The cordilleran ice sheet

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Introduction

The Cordilleran ice sheet, the smaller of two great continental ice sheets that covered North America during Quaternary glacial periods, extended from the mountains of coastal south and southeast Alaska, along the Coast Mountains of British Columbia, and into northern Washington and northwestern Montana (Fig. 1). To the west its extent would have been limited by declining topography and the Pacific Ocean; to the east, it likely coalesced at times with the western margin of the Laurentide ice sheet to form a continuous ice sheet over 4,000 km wide. Because most of the marginal environments of the Cordilleran ice sheet were not conducive to preserving an extensive depositional record, much of our understanding of this ice sheet has come from limited areas where preservation is good and access unencumbered, notably along its lobate southern margin in northern Washington State and southern British Columbia.

Arrival of geologists into Puget Sound late in the 19th century initiated study of the Cordilleran ice sheet. The landscape displayed unmistakable evidence of past glaciations, but a sporadic sequence of deposits along valley walls and coastal bluffs only hinted at a long and intricate history of ice-sheet occupations. By the mid-20th century extensive field studies had developed a framework for Pacific Northwest Quaternary history. Evidence of four glaciations, summarized by Crandell (1965) and detailed by Armstrong *et al.* (1965), Mullineaux *et al.* (1965), and Crandell (1963), followed the precedent from the American Midwest: four continental-scale glaciations, correlated across broad regions. In the Pacific Northwest, the youngest ice-sheet glaciation (Fraser), was constrained by radiocarbon dates and correlated with the Wisconsin Glaciation of the mid-continent. Earlier glaciations (given the local names *Salmon Springs*, *Stuck*, and *Orting*) were identified only in the southeastern Puget Lowland. Crandell (1965) suggested that they spanned early through late Pleistocene time.

In the latter part of the 20th century, improved understanding of global and regional stratigraphy, and emphasis on geomorphic processes, have brought new information from studies of the Cordilleran ice sheet. These advances are the topics of this chapter. The record of global warming and cooling recorded in deep-sea cores shows that there were many glaciations during the Quaternary Period, not just four. Global perspectives on past sea-level variations prove critical to understanding tidewater glacier systems like the southwestern part of the Cordilleran ice sheet. New dating techniques yield crude but consistent chronologies of local and regional sequences of alternating glacial and nonglacial deposits. These dates secure correlations of many widely scattered exposures of lithologically similar deposits and show clear differences among others.

Besides improvements in geochronology and paleoenvironmental reconstruction (i.e. glacial geology), glaciology provides quantitative tools for reconstructing and analyzing any ice sheet with geologic data to constrain its physical form and history. Parts of the Cordilleran ice sheet, especially its southwestern margin during the last glaciation, are well suited to such analyses. The approach allows interpretation of deposits and landforms at the now-exposed bed of the former ice sheet, and it also suggests likely processes beneath other ice sheets where reconstructions are less well-constrained.

Finally, expressions of the active tectonics of western North America are now widely recognized across the marginal zone of the Cordilleran ice sheet. Such conditions were little appreciated at mid-century. Only since the 1980s have the extent and potential influence of recent tectonics on the landscape of western Washington been appreciated. The regional setting for repeated glaciations owes much of its form to those tectonic influences; conversely, deformation and offset of ice-sheet deposits may be critical in unraveling the Quaternary expression of the region's tectonics.

Perhaps the greatest development in recent study of the Cordilleran ice sheet, especially its southwestern boundary, has been the focus of scientific attention on this region – not only by geoscientists but also by resource managers, land-use planners, and the general public. In the last several decades, this glacial landscape has become a region of rapid population growth. In part because of these social pressures, the level of scientific study here has rapidly increased, which will likely render the story of the Cordilleran ice sheet presented in this synoptic paper even more quickly outdated than its predecessors.

