairly high, the possibility that perhaps some of these drifts are older than those at the Double Bluff type locality cannot be discounted at this stage. In view of the margin of error of the TL dates, the age of the Double Bluff Drift is thought to range from about 150 ka to 250 ka.

Laser-argon age

South of Pebble Beach on Camano Island, a few meters of cross-bedded sand between Double Bluff till and the TL-dated clay contains many thin beds composed almost entirely of pumice clasts 0.5–1.5 cm in diameter (Fig. 15). Laser-argon measurements of the pumice gave well-defined ages of 0.98 ± 0.07 Ma on plagioclase and 1.04 ± 0.04 on biotite. The difference between the TL and laser-argon ages remains unresolved. The discrepancy between the TL age of 291 ± 86 ka and laser-argon ages could be explained by the detrital nature of the pumice clasts. The similarity of the age of the pumice to the age of Lake Tapps tephra suggests that the Camano pumice could be reworked Lake Tapps tephra. However, the pumice is considerably coarser than the typical Lake Tapps ash, and its major-element geochemistry does not match that of the Lake Tapps tephra at its type locality at Salmon Springs or at Auburn, Algona, Peasley Canyon, or Frigid Creek (Table 4). Thus, despite their similar ages, the Camano pumice is not likely to be Lake Tapps tephra. The TL age on the interbedded clay is consistent with TL ages of sediments beneath Double Bluff Drift at Double Bluff and Lagoon Point, and the detrital nature of the pumice could mean that it is reworked from older sediments, but the concentration of the Camano pumice clasts argues against reworking of an older tephra.

Figure 14. Rhythmically laminated (varved?) clay, TL dated at 320 ± 100 ka, that lies on Double Bluff till (out of view to the right of the photo) at Lagoon Point, Whidbey Island.

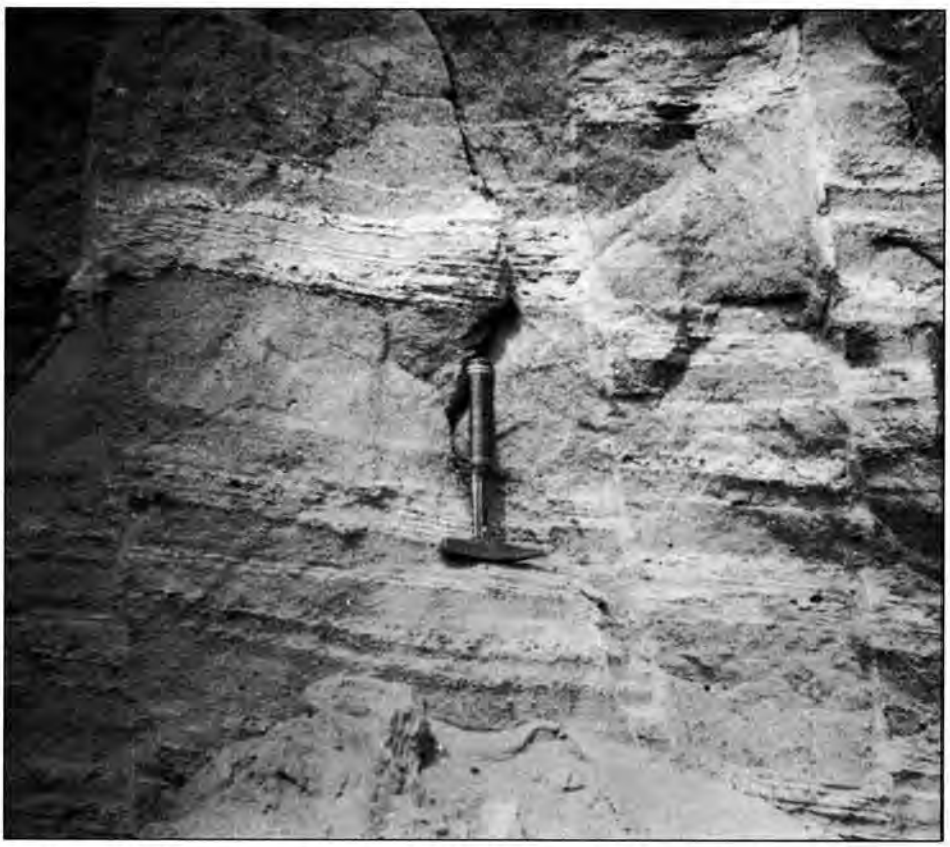

Figure 15. Pumiceous sand beneath Double Bluff till at Pebble Beach, Camano Island. The white layers interbedded with the sand consist entirely of pumice.

Whidbey Formation

The Whidbey Formation crops out extensively in the sea cliffs of Whidbey and Camano Islands and in places along coastlines of the mainland, but the base is exposed only at a few places where the underlying Double Bluff Drift is exposed. Outcrops of the Whidbey Formation are scarce inland because in most places it is covered by younger depos-

Figure 16. Whidbey Formation pumiceous sand, wood-bearing silt, and peat, TL dated at 102 ± 38 ka, Lagoon Point, Whidbey Island. The sample site is about 1 m above beach level just to the right of this photo. The base of the bluffs in the photo is at beach level.

its. The Whidbey Formation is locally overlain by Possession Drift, but elsewhere it is separated from deposits of post-Possession age by an unconformity.

Hansen and Mackin (1949) described an 18- to 24-m-thick section of peat-bearing sand, silt, and clay above "sea level till and outwash" at Possession Point on southern Whidbey Island (Figs. 8 and 14). A similar stratigraphic sequence more than 70 m thick, well exposed in the 2-km-long sea cliff exposure between Double Bluff and Useless Bay, was designated as the type section of the Whidbey Formation (Easterbrook and others, 1967).

The Whidbey Formation consists mostly of sand interbedded with silt, clay, peat, and widely scattered lenses and beds of gravel. Hansen and Mackin (1949) interpreted the lenticular coarse sand and gravel to be channel deposits formed as a result of ". . . very slow aggradation by meandering streams flanked by flood-plain lakes and swamps" in which silty clay and peat were deposited. All sediments included in the Whidbey Formation presumably were formed in such a depositional environment.

Near its type locality, the Whidbey Formation is at least 65 m thick, but an unknown thickness may have been eroded from the top of the section. More than 90 m of stratified sediments are exposed in the sea cliffs east of Double Bluff and also at bluffs 1 km to the east, but the upper 30 m in both places apparently consists of Esperance sand deposited unconformably on the Whidbey Formation.

In the lower part of the Whidbey Formation at Possession Point, Hansen and Mackin (1949) found abundant lodgepole pine (*Pinus contorta*) pollen, which they thought indicated an early interglacial forest. Higher in the Whidbey sediments, they found less lodgepole pine pollen and more western hemlock (*Tsuga heterophylla*) pollen. These

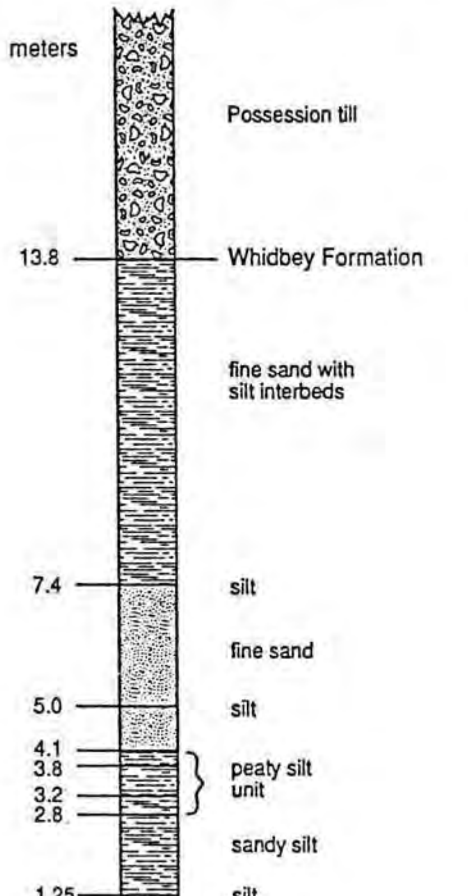

Figure 17. Stratigraphic section at Lagoon Point. The sample dated at 151 ± 38 ka came from the peaty silt unit near the base of the Whidbey Formation sediments.

Table 4. Chemical composition of Camano pumice and Lake Tapps tephra at its type locality. * microprobe analysis by J. A. Westgate, University of Toronto; ** data from Westgate and others, 1987

	Camano pumice*	Lake Tapps tephra at Salmon Springs (UT88)**
SiO_2	73.16	78.6
TiO_2	0.31	0.20
Al_2O_3	13.61	12.3
FeO	2.59	0.69
CaO	1.67	0.81
Na_2O	3.21	3.6
K_2O	4.56	3.7

Figure 18. Whidbey Formation silt, clay, and peat overlain by pumiceous sand, Blowers Bluff, Whidbey Island. A TL age of 107 ± 17 ka was measured from clay near beach level.

changes in the pollen spectra suggest that warming of the climate followed retreat of the Double Bluff glacier. In the lower part of the section at Everett, they found that western hemlock and Douglas-fir (*Pseudotsuga menziesii*) pollen are most abundant, indicative of further climatic amelioration. Progressive increases in the amount of lodgepole pine pollen in sediments higher in the Everett section are accompanied by disappearance of Douglas-fir pollen and decrease in western hemlock pollen, implying climatic instability, possibly contemporaneous with the advance of glacial ice into the Puget Lowland.

Pollen analyses of peat beds in the Whidbey Formation at the type locality (Easterbrook and others, 1967) show that the lowest peat beds contain mostly spores, accompanied by small amounts of tree pollen. Pollen in overlying peat consists of about equal amounts of Douglas-fir and hemlock, in addition to lesser amounts of lodgepole pine, Engelmann spruce, fir, and juniper pollen. Alder pollen predominates in the peat near the top of the formation, and spores and pine pollen are abundant.

The climatic implications of these pollen assemblages can be inferred from the modern altitudinal distribution of these trees in the Cascade Range of western Washington. This distribution is based on surface samples from the uppermost layer of several peat bogs at various altitudes on the west slope of the mountains; the samples show that modern pollen rain in the Canadian Zone (elevations of 1,060–1,970 m) is dominated by pine pollen and contains less hemlock pollen. Pollen rain in the upper Humid Transition Zone (elevations of 600–1,060 m) is also dominated by pine but has more abundant hemlock and Douglas-fir pollen than the Canadian Zone. However, modern pollen rain in the lower Humid Transition Zone (sea level to 600 m) shows a combined predominance of hemlock and Douglas-fir, the latter pollen being more prevalent at drier localities. At all lowland sites, pine is a minor part of the pollen rain, comprising less than 20 percent of the total tree pollen; spruce (*Picea*) pollen is present in small amounts.

Comparison of modern pollen data with the fossil pollen assemblages in the Whidbey Formation suggests that the predominance of nearly equal percentages of Douglas-fir and hemlock pollen in peats probably represents a climate similar to that of the present in the lowland. Pollen spectra dominated by alder and containing pine and lesser amounts of other conifer pollen resemble the modern pollen rain at some places in the Puget Lowland, as shown by samples from a peat bog near the northern end of Whidbey

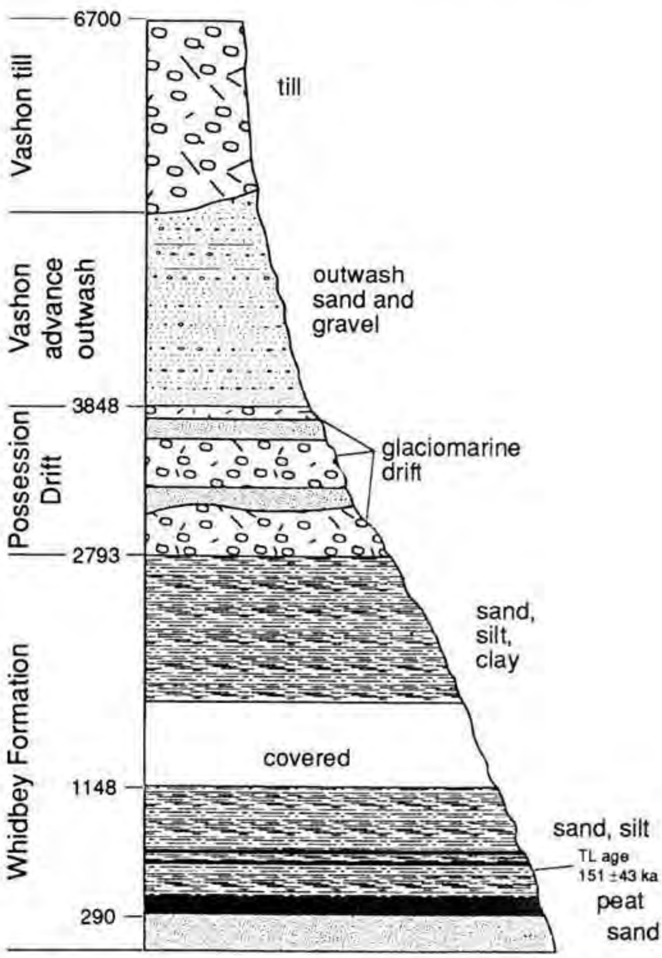

Figure 19. Stratigraphic section at Point Wilson. A TL age of 151 ± 43 ka was measured from clay in Whidbey Formation interbedded sand, silt, clay, and peat near beach level.