Chronology and the Stratigraphic Record

Quaternary Framework

More than one hundred years after Bailey Willis published "Drift Phenomena of Puget Sound" (1898), geologists continue efforts to identify and correlate the Quaternary stratigraphic units across the area episodically covered by the southern part of the Cordilleran ice sheet (Fig. 1). Nearly

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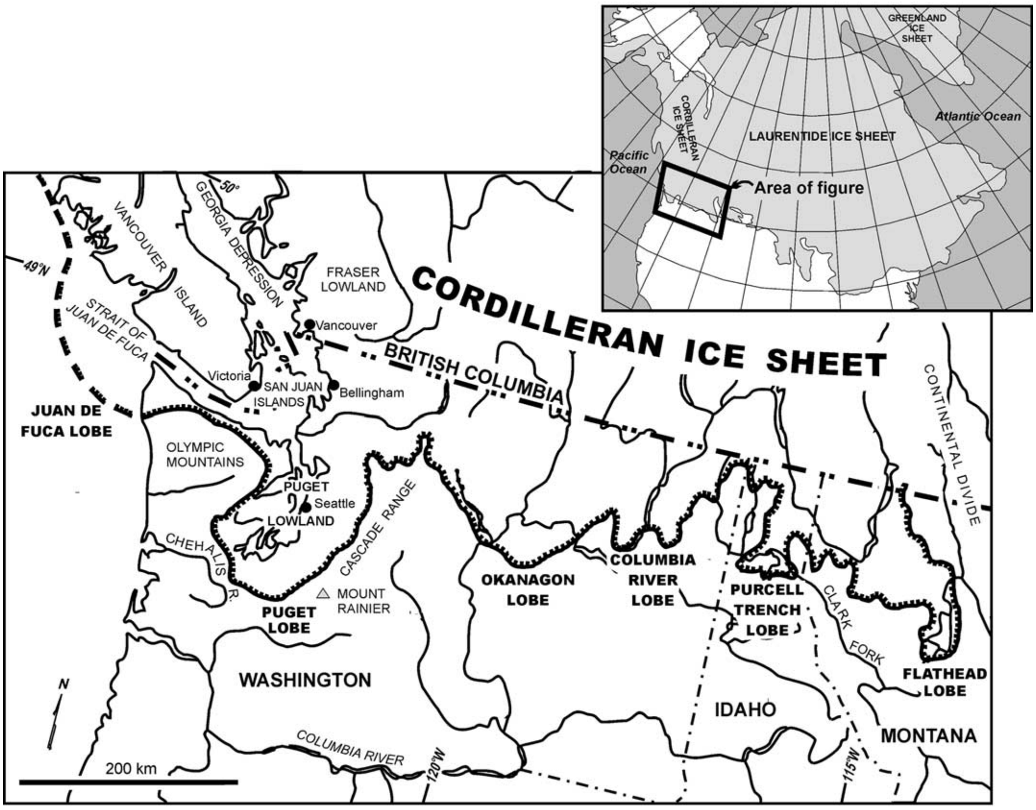


Fig. 1. Map of southern extent and lobes of the latest Pleistocene advance of the Cordilleran ice sheet in Washington and British Columbia.

a half century of field investigations in the southern Puget Lowland (Armstrong *et al.*, 1965; Crandell *et al.*, 1958; Mullineaux *et al.*, 1965; Noble & Wallace, 1966; Waldron *et al.*, 1962) and in the northern Puget Lowland (Clague, 1981; Easterbrook, 1986, 1994) show that ice sheets have advanced south into the lowlands of western Washington at least six times. The global climatic template of the marine-isotope record illustrates the likely number and frequency of glacier advances. It suggests that the current half-dozen known glacier advances do not include every advance into the region in the last 2.5 million years. The last three ice advances correlate with marine oxygen isotope stages (MIS) 2, 4, and probably 6 (Fig. 2). The most recent advance was the Fraser glaciation, discussed in detail later in this chapter.

Little is known about the climate in the lowlands of southern British Columbia and western Washington during most of the Pleistocene; but recent research has focused on either MIS 2 (Hansen & Easterbrook, 1974; Heusser, 1977; Heusser *et al.*, 1980; Hicock *et al.*, 1999; Mathewes & Heusser, 1981; Whitlock & Grigg, 1999), MIS 2 and 3 (Barnosky, 1981, 1985; Grigg *et al.*, 2001; Troost, 1999), or MIS 5 (Heusser & Heusser, 1981; Muhs *et al.*, 1994; Whitlock *et al.*, 2000). From these studies we know climate during MIS 3 was somewhat cooler than today and sea level was lower. The climate of MIS 5 was similar to today's, with marine deposits commonly found near modern sea level.

Recognition of nonglacial environments in the depositional record is essential to unraveling the chronology here. The present Puget Lowland may be a useful analog for earlier nonglacial periods. Areas of nondeposition, soil formation, or minor upland erosion dominate most of the lowland (Fig. 3). Sediment is only accumulating in widely separated river valleys and lake basins, and in Puget Sound. Were the present lowland again invaded by glacier ice, it would bury a complex and discontinuous nonglacial stratigraphic record. Thick sedimentary sequences would pinch out abruptly against valley walls. Sediment deposited in valleys could be 100 m lower than coeval upland sediment or organic-rich paleosols. Thus, the thickness and lateral continuity of nonglacial sediment of any one nonglacial interval will be highly variable owing to the duration of the interval, subsidence and uplift rates, and the altitude and surface topography of fill left by the preceding glacier incursion (Troost, 1999).

West of the Cascade Range, Cordilleran glaciations were typified by the damming of a proglacial lake in the Puget Sound basin, the spreading of an apron of outwash, deep subglacial scouring and deposition of till, formation of large recessional outwash channels, formation of ice-contact terrain, and deposition of glaciomarine drift in the northern lowland. Glacial periods were marked by a change to colder-climate vegetation and increased deposition and erosion. Thick glaciomarine, glaciolacustrine, and outwash deposits accumulated in proglacial and subglacial troughs, capped