Figure 20. Whidbey Formation silt, clay, and peat TL dated at 151 ± 43 ka at Point Wilson. The sample site is at the base of the section in the photograph.

Island (Heusser and Heusser, 1981). The highest Whidbey peat at the type section contains a small amount of Engelmann spruce, not found in modern pollen on Whidbey Island.

Amino-acid ages

An amino-acid age of 97 ± 35 ka (Blunt and others, 1987) has been calculated from shells in laminated marine clay mapped as Whidbey Formation (Easterbrook, 1968) at South Whidbey State Park on Whidbey Island. Amino-acid ages of 96 ± 35 ka from leucine and 107 ± 9 ka from allo/iso ratios were calculated from marine shells from sediments correlated with the Whidbey Formation north of Lagoon Point (Blunt and others, 1987).

Thermoluminescence ages

Four TL ages have been measured on clay in the Whidbey Formation (Easterbrook and others, 1992):

- At Lagoon Point, interbedded sand, wood-bearing silt, and peat of the Whidbey Formation overlie Double Bluff Drift. TL measurements of a silty-clay bed in Whidbey fluvial sediments (Figs. 16 and 17) yielded an age of 102 ± 38 ka.
- A TL age of 106 ± 17 ka was measured on clay in the lower part of the Whidbey Formation near sea level at Blowers Bluff (Fig. 18). The clay bed is overlain by several meters of pumice-rich sand and interbedded sand and silt, which is unconformably overlain by Possession till.
- A clay lens in interbedded sand, silt, clay, and peat at sea level about 1 km north of West Beach on Whidbey Island was TL dated at 142 ± 10 ka. The formation is here unconformably overlain by more than 30 m of Esperance sand and Vashon till.
- A TL age of 151 ± 43 ka was measured on clay in Whidbey Formation interbedded sand, silt, clay, and peat at Point Wilson (Figs. 19 and 20). The Whidbey Formation there is overlain by Possession glaciomarine drift.

Calc-alkaline pumice is concentrated in lenses in Whidbey sand at several localities on Whidbey Island. Geochemical analyses (Table 5) suggest that they are comagmatic and useful as stratigraphic marker beds. Their age has not yet been measured.

Table 5. Electron microprobe analyses of major elements in pumice from the Whidbey Formation. (Analyses by J. A. Westgate)

	Upper Blowers Bluff (bulk)	Upper Blowers Bluff (glass)	Lower Blowers Bluff (bulk)	Lower Blowers Bluff (glass)	Freeland (bulk)	Freeland (glass)
SiO_2	65.76	68.33	67.89	68.65	73.44	72.35
TiO_2	0.54	0.19	0.50	0.18	0.29	0.17
Al_2O_3	16.12	17.93	15.30	18.03	13.92	15.33
Fe_2O_3	4.05	1.28	3.40	1.17	2.12	1.17
MnO	0.11	0.03	0.06	0.02	0.06	0.02
MgO	1.92	0.18	1.60	0.08	0.85	0.35
CaO	4.39	4.18	3.80	4.23	1.56	2.49
NaO	4.34	5.11	4.47	4.93	3.11	4.22
K_2O	2.58	2.66	2.78	2.60	4.52	3.75
P_2O_5	0.20	0.11	0.19	0.09	0.13	0.14

Possession Drift

Possession Drift consists of till, sand and gravel, and glaciomarine drift. The type locality of the Possession Drift is sea cliffs at Possession Point, where about 35 m of compact Possession till is underlain by peat-bearing sand and silt of the Whidbey Formation and overlain by Vashon Drift (Easterbrook and others, 1967). In sea-cliff sections at Blowers Bluff, along the east side of Useless Bay on Whidbey Island, at Port Williams on the Olympic Peninsula, and on the mainland east of Camano Island, deposits correlated with Possession Drift consist of glaciomarine diamictons containing marine shells and shell fragments (Blunt and others, 1987).

Possession Drift is discontinuous laterally and pinches out within a few hundred meters in many sea-cliff exposures. At its type locality, it pinches out laterally to the west, and its stratigraphic position becomes an unconformity between the Whidbey Formation and younger sediments. At many other places, its stratigraphic position is represented only by an unconformity.

Calculations based on amino-acid analyses of marine shells in Possession glaciomarine drift at Port Williams, Stillaguamish, and Blowers Bluff suggest a mean age of 80 ± 22 ka (Blunt and others, 1987; Easterbrook and Rutter, 1981, 1982). No laser-argon or TL dates have yet been obtained from Possession Drift.

CONCLUSIONS

Laser-argon, fission-track, TL, amino-acid, and paleomagnetic dating techniques now provide the basis for establishing a firm chronology for the Orting, Stuck, Salmon Springs, Double Bluff, and Possession Glaciations and the interglacial Alderton, Puyallup, and Whidbey Formations. Because the Orting Drift is reversely magnetized and lies beneath sediments laser-argon dated at 1.6 Ma, its age is considered to be between 1.6 and 2.4 Ma. The Alderton Formation is reversely magnetized, and laser-argon dating of volcanic ash yielded an age of 1.6 Ma. Stuck Drift has not yet been directly dated, but it lies between nonglacial sediments laser-argon dated at 1.6 Ma (see Fig. 2). The Puyallup Formation is reversely magnetized, and laser-argon dates of 1.69 ± 0.11 Ma and 1.64 ± 0.13 Ma have been obtained from pumice in fluvial sediments. The Salmon Springs Drift is reversely magnetized and contains the Lake Tapps tephra, fission-track dated at 1.06 ± 0.11 Ma.

Clay beneath Double Bluff Drift has been TL-dated at 289 ± 74 ka and 291 ± 86 ka. Double Bluff glaciomarine drift has been TL-dated at 177 ± 38 ka, and clay overlying Double Bluff till has been TL-dated at 320 ± 100 ka. These ages correspond quite well with amino-acid ages from mollusk shells in Double Bluff glaciomarine drift that range from 111 to 178 ka at the type locality and from 150 to 250 ka elsewhere in the central Puget Lowland. Double Bluff sediments are normally magnetized.

TL dates of 102 ± 38 ka, 106 ± 17 ka, 142 ± 10 ka, and 151 ± 43 ka have been obtained from clay in Whidbey Formation interglacial fluvial sediments. These ages compare well with amino-acid ages of 97 ± 35 ka, 96 ± 35 ka, and 107 ± 9 ka. The Possession glaciomarine drift at three localities has been dated by amino-acid analyses of marine shells that give a mean age of 80 ± 22 ka.

ACKNOWLEDGMENTS

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Patterns and Processes of Landscape Development by the Puget Lobe Ice Sheet

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ABSTRACT

The Puget lobe, the southwesternmost extension of the Cordilleran ice sheet, last advanced into the Puget Lowland of western Washington about 15,000 years ago. It left behind a varied record of both depositional and erosional landforms, which together still dominate the landscape of the region. This ice advance provides an excellent opportunity to examine the connection between glacier processes and the resulting products of glaciation. The role of water, in both transporting sediment and determining material properties of the glacier bed, is particularly significant for the geologic record of this ice sheet.

Most prominently, the advancing ice sheet deposited voluminous sediment on a proglacial outwash plain that extended from the Olympic Mountains to the Cascade Range, herein recognized as the "great lowland fill". Subsequent overrunning by the ice sheet modified this surface in several ways: most pervasively, by the deposition of basal till, which was commonly shaped into drumlins, and most notably, by the excavation of deep linear troughs now occupied by large lakes and the marine waters of Puget Sound. Excavation of the troughs and valleys of the Puget Lowland required the transport of about 1,000 cubic kilometers of sediment, almost entirely during ice occupation and primarily by subglacial water.

Trough formation required vast subglacial water discharges and probably persisted throughout the several thousand years of ice occupation. Drumlins, however, show very little evidence of subglacial water activity and are inferred to have formed primarily during ice recession. This suggests time-transgressive formation of drumlins as the ice front retreated from south to north, a characteristic narrow drumlin-forming zone (probably less than 20 kilometers wide) persisting at any one location for only a few decades.

INTRODUCTION

More than a century of study of the Cordilleran ice sheet has led to a well-developed picture of ice-sheet chronology and glacial-age deposits. Yet many of the basic questions of ice-sheet behavior and effects, particularly how that ice has shaped the landscape of the Puget Sound region, are addressed only imperfectly or not at all by such studies.

The area affected by the Puget lobe, the southwesternmost extension of the Cordilleran ice sheet (Fig. 1), provides an exceptional opportunity to examine the connection between glacier processes and the geomorphic products of the glacier system. Most importantly, the Puget lobe system is well constrained, with good age control and clearly recognized boundaries. The geomorphic products of glaciation are well displayed across the Puget Lowland, despite the region's luxuriant vegetation cover. Finally, a number of ancillary studies of late Pleistocene environments in the region, particularly those reflecting climatic variation from local glaciers (Porter, 1977), vegetation (Barnosky and others, 1987), and global climate (Kutsbach, 1987), establish an independent environmental context and provide invaluable checks on the results of ice-sheet reconstruction and modeling.

REGIONAL SETTING

Geography

The southwest part of the Cordilleran ice sheet occupied a distinctive geographic region (Fig. 1). A broad topographic basin in British Columbia, the Georgia Depression, extends southward into Washington state. The depression splits at the northeast corner of the Olympic Peninsula. One branch, the Strait of Juan de Fuca, trends west between the Olympic Mountains and Vancouver Island. The other branch continues south, forming the Puget Lowland between the Olympic Mountains and Cascade Range. Although low hills at about latitude 46°45′ define the southern limit of ice advance in the Puget Lowland, the lowland province itself continues south for an additional several hundred kilometers.

The southern Cordilleran ice sheet expanded into the Puget Lowland during several episodes of Pleistocene glaciation (Crandell and others, 1958; Easterbrook and others, 1967; Easterbrook, 1986). Products of the most recent advance, the Vashon Stade of the Fraser glaciation of Armstrong and others (1965), provide the best picture of ice-sheet growth and decay. Ice caps on the mountains of Vancouver Island and the British Columbia mainland expanded

Figure 1. Location of the Puget Lowland in northwestern Washington state, showing the southwestern limit of the Cordilleran ice sheet (hachured line) during the Vashon Stade of the Fraser glaciation, about 15,000 years ago. The continental ice sheet divided into two lobes, the Juan de Fuca lobe southwest of Vancouver Island and Puget lobe just west of the Cascade Range. Once the Strait of Juan de Fuca was blocked by ice, meltwater from the Puget lobe escaped southwest over a topographic divide through a spillway system (arrows) that ultimately drained to the Pacific Ocean (after Booth 1986a, fig. 1).

and coalesced, gradually extending into the lowland valleys and the Georgia Depression (Clague, 1981). However, as the ice tongue moved southward into the Puget Lowland, it probably received no additional lateral input because glaciers in the Olympic Mountains and Cascade Range had already retreated from their late Pleistocene maximum limits (Porter, 1976; Booth, 1987).

Chronology

During the Fraser glaciation, the advancing and retreating Puget lobe was among the more rapidly moving North American ice sheets. Only a few thousand years span the advance of the lobe across the Canadian border, attainment of maximum southern position, and retreat back to the foothills of the British Columbia mountains. A movement of the terminus of more than 200 km in each direction was therefore accomplished in this period, resulting in an average velocity of the terminus of at least 100 m per year. Yet the geomorphic record in the Puget Lowland is primarily that of an ice sheet at maximum stage (see also Sugden, 1979), with well-defined ice limits and many indicators of ice-flow direction that are consistent with a sustained ice-maximum position (Thorson, 1980; Booth, 1990). Thus the lobe undoubtedly maintained its maximum position for at least some fraction of its total history, with limiting radiocarbon dates (see below) permitting 500–1,000 years of maximum or near-maximum conditions. Advance and retreat rates were therefore even higher than their minimum, averaged value of 100 m per year.

Abundant radiocarbon dates constrain both the advance and the retreat chronology of the Puget lobe and the neighboring Juan de Fuca lobe (Fig. 2). The ice sheet entered the Georgia Depression some time after 18.3 ± 0.17 ka (GSC-2322; Armstrong and Clague, 1977; Fig. 2). The advance of the Puget lobe is further constrained in the Seattle area, 160 km south of the international boundary, by a pre-glacial age of 15.0 ± 0.4 ka (W-1227; Mullineaux and others, 1965).

Final advance to the ice-maximum position and subsequent retreat are also detailed for both lobes. The Juan de Fuca lobe has one dated locality near its terminus, a post-retreat date of 14.46 ± 0.20 ka (Y-2452; Heusser, 1973) 50 km upglacier of its ice-maximum limit. The Puget lobe advanced an additional 100 km south of Seattle and then retreated a like amount between 15.0 ka (see above) and 13.65 ± 0.55 ka (L-346a; Rigg and Gould, 1957). Yet nearly equivalent dates of 13.60 ± 0.15 ka and 13.65 ± 0.35 ka are also reported from shells on Whidbey Island, an additional 30 to 50 km north of Seattle (BETA-1716 and BETA-1319; D. P. Dethier and others, U.S. Geological Survey, written commun., 1986). Grounded ice of both the Puget and Juan de Fuca lobes therefore must have retreated at least this far upglacier by that time. Required minimum rates of terminal advance and retreat are about 200 m per year during this period. Allowing for slower advance rates (Weertman, 1964) and sufficient time to achieve equilibrium at ice-maximum position, actual rates of terminus retreat were probably closer to 500 m per year.

RECONSTRUCTION OF THE PUGET LOBE

Introduction

Determining the physical behavior of an ice sheet first requires some knowledge of its physical dimensions. The flux of both ice and water must be estimated as well, because the combined effects of these agents on the glacier bed causes geomorphic changes. Unfortunately, most glacier systems are ill suited to such a reconstruction. If applied to an existing glacier, the necessary parameters are easily measured, but their effects on the glacier bed are obscured. Pleistocene ice sheets suffer from the inverse problem—typically, the record of their passage is grossly incomplete or their mass-balance regime is completely unknown. The Puget lobe does not completely avoid these shortcomings of other vanished ice sheets, but sufficient data are available and constrained by independent checks such that reconstruction is both feasible and instructive. The discussion of method and results that follows is summarized from Booth (1986a), which includes a more extensive review of techniques, data sources, errors, and independent verifications.

Method

This reconstruction is based first on the compilation of available geologic data on the physical extent of the ice sheet (Fig. 3) and second on the calculation of an equilibrium mass balance for the Puget lobe itself. The compilation of ice limits relies heavily on Thorson (1980) in the south and Wilson and others (1958) in the north. In addition, we follow several basic assumptions: ice-surface contours lie perpendicular to flow indicators, flow lines do not converge or diverge without commensurate changes in ice thickness or net mass balance, and longitudinal stress gradients (that is, the downglacier change in the depth-slope product) are low at the scale of the reconstruction. Some of these assumptions are contradicted in certain localities; we discuss the resulting implications in a later section.

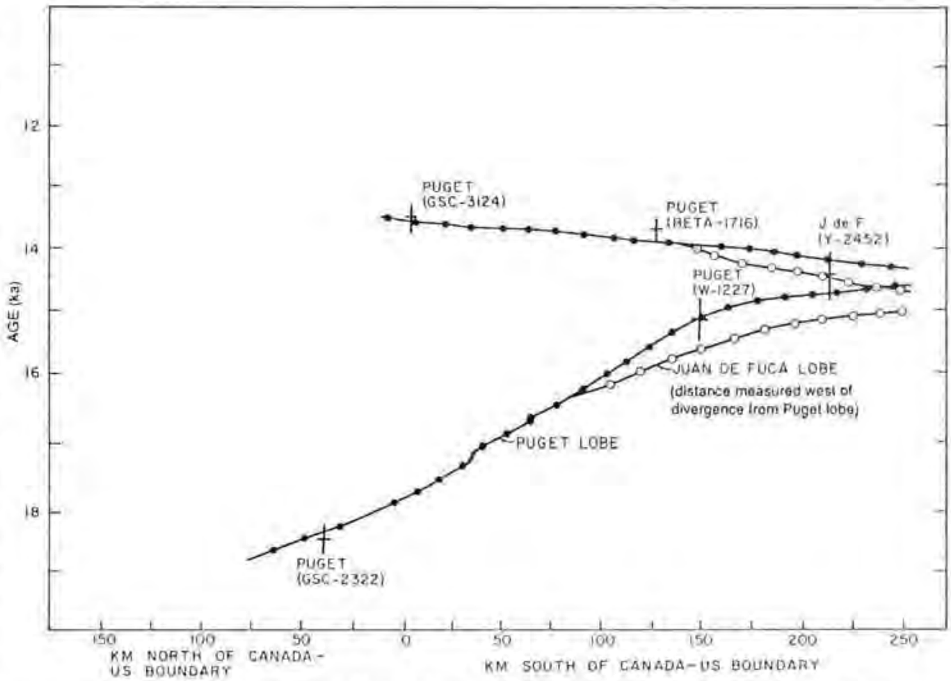

Figure 2. Advance and retreat of the Puget and Juan de Fuca lobes, with key limiting radiocarbon dates and standard errors (adapted from Booth, 1987).

The mass balance of an ice sheet during the Pleistocene cannot be known with certainty, but analogy with existing glaciers provides a reasonable working estimate. In the Pacific Northwest, long-term studies of several modern maritime glaciers yields a local mass-balance relation (Meier and others, 1971; Porter and others, 1983) that expresses net accumulation or ablation as a function of elevation above or below the equilibrium line altitude (ELA). An analogous expression for total ablation provides a measure of the water produced that discharges through the glacier, irrespective of whether that melting is compensated for by snow accumulation. We have used such a relation compiled from high-quality data from two Norwegian maritime glaciers (Mueller, 1977).

Results

With a reconstructed ice sheet and a relation between relative altitude and mass balance, an ELA that brings the lobe into balance can be determined by trial and error. Using these procedures, accumulation and ablation of the Puget lobe is balanced with an ELA between 1,200 and 1,250 m. The contribution to ice velocity from internal ice deformation can be calculated (Paterson, 1981) and accounts for less than 2 percent of the total flux. Thus basal sliding must account for nearly all the predicted motion, which is several hundred meters per year across nearly the entire source and ablation areas of the lobe. The ice flux peaks at the ELA, in contrast to the monotonic increase in water flow downglacier.

LANDSCAPE-FORMING PROCESSES

Glacial Sedimentation

The great lowland fill

As the ice sheet advanced, proglacial rivers carried fluvial sediment away from the front of the ice. Named the Esperance Sand Member of the Vashon Drift by Mullineaux and others (1965), this sediment is recognized in virtually every part of the Puget Lowland and is now mapped as Vashon advance outwash. Crandell and others (1965) first suggested that this deposit may have been continuous across the modern-day arms of Puget Sound. In the Georgia Depression, Clague (1976) inferred that the correlative deposit (the Quadra Sand) probably filled that trough as well. In the Puget Lowland, this infilling was particularly likely because the basin became a closed depression once the ice advanced south past the entrance of the Strait of Juan de Fuca, raising the base level of the Puget Lowland by blocking the only sea-level drainage route. Lacustrine sediment, the Lawton Clay Member of the Vashon Drift (Mullineaux and others, 1965), records initial lowland ponding during this time, but coarser outwash subsequently accumulated as the ice advanced farther south. Lacustrine deposits lying stratigraphically above both the advance outwash and basal till record a later, analogous environment during ice recession as well.

This model of proglacial sedimentation, and its logical prediction of a basin-wide outwash plain, can be tested. We seek to reconstruct this relict, lowland-wide outwash sur-

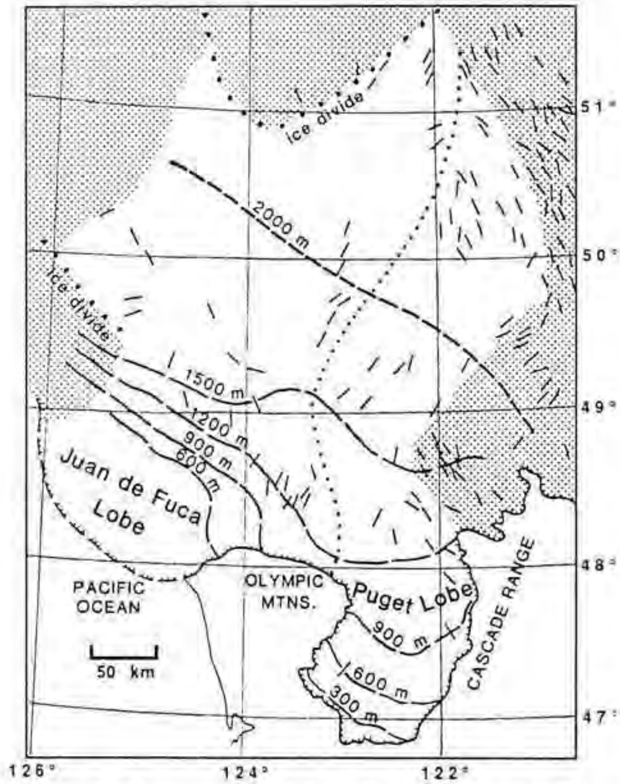

Figure 3. Reconstruction of the Puget and Juan de Fuca lobes at maximum stage. Short lines show orientations of representative striations and other glacial lineations; hachured line shows maximum extent of ice. Heavy lines are contours of ice-surface altitudes above modern sea level, uncorrected for glacial-age sea-level lowering or isostatic depression. Unshaded ice-covered area included all ice of the Puget and Juan de Fuca lobes, with an inferred separation between the lobes indicated by the dotted line (adapted from Booth, 1986a).

face by recognizing those landforms that should best reflect the surface's original elevation, should that surface in fact still exist (Fig. 4). Geologic mapping throughout the region (see index map in Booth, 1991a) is sufficient to demonstrate that the elevation of most topographic prominences across the Puget Lowland is defined by the thickness of Vashon advance outwash. Existing mapping is not, however, adequate to accurately locate the elevation of the till-outwash contact. As a result, the thickness of till overlying the outwash, typically on the order of a few meters (and so only a few percent of the outwash thickness), must be ignored for purposes of this reconstruction, but the introduced error is minor.

Following these conventions, a crude topographic map of the proglacial outwash surface can be generated (Fig. 5). The generally southward slope of the reconstructed surface reflects drainage out the south end of the Puget Lowland, over the Black Hills and into the Chehalis River. That slope is not corrected for isostatic rebound (Thorson, 1989) because original deposition would have occurred prior to depression of the crust by the weight of the ice sheet. The "dishing" at eastern and western margins probably reflects inwash of fluvial sediment from mountain drainages during deposition, spatially smoothed by the topographic reconstruction. This outwash surface is not simply an artifact of the mapping conventions. Once recognized, its presence is readily apparent throughout the Puget Lowland (Fig. 6). Despite subsequent, localized excavation and incision (see below), this surface persists today as the most prominent single landform of the entire Puget Lowland.

The timing of outwash-plain deposition is well constrained. Outwash did not inundate the Seattle area until shortly before 15,000 yr B.P. (W-1227; Mullineaux and others, 1965). Initial deposition may have begun as long as a few thousand years earlier, but until drainage out the Strait of Juan de Fuca was blocked (about 16,000 yr B.P.), aggradation farther south would have been limited in thickness. This deposition, however, must have been complete across the entire lowland prior to ice maximum at about 14,500 yr B.P. (Booth, 1987) because basal till, reflecting the overriding of the ice sheet itself, is ubiquitous as the capping deposit.

In a tectonically active environment (Gower and others, 1985), the formation of a near-planar surface at about 15,000 yr B.P. invites assessment of subsequent crustal deformation. Unfortunately, the imprecision with which this outwash plain can be reconstructed would make warping of even 10 or 20 m difficult to demonstrate. However, uplift of about 7 m has been suggested for an earthquake about 1,100 years ago in the central Puget Sound (Bucknam and others, 1992); the likelihood of multiple events since 14,000 yr B.P. is also high (Atwater, 1987). A plausible maximum rate of 7 m per 1,100 yr, however, could not be sustained throughout the Holocene, based on the reconstruction of Figure 5. Indeed, only one-half, and probably less than one-third, of that rate is consistent with these data, a constraint on the recurrence history of large, shallow earthquakes in the Puget Lowland.

Till deposition and drumlin formation

Lying directly above the outwash is a gray to blue-gray till layer. The till is compact, hard, and clay rich; it commonly displays strongly developed pebble fabrics that parallel the regional ice-flow indicators (Goldstein, unpub. data; Booth, 1990). On the basis of these characteristics, it has long been interpreted as a basal till, deposited beneath actively flowing ice (Garling, Molenaar, and others, 1965). Particularly in the southern Puget Lowland, the till has been extensively molded into north-trending drumlins that typically extend a few tens or hundreds of meters in plan view and have as much as 10 m of vertical relief.