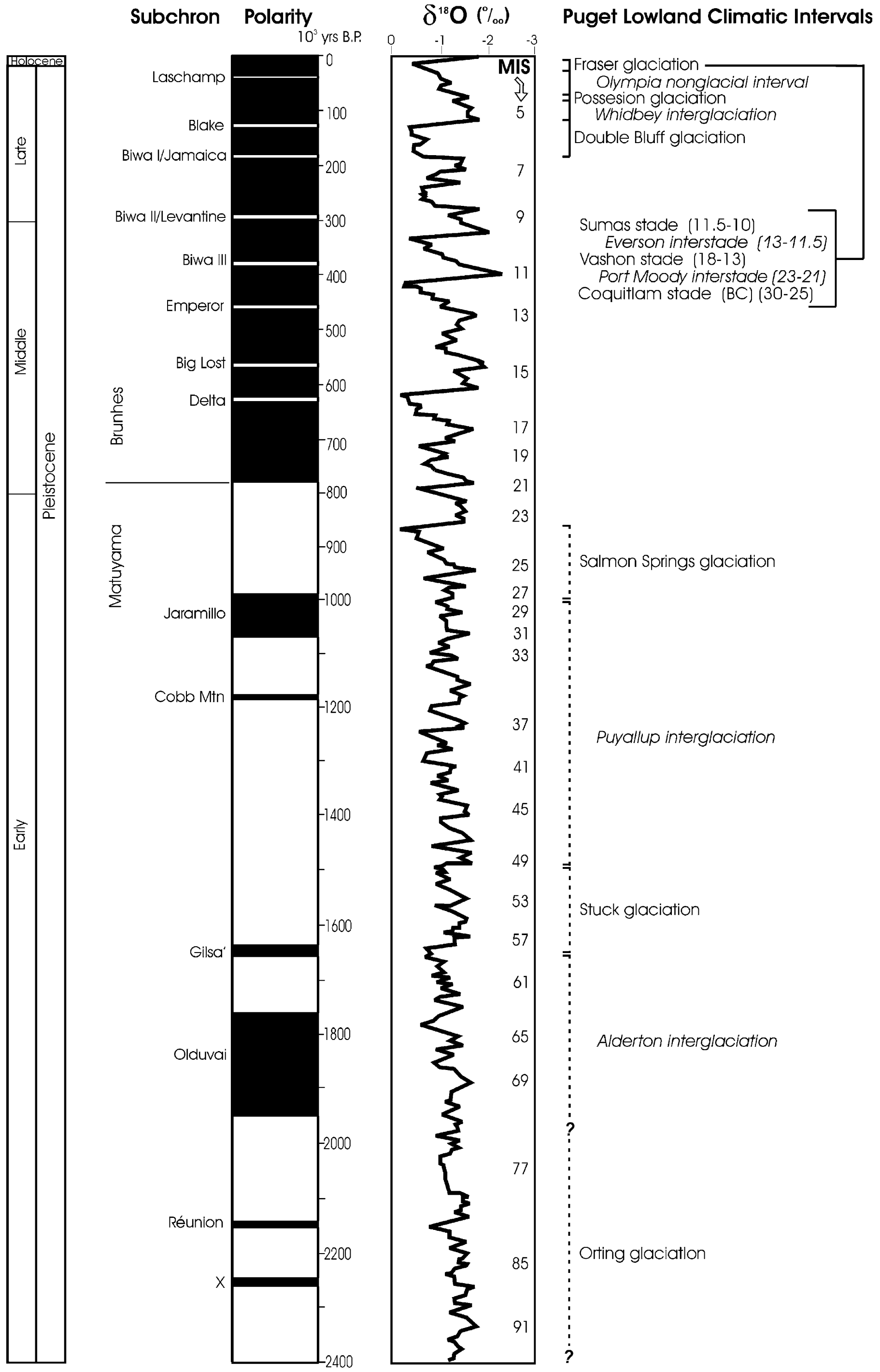


Fig. 2. Comparison of the marine oxygen-isotope curve stages (MIS) using the deep-sea oxygen-isotope data for ODP677 from Shackelton et al. (1990), global magnetic polarity curve (Barendregt, 1995; Cande & Kent, 1995; Mankinnen & Dalrymple, 1979), and ages of climatic intervals in the Puget and Fraser lowlands. Ages for deposits of the Possession glaciation through Orting glaciation from Easterbrook et al. (1981), Easterbrook (1986), Blunt et al. (1987), and Easterbrook (1994). Additional ages for deposits of the Puyallup Interglaciation from R.J. Stewart (pers. comm., 1999). Ages for the Olympia nonglacial interval from Armstrong et al. (1965), Mullineaux et al. (1965), Pessl et al. (1989), and Troost (1999). Ages for the Coquitlam Stade from Hicock & Armstrong (1985); ages for the Port Moody Interstade from Hicock & Armstrong (1981). Ages for the Vashon Stade from Armstrong et al. (1965) and Porter & Swanson (1998). Ages for the Everson Interstade from Dethier et al. (1995) and Kovanen & Easterbrook (2001). Ages for the Sumas Stade from Clague et al. (1997), Kovanen & Easterbrook (2001), and Kovanen (2002).

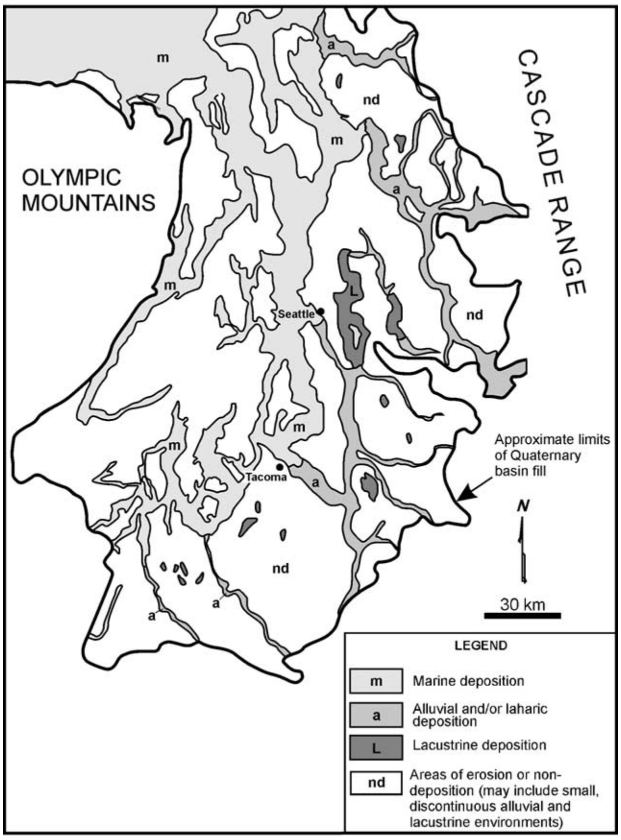


Fig. 3. Modern Puget Lowland depositional environments, providing one example of the extent of deposition during interglacial periods. Most of the land area is either erosional or non-depositional (except for minor upland soil formation). Modified from Borden & Troost (2001).

intermittently by subglacial till, predominantly of meltout origin. Likewise, subglacial drainage carved deep erosional troughs subsequently filled with postglacial volcanic debris flows and alluvium. Thus, there are many unconformities and buried topographies in the stratigraphic record.