Across the Puget Lowland, the basal till displays other common characteristics. It ranges between 1 and 10 m thick, but it generally averages less than 5 m. This till blankets the ground surface and commonly truncates bedding in underlying units at all spatial scales—not only along the sidewalls of many large river valleys, but also along the lateral margins of individual drumlins. The lower 10–50 percent of the till is enriched in material from the underlying sediment, be it pebble- and cobble-size clasts derived

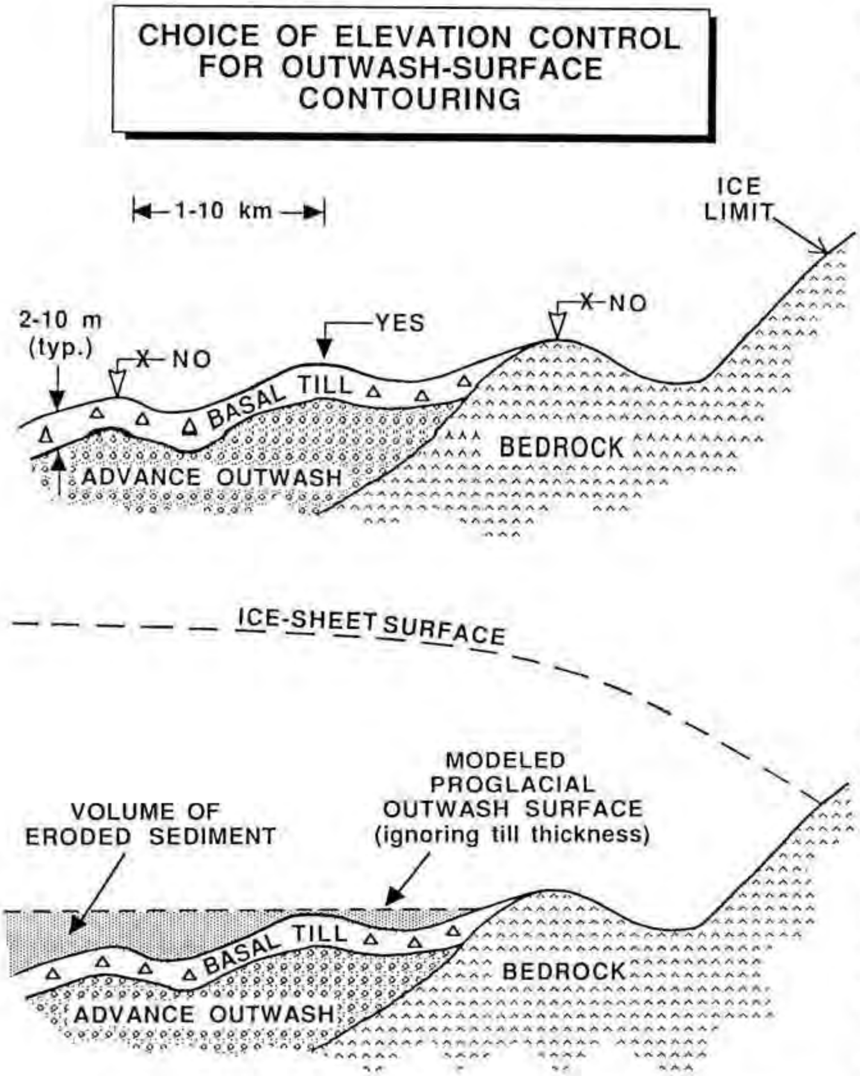

Figure 4. Mapping conventions for reconstruction of the outwash surface. Elevations selected for plotting (upper part of this figure) are local topographic maxima on glacially deposited surfaces (labeled "YES"), not bedrock or lower nearby depositional surfaces (labeled "NO"). The thickness of till is ignored. Once the reconstructed surface is defined (lower part of this figure), the volume of sediment subsequently eroded can be readily calculated.

from outwash or silt and clay from the local lacustrine sediments. These local components, however, become progressively surpassed by farther traveled components higher in the till. This vertical dilution is primarily manifested by an upward decrease in clast and matrix grain size, perhaps representing progressive diminution during long-distance transport. In the southern lowland, there is also an upward increase in the proportion of subangular clasts derived from bedrock farther north in the lowland, which markedly contrast to the well-rounded fragments that originated from the underlying outwash.

Various subglacially or englacially water-sorted deposits are scattered throughout the basal till, reflecting a wide range of flow conditions. (See also Brown and others, 1987.) Where present, these lenses of bedded gravel, sand, or silt generally have sharp, erosional lower contacts, but their upper contacts are gradational with the overlying till.

At any single locality, the character of these lenses probably reflects changing subglacial water conditions over time. The geometry of lens contacts and the nature of the interbedding suggest that these sediments were deposited in cavities that contained diminishing water discharge that eventually was insufficient to support the weight of the ice above. Ultimately, each cavity was closed off, permitting till deposition to be re-established. In the southern lowland, these sorted lenses constitute only a minor component of the material above the advance outwash (generally much less than 10 percent at any one section). This low proportion suggests that sites of till deposition were not sites of significant simultaneous meltwater deposition as well.

The sorted lenses may also be used as a crude measure of the total strain experienced by the till. They consistently appear to be in their original form and geometry, and where plastic or brittle deformation structures do exist, the total strain does not appear to be more than 1 or 2. (See also Brown and others, 1987.) These characteristics hold true for lenses located at all vertical positions in the till and regardless of grain size.

As has been shown for parts of the midcontinent region of the United States formerly affected by the Laurentide ice sheet (Clayton and others, 1989), deformation either by pervasive shear or repeated folding appears to have been only a minor process in the genesis of some tills. If either of these two deforming mechanisms had indeed occurred to a significant extent, relict subglacial or englacial structures such as the sorted lenses and vertical compositional variations, as well as intact weathering rinds on clasts, would not have survived, but rather would have become "homogenized" in the manner envisioned by Boulton (1987). The observed characteristics of the till suggest that it was formed predominantly by accretion, as demonstrated for both sheets and drumlins of till studied elsewhere (Goldstein, 1989).

Timing of drumlin formation

The pattern of drumlin orientation allows us to infer the relative timing of drumlin formation. Although drumlin orientations fan over an arc of about 120 degrees in a pattern that is controlled primarily by the ice flow into southward-widening outline of the Puget Lowland, the trends of some individual drumlins and local groups of drumlins are discordant to this regional pattern (Fig. 7). This disparity is

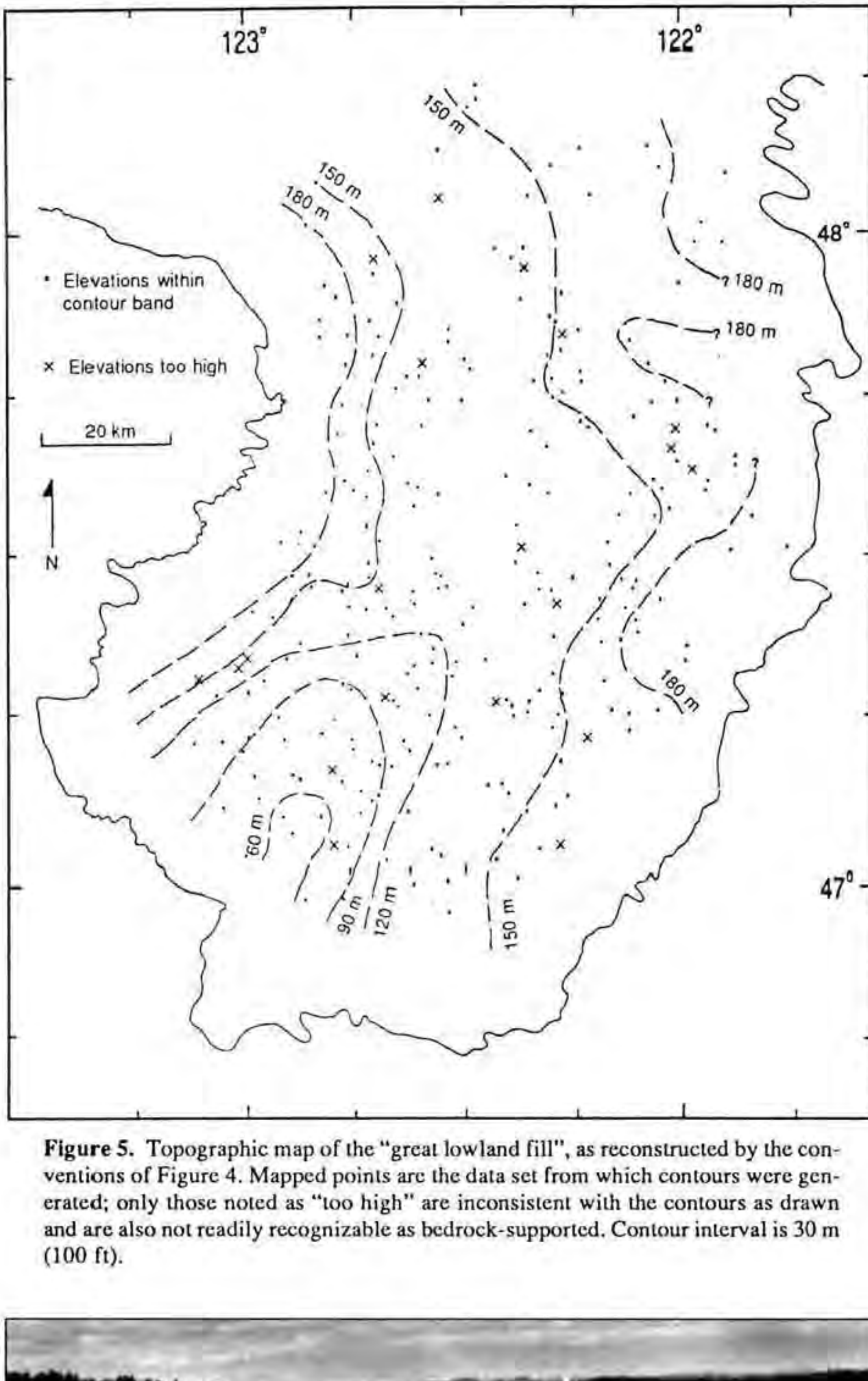

Figure 5. Topographic map of the "great lowland fill", as reconstructed by the conventions of Figure 4. Mapped points are the data set from which contours were generated; only those noted as "too high" are inconsistent with the contours as drawn and are also not readily recognizable as bedrock-supported. Contour interval is 30 m (100 ft).

Figure 6. View of the "great lowland fill" to the southwest toward the south end of one of the Puget Lowland troughs (Lake Washington). Note that although individual hills may be of lower altitude, their maximum altitudes form a well-defined concordant surface, here about 140 m (450 ft) above sea level.

particularly noticeable in the west-central and ice-marginal areas of the southern lowland, where drumlins are best developed.

Demonstrating this discordance is hampered, but not precluded, by the method of ice-sheet reconstruction. For the ice-sheet interior, ice-flow directions were commonly estimated from drumlin orientations themselves, which obviously discounts any discordance between these two directions that might actually have existed. Near the ice margins, however, primary data such as the elevation of glacial erratics and the trend of glacially scoured bedrock is used in the ice-sheet reconstruction. In these areas, any discordance between drumlin orientation and ice-flow direction (or perpendicularity to ice-surface contours) is demonstrable.

Where documented, drumlin orientations are commonly more nearly perpendicular to the reconstructed ice-marginal channels of Thorson (1980) that mark recessional positions (Fig. 8) than to ice-maximum ice-surface contours or ice limits. This implies that the drumlins may have been formed within a narrow zone behind the margin (see also Mooers, 1989), perhaps no wider than a few kilometers, in a time-transgressive manner as the ice sheet decayed and retreated northward. This retreat was apparently that of an active, rather than a stagnant, ice margin, as deduced from the limited distribution of upland dead-ice topography in these areas (Thorson 1980; Booth, 1990).

Successive formation of drumlins during retreat, rather than simultaneous formation during ice maximum, implies that individual drumlins probably formed during a few decades at most. If the drumlin till formed predominantly by accretion, rather than by deformation, then the till must have accumulated vertically as rapidly as 0.5–1.0 m per year, using an average till thickness of 10 m (Garling, Molenaar and others, 1965; Molenaar and Noble, 1970).

Glaciofluvial Erosion of the Lowland Trough

Prominent in any map of the Puget Lowland is the system of sub-parallel troughs (Fig. 9) that today extend to

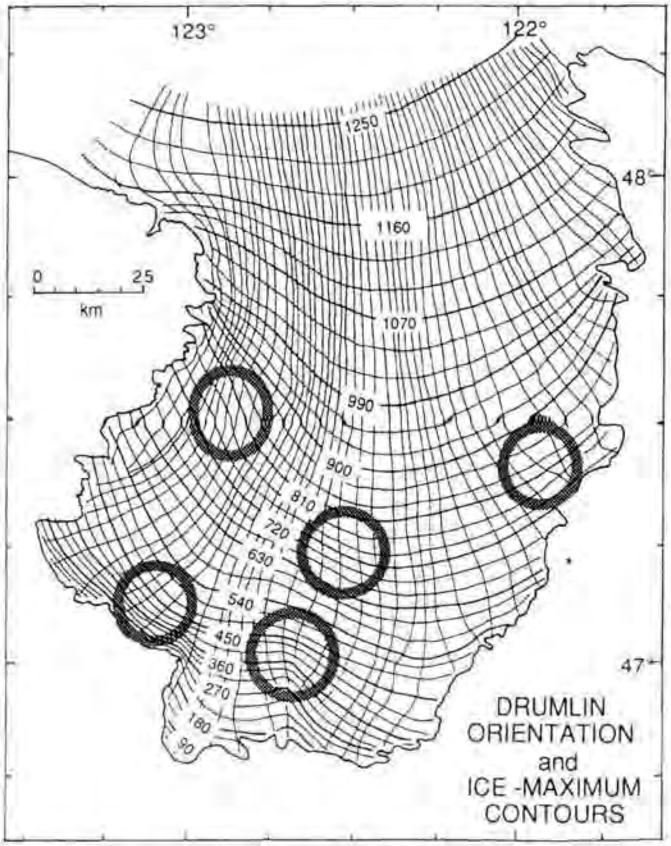

Figure 7. Basal flow lines of the Puget lobe, derived mostly from drumlin trends, superimposed over a detailed reconstruction of the surface of the ice sheet (after Thorson, 1980, figs. 2 and 4). In many areas (circled), drumlin trends are not perpendicular to reconstructed contours, contrary to the pattern that would be expected if the drumlins had formed during maximum stage.

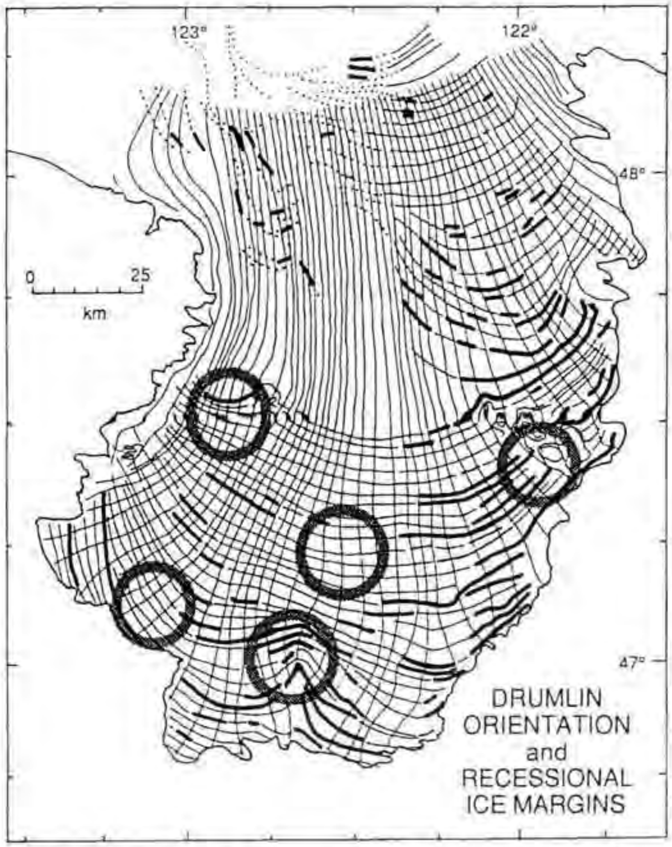

Figure 8. Basal flow lines of the Puget lobe derived mainly from drumlin trends (as in Fig. 7) superimposed on a reconstruction of the location of the ice margin during recession. Heavy lines represent demonstrable ice-marginal positions as mapped in the field, whereas thinner lines were assumed from theoretical considerations (after Thorson, 1980, fig. 8). Where differences are substantial, basal flow lines are more nearly perpendicular to the successive recessional positions shown here than to the ice-maximum surface contours (Fig. 7), suggesting that drumlin formation was near-marginal and time-transgressive.

depths as great as 400 m below the surface of the drumlinized uplands. Whereas these troughs appear to be the result of glacial occupation of a pre-glacial drainage system (for example, Willis, 1898), they actually post-date deposition of the "great lowland fill" of Vashon advance outwash (Crandell and others, 1965). These features thus represent the primary deviation in the modern landscape from the idealized smooth surface formed by the advance outwash. The troughs are much deeper than can be explained by the primary (that is, depositional) relief of an outwash plain; hence, their formation required post-depositional erosion.

If the outwash surface was originally smooth, then the volume of subsequently eroded material can be calculated directly from the differences in altitude between the reconstructed surface and the modern topography (Fig. 10). This calculation shows that nearly 1,000 km^3 of sediment has been eroded, equivalent to a layer about 100 m thick over the ice-occupied area of the Puget Lowland as a whole. An even greater volume of eroded and subsequently redeposited sediment is suggested by recent seismic profiles across several of the channels of Puget Sound (M. Holmes, U.S. Geological Survey, written commun., 1991).

The timing of trough erosion into the great lowland fill is constrained to the limited period between deposition of the advance outwash and subaerial exposure of the glacier bed during the ice recession. Basal till drapes down the flanks of many of the eroded troughs throughout the lowland, to as low as sea level in places (for example, Booth, 1991a). Furthermore, an extensive series of lowland lake basins (Thorson, 1989), excavated into the advance-outwash deposits to elevations well below the reconstructed upper advance-outwash surface, already existed by the time retreat of the Puget lobe was under way. Thus the lake troughs were formed primarily, or exclusively, by subglacial processes.

In the context of the regional glacial history, these constraints mean that, in places, nearly half a kilometer of subglacial erosion was accomplished during the brief period that the Puget lobe occupied the southern lowland. Postglacial excavation is not likely to have contributed much to

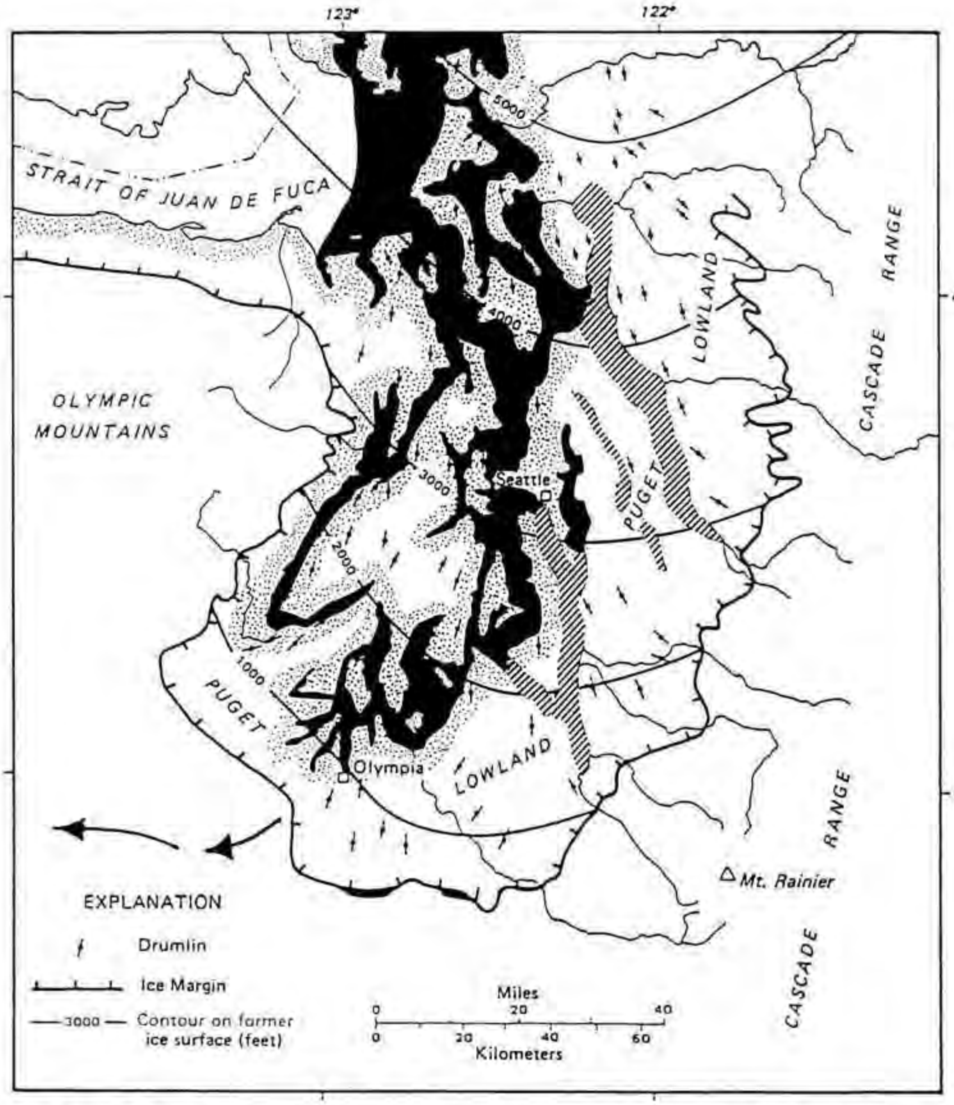

Figure 9. The Puget lobe at maximum stage, showing the orientation of selected drumlins and other streamlined hills, as well as the modern topographic expression of the deep lowland troughs that are either water-filled (solid black) or infilled by Holocene sediment but inferred to have likely existed in late-glacial time (diagonal lines; see Hall and Othberg, 1974). Arrows indicate the location of the lobe's southern drainage system along the modern Chehalis River valley (see also Fig. 1; after Easterbrook, 1979).

the excavation of the troughs because during deglaciation the southern lowland was once again subjected to proglacial lacustrine deposition within the newly created troughs in front (south) of the retreating Puget lobe (Thorson, 1980). Massive subglacial excavation of the troughs was therefore simultaneous with subglacial formation of drumlins on the intervening uplands.

These troughs were probably carved primarily by subglacial meltwater. (See Booth and Hallet, 1993.) Their (rather sinuous) morphology is well expressed not only in the ice-sheet interior, where both ice and subglacial water fluxes are high, but also in areas out to the ice terminus, where ice flux declines rapidly but subglacial water discharge continues to increase. This conclusion is in full accord with the inferences drawn from other recent investigations of Pleistocene glacier-occupied troughs in southern British Columbia (Eyles and others, 1990) and New York (Mullins and Hinchey, 1989) and of tunnel valleys with remarkably similar dimensions and relief in Nova Scotia (Boyd and others, 1988) and Germany (Ehlers, 1981). Across the Puget Lowland, these troughs number about ten (Fig. 9). To accommodate the total water discharge of the Puget lobe (Booth, 1986a), each of them probably carried an average discharge of some several hundred cubic meters per second.

Although this predicted magnitude of subglacial fluvial erosion is very large, its plausibility can be tested. Glacial meltwater streams have notoriously high sediment concentrations (volume of total sediment per volume of water); measured values include 0.0026 and 0.0049 from modern channels draining glaciated basins in central Asia (Chernova, 1981) and 0.001 for the concentration of suspended (only) sediment leaving the Malaspina Glacier in Alaska (Gustavson and Boothroyd, 1987). On the basis of these data, together with a range of measured ratios of suspended-sediment load to total load in glacial streams (Drewry, 1986), a minimum total sediment concentration of 0.002 is likely. With an average discharge of 3,000 m^3/s (Booth, 1986a), meltwater of the Puget lobe could transport the entire "missing" volume of its proglacial outwash plain (Fig. 10C), now reflected by the modern valleys and channels of Puget Sound, in about 5,000 years.