Sediment lithology helps differentiate glacial from nonglacial deposits, given that source areas for glacial deposits are usually other than the headwaters of the current streams. This technique, although not new, is finding renewed use for proposing late Pleistocene glacier readvances (Kovanen & Easterbrook, 2001) and for interpreting bulk geochemistry analyses of central lowland deposits (Mahoney *et al.*, 2000).

Tectonic Setting

Plate movement of western North America governs the structural setting of the southwestern margin of the Cordilleran ice sheet. The Juan de Fuca plate (JDF) moves northeast and subducts beneath the North America plate at about 4 cm per year (Fig. 4a). From strike-slip plate movement farther south and crustal extension across the Basin and Range province, a series of crustal blocks between northern Oregon and southern British Columbia are colliding with the relatively fixed buttress of Canada's Coast Mountains (Wells *et al.*, 1998). The region is shortening N-S by internal deformation of the blocks and by reverse faulting along block boundaries.

Both the bedrock and overlying Quaternary sediment in the Puget Lowland have been deformed by faults and folds as a result of this tectonic activity. The Seattle fault is one of several active structures of the Puget Lowland showing displacement in the last 10,000 years. It separates the Seattle basin from the Seattle uplift, two of the structural blocks involved in the shortening in Oregon and Washington (Fig. 4b). Its displacement history embraces about 8 km of south-side-up movement since mid-Tertiary time (Johnson *et al.*, 1994; Pratt *et al.*, 1997), including 7 m of uplift during a great earthquake 1,100 years ago (Atwater & Moore, 1992; Bucknam *et al.*, 1992). This fault may have moved several times in the last 15,000 years; episodic movement throughout the Quaternary is likely, although not yet documented. Current investigations suggest that a similar fault may pass west-northwest near Commencement Bay at Tacoma (Brocher *et al.*, 2001). Other faults trending east-west or southeast-northwest cross the glaciated lowlands both north and south of the Seattle fault (Johnson *et al.*, 1996, 2001; Pratt *et al.*, 1997), with likely displacements of meters to tens of meters, thereby complicating interpretation of the Quaternary stratigraphic record.

Evidence of Pre-Fraser History and Depositional Environments

Puget Lowland

Abundant but fragmentary evidence of pre-Fraser glacial and interglacial deposition in the Puget Lowland exists in many geologic units named and described at type sections (Table 1, Figs 2 and 5). Because the evidence is scattered and discontinuous, reconstructions of pre-Fraser depositional environments and climate are sparse. Only the two latest nonglacial periods (MIS 3 and 5) are well known through abundant organic-bearing sediments and good exposures.

Evidence of nonglacial deposition during MIS 3 (broadly coincident with the Olympia nonglacial interval, defined by Armstrong *et al.*, 1965) has been found in bluff exposures and boreholes across the Puget Lowland. These deposits accumulated between about 70,000 yr ago and 15,000 ^{14}C yr B.P.; although a time-stratigraphic unit, the Olympia interval also has a defined type section at Fort Lawton in Seattle (Mullineaux *et al.*, 1965). During MIS 3, most of the lowlands of Washington were ice-free, allowing for subaerial deposition and weathering. Deposits of the Olympia nonglacial deposits (informally named the Olympia beds in western Washington and the Cowichan Head Formation in southwestern British Columbia) thus consist of peat, tephra, lahars, mudflows, lacustrine, and fluvial deposits (Fig. 6). Dozens of radiocarbon dates from this interval confirm nonglacial conditions from about 15,000 ^{14}C yr B.P. to beyond the limit of radiocarbon dating (ca. 40–45,000 ^{14}C yr B.P.) (Borden