This time interval is 3 to 4 times longer than the known duration of ice occupation during maximum stage. However, erosion of subglacial sediment would have occurred not only at ice-maximum stage but also throughout the periods of ice advance and retreat. This interval of active-ice occupation probably persisted for about 3,000 years (Booth, 1987). Moreover, the entire sediment load probably was not removed from the lowland; sediment deposited in and subsequently eroded from the north would have redeposited farther south but still within the ultimate ice limits. This "recycling" of sediment, not recognized in the sediment-budget data from the modern glaciers, could have substantially reduced the total net volume of transported

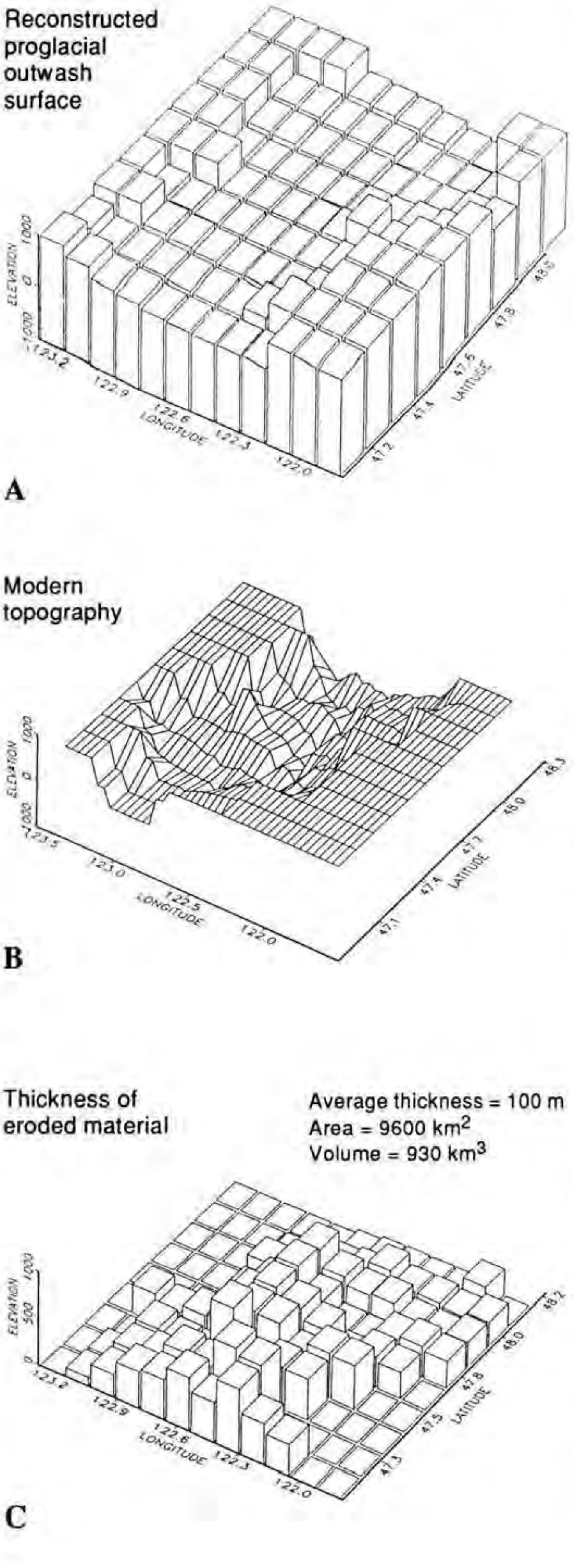

Figure 10. Calculation of the eroded volume of the "great lowland fill". Because land exposures are sparse north of the entrance to the Strait of Juan de Fuca and because the drainage system of the ice sheet would have changed significantly once the strait was filled with ice, this surface is only well defined in the central and southern Puget Lowland.

A. Original surface. Each block represents the area of one 7.5-minute quadrangle and is based on the contour map of the fill surface (Fig. 5). Elevations in meters.

B. Modern topography, truncated at 1,000 m elevation beyond the ice margin. Density of plotted data is double that of Figure 10A and is based on the local average of much denser digital elevations (meters above sea level) and more sparse bathymetric data (meters below sea level).

C. Volume of eroded material, calculated as the difference between the original fill surface (10A) and the modern topography (10B). Block areas correspond to those of Figure 10A.

sediment, depending on how far in front of the advancing ice sheet the fully aggraded outwash plain extended. On modern glaciers, for example, this need be only a few tens of kilometers. (See Boothroyd and Ashley, 1975.) With these corrections, removing the calculated missing sediment volume by subglacial meltwater alone thus appears quite plausible and probably inevitable.

Postglacial processes in the Puget Lowland are implausible alternative agents of significant landscape formation. For example, tidal currents in the modern channels of Puget Sound do not transport even chemical constituents out the Strait of Juan de Fuca (Carpenter and others, 1985), and so their ability to transport sediment would be even more limited. Holocene erosion by rivers also has only limited effect. If that erosion occurs below the reconstructed level of the outwash surface, it only redistributes the sediment but does not change the calculation of total "missing" volume. Holocene erosion of sediment from above or outside the outwash surface, with subsequent transport into the central lowland, would be significant only if the rate of landscape denudation was a substantial fraction of the 100-m thickness of sediment calculated to be "missing" from the proglacial outwash surface (Fig. 10C). In fact, denudation rates for the Cascade Range (as inferred from modern drainage-basin sediment loads, such as discussed in Nelson, 1971) integrated over the last 13,000 years are about two orders of magnitude too low to be relevant.

DISCUSSION

Processes of Drumlin and Trough Formation

With neither recognizable deformation nor abundant sorted sediment in evidence, we conclude that subglacial water was largely drained away from wherever (and whenever) drumlinization was occurring. Likely drainage routes would have existed through both thick and widespread deposits of advance outwash and in adjacent subglacial channels. Under such conditions, accreting drumlin till would therefore have maintained relative rigidity because pore-water pressures were low. Brown and others (1987) and Booth (1991b) estimate that only within about 10–15 km of the ice margin, however, were subglacial pore-water pressures significantly reduced by drainage; farther upglacier, pore pressure alone could have supported virtually all of the ice load across the ice-sheet interior because of longer and impeded drainage routes out to the ice margin.

These theory-based estimates of pore-pressure dissipation are consistent with the field evidence presented earlier. Both argue for a drumlinizing zone near the ice margin and perhaps only a few kilometers wide, within which rapidly accreting till was subjected to only minor effects of deformation and meltwater sorting. We expect that this zone, along with the upglacier limit of the drumlin field, progressively shifted northward during recession. Thus the lowland sediments now exposed display properties inherited from this last-stage retreat, but these properties do not necessarily reflect subglacial conditions during the vast majority of ice occupation.

This model contrasts with that of Shaw and others (1989), in which massive subglacial fluvial erosion was thought to have been responsible for drumlin formation. Instead, we envision that most of the subglacial meltwater was channeled into distinct troughs beneath the Puget Lobe; drumlins, in contrast, formed in well-drained areas where low basal pore pressures inhibited till deformation (Alley, 1992). Simultaneous trough excavation by subglacial meltwater and drumlin formation by till accretion on the intervening uplands would have represented a dramatic lateral gradient in subglacial processes.

Brown and others (1987) postulate that at the glacial maximum, 20–60 discrete subglacial tunnels, each several meters wide and spaced 2–5 km apart, would have been sufficient to collect and drain away all the meltwater reaching the bed. However, their estimates were based on the annual average water flux during equilibrium. During recession, the volume of meltwater production would have been even greater, especially peak flows during the melt season. We judge more likely that most of the meltwater reaching the bed of the Puget lobe during recession was confined along only a few (8–10) very large subglacial channels, which in turn excavated the deep troughs that remain a part of the lowland landscape today.

Relative Significance of Glacial-Geologic Processes

This perspective on the late-glacial erosional and depositional history of the Puget Lowland has implications for the way in which glacial processes are interpreted. First, the rate of ice-sheet advance may depend most critically not on climatic factors, but rather on rates of sediment production, transport, and deposition. As the advance outwash is generally capped by basal till, the ice sheet must have progressed in a subaerial, as opposed to a lacustrine or tide-water, environment; exposures that suggest a gradation from lacustrine sediment directly to an aqueous basal till are recognized only in unique and very localized environments (Booth, 1986b). This could occur only if ice-advance rates were no faster than progradation rates of the outwash sediment. Had the ice margin advanced beyond its apron of outwash, the dramatic increase in terminus ablation rate in an aqueous (calving) margin (Brown and others, 1982) probably would have halted ice advance. Thus the estimated rates of ice advance for the Puget lobe, 100–200 m per year (Booth, 1987), reflect only a lower bound on the independent magnitude of climatic forcing of the ice advance. The mass balance of the ice sheet, if not for potential calving at a subaqueous terminus, could have been much more positive and ice maximum achieved somewhat earlier.

This model of glaciofluvial deposition and erosion also challenges some commonly held assumptions about the relative importance of ice and water in the glacial environment. At the scale of a continental ice sheet, Drewry's (1986) tentative ranking of glacial-erosive mechanisms (his table 6.4), which places direct ice action as more significant than fluvial processes, is unchallenged. Yet in near-marginal areas, the increasing discharge of meltwater relative to the decreasing flux of ice suggests an opposite relation between these agents of geomorphic work. (See also Booth and Hallet, 1993.) The reconstructed late-glacial history of the central Puget Lowland, that of voluminous sedimentary infilling followed by locally intense linear scour, is largely a history of fluvial activity, and the modern landscape still plainly reflects those processes.

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Prehistoric Earthquakes in Western Washington

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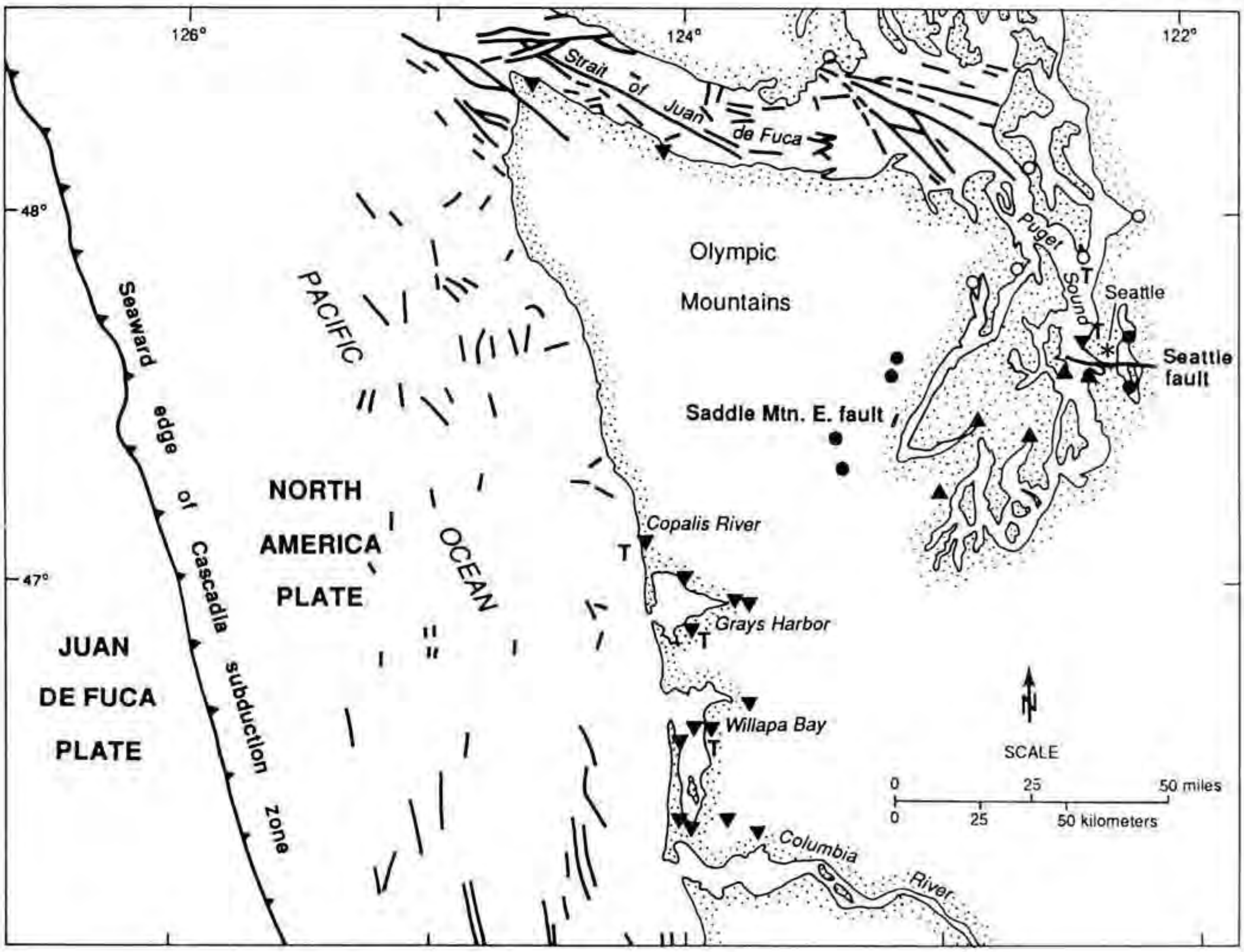
Holocene geology in western Washington gives evidence for kinds of earthquakes not represented in the region's 200 years of written history (Fig. 1, Table 1). In the Puget Sound area, coastal features (Fig. 2) and fault scarps suggest late Holocene earthquakes on historically languid structures in the North America plate, and landslides suggest Holocene shaking stronger than that caused by the largest 20th-century earthquakes in the subducted Juan de



EXPLANATION

Fault known or believed to have been active in the Holocene (dashed where approximately located; sawteeth give direction of dip of plate boundary)

- O Locality with evidence for only gradual vertical movement
- T Evidence of tsunami
- ● Slope failure – block slides and turbidity currents east of Seattle; rock avalanches in Olympic Mountains

Locality with published evidence for abrupt, probably coseismic, vertical movement in the past 2000 years

- ▲ Uplift
- ▼ Subsidence (shown only where radiocarbon dated)

Figure 1. Location of features bearing on prehistoric Holocene seismicity in western Washington (modified from Atwater, 1992).

Fuca plate. Along the Pacific coast, killed forests (Fig. 3) and buried soils imply coseismic subsidence of late Holocene age, probably from great (moment magnitude 8) earthquakes on the historically quiescent boundary between the plates. Still farther west, turbidites off southern Washington imply recurrent Holocene shaking near the seaward edge of the plate boundary. These signs of prehistoric earthquakes give paleoseismology an important role defining Washington's seismic hazards.

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Table 1. Selected reports about evidence for prehistoric earthquakes of Holocene age in western Washington

Locality (Fig. 1)	Evidence	Main interpretation	Reference
Southeastern Olympic Mountains	Scarp	Seismic slip on Saddle Mountain East fault	Wilson and others (1979)
Central Puget Sound	Steep gradient in gravity and in depth to bedrock; terminated seismic reflectors	Quaternary movement on an east-west structure (the Seattle fault) at latitude of Seattle	Gower and others (1985), Yount and others (1985), Yount and Gower (1991), Yount and Holmes (1992)
Southern Puget Sound	Emerged tidelands	Uplift; that near Seattle attended by slip on the Seattle fault	Bucknam and others (1992)
Northern Puget Sound	Sand sheets	Tsunami generated in Puget Sound by earthquake on the Seattle fault	Atwater and Moore (1992)
Southeastern Olympic Mountains	Rock avalanches	Seismic shaking	Schuster and others (1992)
East of Seattle	Landslides and turbidites	Seismic shaking	Jacoby and others (1992), Karlin and Abella (1992)
Pacific coastal estuaries and western Strait of Juan de Fuca	Buried soil of marshes and swamps, some mantled with sand	Subsidence and tsunamis from great plate-boundary earthquakes on the Cascadia subduction zone	Atwater (1987), Reinhart and Bourgeois (1987), Atwater and Yamaguchi (1991), Atwater and others (1991), Atwater (1992)
Copalis River	Intruded and extruded sand	Ground-water eruption caused by an earthquake but probably not by ordinary liquefaction	Atwater (1992)
Submarine parts of North America plate, and adjacent coast	Terminated seismic reflectors; displaced glacial till	Holocene faults	Wagner and others (1986), Wagner and Tomson (1987), Snavely (1987, p. 320)
Submarine channels near seaward edge of plate boundary west of Columbia River	Similar number of turbidites in different places	Seismic shaking, probably during great earthquakes on the Cascadia subduction zone	Adams (1990)

Figure 2. Sand bed in Seattle deposited by a tsunami about 1,000 years ago in Puget Sound. The bed is 6 cm thick; its base is at the bottom of the scale, which is 15 cm long. The bed overlies tidal-marsh mud (light tone) and peat (dark), and it underlies tidal-flat mud. The sand and some of the tidal-flat mud accumulated around growth-position stems of sedges (center) that had lived on the pre-earthquake marsh. Subsidence at the time of the tsunami led to lasting submergence that killed and preserved the sedges (Atwater and Moore, 1992). The tsunami and subsidence were probably generated by an earthquake that, according to Bucknam and others (1992), was accompanied by slip on the Seattle fault. The inferred earthquake occurred during the same season and year as did landsliding into Lake Washington (Jacoby and others, 1992).

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Figure 3. Standing dead trunks of western red cedar along the lower Copalis River (Fig. 1). The trees died about 300 years ago from tidal submergence that resulted from abrupt subsidence, and this subsidence probably accompanied a great earthquake on the Cascadia subduction zone (Atwater, 1992). Saplings of Sitka spruce live on the snag at left. Man at bottom center stands on brackish-water tidal marsh; live trees in background live on upland.

Abstracts

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DETACHMENT ORIGIN FOR REPUBLIC GRABEN, NE WASHINGTON

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The Republic graben is one of three N-trending, en echelon Eocene grabens or half-grabens that occur between two Eocene metamorphic core complexes. To the west, the Okanogan complex is flanked on its west side by top-to-the-west, low-angle mylonite and overlying detachment fault that cuts rocks as young as 50 Ma (Parrish et al, 1988). To the east, the Kettle complex is flanked on its east side by a top-to-the-east, low-angle mylonite and detachment fault that cuts middle Eocene (~50 Ma) volcanic rocks. The mylonites beneath the brittle detachment faults, formed at 10–15 km depths, project above the core complexes in the direction of the central grabens. The internal structure of the Republic graben suggests that it is floored by a low-angle, top-to-the-east, detachment fault that may have been the shallow, updip continuation of the Kettle mylonitic detachment fault prior to doming of the core complex. The Republic graben exposes a predominately west-dipping section of Permian and Eocene volcanic and sedimentary strata. The youngest Eocene strata, exposed on the west side of the graben, consist of a basal lacustrine shale facies overlain by a coarsening-upward sandstone to conglomerate sequence with a western source. An isopach map of the basal lacustrine shale facies reveals an east-thinning wedge, thinning from over 1100′ to 0′ in a one mile distance. The Eureka Creek fault (east side down) bounds the lacustrine section on the west. The fault was host to epithermal Au mineralization (50.2 ± 0.3 Ma, $^{40}Ar/^{39}Ar$ adularia) during initial growth of the lacustrine basin; debris flows at the base of the lacustrine section contain epithermal vein clasts. This suggests that the fault surface was exposed during lake deposition and that the lake basin was fault-controlled. The wedge shape of the lake basin deposits and the westward tilt of the underlying strata suggest that the fault has a listric geometry. A second detachment fault, exposed several miles to the east within the graben ("Lambert thrust" of Muessig, 1967), offsets stratigraphy 3–5 miles to the east along a low-angle (<10°) zone of breccia. The Eureka and Lambert detachments are probably splays off a master detachment which locally defines the eastern margin of the graben (St. Peter fault). This fault dips 15°W and is cut by a pair of steep, en echelon, west-side-down, normal faults that define the eastern graben margin for most of its length. The St. Peter fault projects to the east over the Kettle complex, and is inferred to continue as the top-to-the-east mylonite zone below the brittle Kettle detachment fault. Similar cases can be made for detachment origins for the Toroda Creek graben (at the updip end of the top-to-the-west Okanogan detachment) and the Keller Creek graben (updip of the Kettle–Lincoln dome detachment).

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AGE AND TECTONIC EVOLUTION OF THE OLYMPIC SUBDUCTION COMPLEX AS INFERRED FROM FISSION-TRACK AGES FOR DETRITAL ZIRCONS

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The Olympic Mountains of NW Washington State expose the Olympic subduction complex (OSC), which consists of an imbricated assemblage of Cenozoic sandstone, mudstone, and minor pillow basalt. We have employed the fission-track (FT) method to better constrain the timing of deposition, subduction accretion, and metamorphism of the central and eastern OSC, which until now were poorly dated.

Of the 15 sandstone samples that were dated, 11 are unreset and thus preserve detrital grain ages related to cooling events in the original source terrain from which the zircons were derived. The youngest grain ages provide a maximum limit for the depositional age of the sandstone. Conversely, the reset samples provide a minimum age for deposition and also can be used to infer the timing of cooling and exhumation of the OSC. For the unreset samples, FT ages for the youngest population of grains in each sample range from 48 to 18 Ma. For the reset samples, the FT data indicate cooling at about 14 Ma from temperatures in excess of 240°C.

Overall, the FT data define a radial pattern with reset rocks in the central part of the OSC, Late Oligocene and Early Miocene clastic rocks lying just outside the perimeter of the reset area, and Middle Eocene to Early Oligocene clastic rocks at a still greater distance. This pattern is consistent with a domal structure for the OSC, which appears to have formed during Neogene uplift of the modern Olympic Mountains. The oldest rocks lie in the flanks of the uplift. The youngest and most deeply exhumed rocks are located in the center of the uplift and coincide with the area of the highest mountainous relief.

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CORDILLERAN TECTONIC SETTING OF WASHINGTON

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Some of the rock units and structures in Washington are representative of Cordillera-wide elements. In the NE quadrant, these include the Belt–Purcell, Windermere, and Paleozoic miogeoclinal rocks; structures recording the Antler orogeny and an Early or Middle Jurassic event, during which volcanic-arc and ocean-floor assemblages were shoved eastward onto the continental margin; and core complexes denuded by Eocene extension. Some late Mesozoic elements and events that are well represented in the US Cordillera (especially California) are either lacking or are greatly modified in Washington and the remainder of the Northwest Cordillera in British Columbia: there is no record of a Nevadan orogeny; except for some tiny scraps in the San Juan–Cascade thrust system, the Franciscan subduction complex and Great Valley fore-arc sequence are absent; and there is no single locus of magmatic-arc activity as there is in the Sierra Nevada and Klamath Mountains. Instead, late Mesozoic plutonic-volcanic belts continue into British Columbia as the Coast Plutonic Belt and Spences Bridge–Chilko arc, and the San Juan–Cascade thrust system lies at the southern end of a 1200-km zone of mid-Cretaceous shortening extending into SE Alaska. As in British Columbia, the Cretaceous paleolatitude of all of these elements in the NW quadrant is unresolved; they may have lain >1,500 km farther south until latest Cretaceous time, according to paleomagnetic data.

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STRATIGRAPHY AND BIOSTRATIGRAPHY OF THE PALEOZOIC AND EARLY MESOZOIC ROCKS OF SAN JUAN ISLANDS AND NORTHWESTERN CASCADE MOUNTAINS, WASHINGTON

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The stratigraphy and paleontology of Paleozoic and early Mesozoic rocks of the San Juan Islands and adjacent parts of the foothills of the Northern Cascade Mountains to the southeast, bear a striking resemblance to the Cache Creek (Pacifica) terrane, oceanic ridge sequence of central British Columbia and are believed to be a displaced southern extension of this terrane. Early Paleozoic basement complex and Devonian rocks found on the San Juan Islands are not known in the Cache Creek terrane to the north but the Mississippian, Permian, Triassic and early Jurassic rocks have similar stratigraphy and paleontology though the section is condensed.

The western part of the Northern Cascade Mountains contains a stratigraphy and paleontology (Chilliwack Group) similar to that of the Quesnellia terrane lying east of the Cache Creek terrane in central British Columbia and in particular correlates with the Harper Ranch Group near Kamloops, B.C.

The San Juan Islands contain the typical Permian Tethyan faunas of the Cache Creek Group and the basalt, thick radiolarian chert sequence and pure limestones of an oceanic ridge. There is no recognizable break between the Triassic and the Permian.

The western part of the Northern Cascade Mountains contains the non-Tethyan, no-American, Western Cordilleran fauna, and the andesites, dacites and basalts, sandstones and shales and argillaceous limestones of an island arc sequence. There is a major Permian–Triassic unconformity.

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THE STRATIGRAPHIC RECORD OF MID-CRETACEOUS OROGENY IN THE METHOW BASIN, WASHINGTON AND BRITISH COLUMBIA

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Stratigraphic, structural, and provenance studies conducted over the last decade within the Methow basin have elucidated the nature of basin forming and filling processes and have implications for the broader problems of late Mesozoic tectonic history in the Pacific Northwest. Key aspects of the basin history that have long been recognized are (a) accelerated subsidence in early Albian time, (b) two-sided basin filling and transition from deep-water to alluvial conditions during Albian and ?Cenomanian time, and (c) folding and faulting of basin strata during late Cretaceous and Tertiary time. More recently, several observations have shed light on the details of basin history.

Most importantly: (a) rapid subsidence of the basin in early Albian time probably accompanied rapid uplift of the eastern margin of the basin along an ancestral Pasayten fault; (b) relations in correlative strata in the Tyaughton basin in southern B.C. indicate that the Methow and Tyaughton basins were separated by an uplifted axial high, the Bridge River–Hozameen terrane, during deposition; (c) a region-wide mid-Cretaceous unconformity is locally angular and everywhere separates alluvial sediments and volcanic rocks from underlying marine and marine-marginal strata.

The mid-Cretaceous unconformity clearly demonstrates that compressional deformation, probably rooted in the concurrently evolving Cascades and Coast Mountains orogen to the west, had begun to encroach on the basin between the middle Albian and 86 Ma. Strata above the unconformity were probably deposited in an intermontane trough. The mechanism by which the Bridge River–Hozameen high was structurally elevated is uncertain, but may have been driven by early stages of regional compression. Greater uncertainty surrounds the mechanism of early Albian subsidence. Early phases of regional compression may have loaded one or both margins of the basin. It is also possible that the eastern boundary of the basin was a normal fault in Albian time, formed either by regional extension or by reactivation of an old break due to flexural bending and loading.