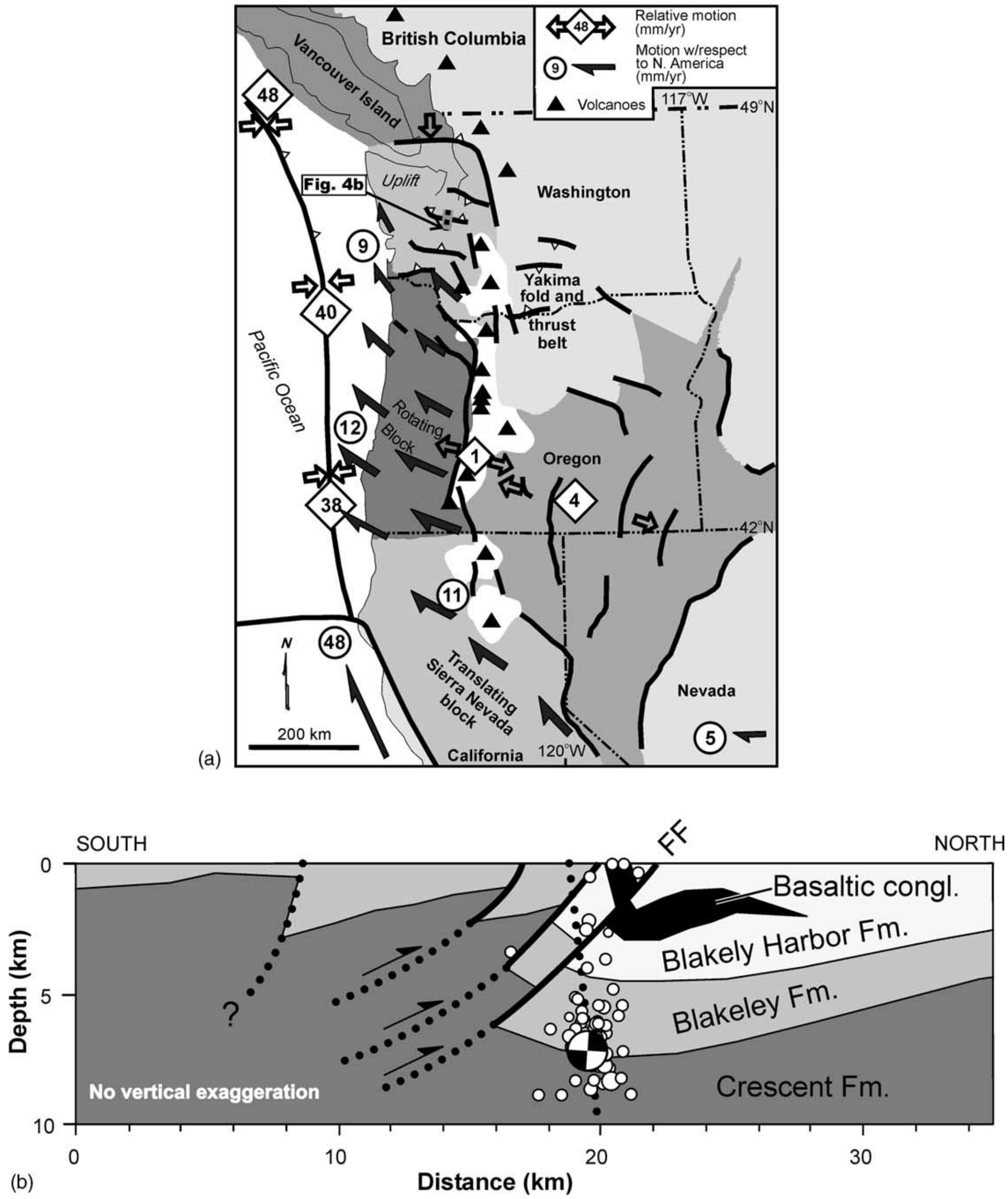


Fig. 4. Crustal blocks and major structures in the Puget Lowland, showing the north-verging compressional motion and the resultant displacement across the Seattle fault zone. Fig. 4a shows the relative motion of the western United States as transferred to western Washington (modified from Wells et al., 1998). Fig. 4b interprets the Seattle fault zone as a series of south-dipping reverse faults (FF = frontal fault; modified by B. Sherrod [USGS] from Johnson et al. (1994).

& Troost, 2001; Clague, 1981; Deeter, 1979; Hansen & Easterbrook, 1974; Troost, 1999). Paleoecological analyses in the Puget Lowland indicate a wide range of paleoenvironments during the Olympia interval. Many locations of Olympia beds yield excellent pollen preservation with a predominance of pine and spruce; freshwater diatomites suggesting clear, shallow lakes and large littoral areas; and macrofossils including mammoth teeth and tusks, *Pinus* sp.? cones and needles, branches, leaf prints, and in situ tree roots (Troost, 2002).