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PRE-TERTIARY STRATIGRAPHY AND MULTIPLE OROGENY IN THE WESTERN NORTH CASCADES, WASHINGTON

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New fossil data from the western North Cascades, coupled with our 1:100K mapping and existing fossil and radiometric ages, confirm the existence of several distinct terranes and indicate multiple orogenies.

Undisrupted blueschist-facies rocks of the Early Cretaceous Easton Metamorphic Suite appear to be structurally highest pre-Tertiary rocks in the northwest Cascade system (NWCS). The regional Bell Pass melange locally underlies the Easton Suite. Most of the melange is Elbow Lake Formation, characterized by ribbon chert of Pennsylvanian(?) to Late Jurassic (but mostly Permian and Triassic) age, basalt tuff, and argillite. These rocks are commonly associated with dark, chert-lithic sandstone of unknown age. Other melange components are early Paleozoic and older crystalline rocks of the Yellow Aster Complex; Permian schists of the Vedder Complex; and ultramafic rock. The Yellow Aster commonly lies at the base of the Bell Pass melange; other crystalline elements commonly are at its top. The Twin Sisters Dunite is separated by high-angle faults from most other units but can be construed to be within the Bell Pass melange or even above the Easton Suite. Largely beneath the Bell Pass melange are volcaniclastic, volcanic, and carbonate strata of the late Paleozoic Chilliwack Group. Associated with Chilliwack strata are volcaniclastic and volcanic strata of the Late Triassic and Jurassic Cultus Formation. Mostly upright and moderately dipping clastic strata of the Late Jurassic and Cretaceous (as young as middle or Late Cretaceous?) Nooksack Group lie structurally beneath tectonically disrupted Chilliwack–Cultus strata and Bell Pass melange. At its base the Nooksack interfingers with pyroclastic rocks of the Middle Jurassic Wells Creek Volcanics.

Helena–Haystack melange (HHM) separates NWCS from the western and eastern melange belts (WEMB) to the SW. The eastern belt is dominated by Triassic chert and basalt and the western by Late Jurassic to Early Cretaceous clastic rocks, though both contain large blocks of older rock. Differing structural styles and lack of lawsonite and Easton rocks within WEMB distinguish WEMB from NWCS.

Age, facies, and structural relations require 3 or 4 significant orogenies. Widespread overturning, structural disruption, and mixing within the Chilliwack–Cultus and Bell Pass units suggest (1) a pre-Cascade, arguably Late Jurassic, event. (2) Mid-Cretaceous Cascade orogeny juxtaposed the Easton, Bell Pass + Chilliwack–Cultus, and Nooksack–Wells Creek packages. WEMB may have been emplaced during Cascade orogeny, but fragments of both NWCS and WEMB within the HHM suggest mixing during (3) a later Cretaceous or earliest Tertiary event. All structural relations were greatly modified by (4) post-middle Eocene orogeny.

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TOWARD A BETTER UNDERSTANDING OF THE PALEOGENE PALEOGEOGRAPHY OF THE PUGET LOWLAND, WESTERN WASHINGTON

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The Puget Lowland was the site of Eocene dextral strike-slip faulting along the Puget fault zone, a structure inferred from patterns and contrasts in basin evolution, and from facies and petrographic mismatches. This fault forms the boundary between pre-Tertiary basement rock of the Cascades to the east and early and middle Eocene, marine, basaltic basement rock (Crescent Formation) of the Coast Range to the west. The fault zone extends northward from Mt. St. Helens beneath the Northcraft volcanic pile, through the Puget Lowland and Puget Sound, then bends northwestward into a zone of transpressive, south-directed thrust faults on southern Vancouver Island. The fault zone was initiated about 55 Ma when the Crescent was erupted in a marginal rift basin. Northward movement of the Coast Range probably continued into at least the Oligocene. Eocene–Oligocene subsidence of Crescent basement in southwest Washington is mainly of thermal origin; much higher subsidence rates for Crescent basement on the northern Olympic Peninsula require an additional driving force, inferred to be loading from southern Vancouver Island thrust sheets.

Some offset on the Puget fault zone was transferred eastward to the Straight Creek fault, leading to development of a broad, rapidly subsiding, pull-apart basin in the eastern Puget Lowland and Cascade foothills. Local sub-basins, such as the prominent gravity low near Seattle, formed within or adjacent to the basin-margin fault zones. Fluvial systems carrying sediment derived from distal crystalline and local metasedimentary sources entered the basin from the east and north. At times, intrabasinal volcanic centers disrupted drainage patterns. Eocene shorelines in the basin trended northwest and migrated to the southwest. Intrabasinal tectonics is reflected in local unconformities and abrupt facies change. Facies in the Raging River Formation, for example, change upward through an interval of 400 m from braided-fluvial conglomerate to bathyal mudstone to deltaic sandstone.

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ALPINE GLACIATION OF WESTERN WASHINGTON

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The oldest evidence of mountain glaciation in Washington is found at the crest of the southern Cascade Range where alpine till is overlain by lava dating 1.75 myr. Subdued and weathered drifts in the northern and southern Cascades and the southwestern Olympic Mountains record undated ice advances preceding the last glaciation. Based on relative-age criteria, these drifts likely were deposited during the last 1 myr; a till in the southern Cascades probably dating to this interval is >0.65 myr old. Glaciers that mantled the Cascades and Olympics prior to ca. 1 myr ago likely were shorter and thinner than subsequent glaciers; in most areas their record either is missing or has not been recognized.

The undated penultimate glaciation may correlate either with marine isotope stage 4 or 6; cosmogenic isotope dating presently underway may help resolve the chronology. During stage 3, when the average glacier cover probably resembled that at the close of the Pleistocene, sediment-laden meltwater streams deposited the Kitsap Formation in the adjacent Puget Lowland.

The age of the last glacial maximum in the Olympics and Cascades has not yet been closely constrained. Widespread late-glacial moraines probably date between 12,000 and 10,000 yr ago, but their possible correlation with the European Younger Dryas oscillation has not been established, nor has the possibility of a regional readvance of glaciers during the early Holocene. Neoglacial drift of middle to late Holocene age is found beyond many existing glaciers; the youngest moraines correlate with Little Ice Age deposits found worldwide.

During the last glacial maximum, glacier equilibrium-line altitudes in the Cascades lay ca. 900 m below present levels. Estimated mean-annual temperatures at that time are consistent with temperature estimates derived from global climate-model simulations that suggest cold, dry conditions during full-glacial times.

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TERTIARY MAGMATISM AND TECTONISM IN AN E–W TRANSECT ACROSS THE CASCADE ARC IN SOUTHERN WASHINGTON

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A section of lava flows and volcaniclastic rocks more than 7 km thick and ranging in age from >36 Ma to 16 Ma crops out in broad NW- to NNW-trending folds along a 100-km transect from west of Mount St. Helens (MSH) to the Cascade crest in the Goat Rocks area. The lower part of the section intertongues with fluvial micaceous arkose 55 km northeast of MSH and with near-shore marine and nonmarine micaceous and coal-bearing arkose between Kelso and Castle Rock west of the mapped area. The magmatic rocks are predominantly low- and moderate-K basaltic andesite and andesite but include local accumulations of calc-alkaline basalt and a 1-km-thick pile of E-type MORB-like tholeiite near the base of the section 20 km west of MSH. Dacite and rhyolite form three local units in the middle of the section, as well as isolated flow and domes. No regional stratigraphic marker has been found. The section is intruded by the 21-Ma Spirit Lake pluton and many finer-grained dikes, sills, and other bodies. An extensive swarm of ENE-striking dikes 25 km northeast of MSH may reflect the regional paleostress field *ca.* 20–25 Ma.

Most of the folding took place after emplacement of the Spirit Lake pluton and before intrusion of hornblende-phyric silicic andesite and mafic dacite about 13–11 Ma. These intrusions, which record the first major appearance of hornblende-bearing magma in the area, occur as radial dikes and sills sprouting from the mineralized hornblende quartz diorite of McCoy Creek 35 km northeast of MSH, and as isolated sills farther east. Erosion related to regional Cascade uplift (defined by tilted 17–15-Ma Grande Ronde Basalt east of the Cascade crest) removed an unknown thickness of section. No rocks within the transect have been dated between 10 Ma and 5 Ma, and only the hornblende-bearing intrusions have ages between 15 Ma and 10 Ma; however, evidence for 12–6 Ma hornblende-bearing eruptions within or near the transect is found in volcaniclastic deposits flanking the Cascades, such as the Ellensburg and Mashel Formations. Whether the paucity of rocks younger than 17–15 Ma reflects decreased rate of volcanism, lack of preservation owing to uplift and erosion, or both (perhaps most likely) is unclear. More age determinations for intrusions, commonly altered and hard to date, are needed to resolve this issue.

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THE OKANOGAN RANGE BATHOLITH, NORTH-CENTRAL WASHINGTON: ROOT OF A LATE JURASSIC(?)–EARLY CRETACEOUS CONTINENTAL-MARGIN ARC

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The west-central Okanogan Range batholith (ORB) comprises a 125–112 Ma (U-Pb zircon and monazite) trondhjemitic suite bounded on the west by the Pasayten fault (PF) and on the east by metamorphic rocks of Permian–Triassic(?) protolith age. The NNW orientation of the batholith and its elongated plutons (parallel to the PF) indicate that intrusion took place during dip-slip on the PF. The ORB is calcic (Peacock index 64), with mean SiO_2=68%, high Al_2O_3 (17%), Na_2O(5%), Sr(851 ppm) and Ba(806), and low K_2O(1.4%), Rb(23), Nb(3), Y(8), Th(2), U(<2), Pb(10), Ni(5) and Cr(5). Initial $^{87}Sr/^{86}Sr$ ratios (Sr_i) increase eastward from 0.7036 to 0.7042. Late Jurassic–Early Cretaceous (K-Ar hbl) quartz diorite and tonalite plutons (MBP) intruded marine strata of the Methow basin west of the PF. Crystallization ages of MBP are poorly constrained and there may be a significant age gap between MBP and ORB. MBP are calcic with mean SiO_2=60%, high Al_2O_3(16.7%), and compared with ORB, similar trace element abundances except for lower Sr(510 ppm), moderate Ba(423) and higher Y(24), Ni(19), Cu(64), Cr(34) and V(173). Sr_i is ~0.7035. Calcic compositions and low Sr_i of both groups require a primitive mafic source (basaltic lower crust or upper mantle). Differences in trace element abundances between ORB and MBP probably reflect differences in their source and(or) magmatic history but do not exclude the possibility that they are parts of a single long-lived subduction-related magmatic arc. The source of MBP may have been newly subducted oceanic lithosphere interacting with a mantle wedge. In the Early Cretaceous, the zone of melting migrated eastward encountering early Mesozoic crust as the depth of the subducted slab increased giving rise to more evolved magmas (ORB). The PF may have been initiated as a shear zone located in the vicinity of an older crustal boundary (Late Jurassic continental margin or suture?).

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ANOTHER LOOK AT THE FRASER RIVER–STRAIGHT CREEK FAULT (FRSCF)

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The FRSCF is a major Middle–Late Eocene dextral strike-slip fault. It originated as a N–S shear which cuts a system of earlier NW-trending strike-slip faults. Offset of the NW-trending Yalakom and Ross Lake faults is ~105 km at the north end of the FRSC structure, while that of the Darrington–Devils Mtn. (DDM) and Cle Elum (CE) faults is ~85 km a the south end. Reversal of this displacement restores many regional geologic units to close proximity: the Tyaughton and Methow basins; Eocene reset metamorphic rocks of the Bridge River Group and North Cascades core; the main Easton metamorphic belt and its southeastern tail; and, displaced belts of the Helena–Haystack melange. Subequal offset at both ends of the FRSC refutes larger displacements (190–140 km) inferred from the apparent offset of the Nason and Settler units (Misch, 1977).

Geologic relations on southern segment of the FRSCF constrain the movement history of the fault. Initiation of the SCF is signaled by folding of the Swauk Fm at ~48 Ma. This folding was localized in the restraining bend of the DDM–CE fault and the nascent SCF. After 48 Ma the SCF became the master structure and subsequent movement on the NW-trending faults was largely dip-slip. A belt of detrital kyanite in Late Eocene sandstones in SW Washington still projects northeast across the SCF to its source in the Nason terrane, thus ruling out significant strike-slip on the SCF since ~41 Ma. Similarly, late Middle Eocene conglomerates trapped along the SCF in northern Washington still lie adjacent to their source, the Marblemount metaquartzdiorite. Final major movement on the SCF is related to the folding of the Naches Fm after ~40 Ma as was dominantly dip-slip. Movement on the FRSCF had largely ceased prior to emplacement of early plutons of the Cascade magmatic arc at ~36 Ma.

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