As expected with deposition during nonglacial periods, Olympia beds vary in thickness, elevation, grain size, and composition over short distances. Topographic relief on the basal unconformity of the Olympia beds near Tacoma exceeds 230 m, 60 m of which lies below modern sea level. The thickest exposures of Olympia beds (>25 m) include multiple

Table 1. Sources of age data for puget lowland stratigraphic units.

Name (Climatic Intervals in Italics)	Type Section Location	Reference for Nomenclature	Reported Age (in 10^3 Years)	Type of Date	Location	Reference for Age	Comment
Sumas glaciation	Near Sumas, Canadian side	Armstrong (1957)	na	na	na	na	na
Sumas Drift		Easterbrook (1963)	11.3–10.0 and pre 11.9	^{14}C yr B.P.	Aldergrove–Fort Langley and Chilliwack R. valley, B.C.	Clague *et al.* (1997)	New dates
			11.5–10.0	^{14}C yr B.P.	Multiple locations, Fraser Lowland and Nooksack valley	Kovanen & Easterbrook (2001), Kovanen (2002)	Compilation and new dates; table of 69 dates on the Sumas interval
Sumas stade		Armstrong *et al.* (1965)	na	na	na	na	na
Everson Glaciomarine Drift, *Everson interstade*	Upstream of Everson, on the Nooksack R.	Armstrong *et al.* (1965)	13.0–11.0	^{14}C yr B.P.	Type section	Armstrong *et al.* (1965)	New dates
			13.0–11.5	^{14}C yr B.P.	Whidbey Is. to Campbell R.	Kovanen & Easterbrook (2001)	Compilation and new dates
			13.6–11.3	^{14}C yr B.P.	Northern Puget Lowland	Dethier *et al.* (1995)	New dates
	Southeast of Cedarville on the Nooksack R.	Easterbrook (1963)	na	na	na	na	Includes the Kulshan glaciomarine drift, Deming Sand, and Bellingham glaciomarine drift
Vashon till, *Vashon glaciation*	Vashon Island	Willis (1898)	>13.5	na	na	Rigg & Gould (1957)	Youngest limiting age
Vashon Drift, *Vashon stade*		Armstrong *et al.* (1965)	25.0–13.5	^{14}C yr B.P.	Multiple, Strait of Georgia to Lake Washington	Armstrong *et al.* (1965)	New dates
			18.0–13.0	^{14}C yr B.P.	Fraser Lowland	Kovanen & Easterbrook (2001)	Compilation
			16.0–13.5	^{14}C yr B.P.	Seattle, Bellevue, Issaquah	Porter & Swanson (1998)	Compilation and new dates
Steilacoom Gravel	Steilacoom plains	Willis (1898), Bretz (1913), Walters & Kimmel (1968)	Younger than 13.5	^{14}C yr B.P.	Ft. Lewis, Tacoma	Borden & Troost (2001)	Multiple, young, sub–Vashon dates
Esperance Sand Member of Vashon Drift	Fort Lawton, Seattle	Mullineaux *et al.* (1965)	15.0–13.5; 15.0–14.5	^{14}C yr B.P.	Seattle; Issaquah	Mullineaux *et al.* (1965), Porter & Swanson (1998)	Limiting ages
Lawton Clay Member of Vashon Drift	Fort Lawton, Seattle	Mullineaux *et al.* (1965)	15.0–13.5; 15.0–14.5	^{14}C yr B.P.	Seattle; Issaquah	Mullineaux *et al.* (1965), Porter & Swanson (1998)	Limiting ages

Port Moody nonglacial deposits	Port Moody	Hicock *et al.* (1982)	23.0–21.0	^{14}C yr B.P.		Hicock & Armstrong (1981)	New dates
Port Moody interstade					na	Hicock & Armstrong (1985)	Interstade informally introduced
Coquitlam Drift	Coquitlam–Port Moody	Hicock (1976)	21.7–18.7	^{14}C yr B.P.	Type section	Hicock & Armstrong (1981)	New dates
Coquitlam stade		Hicock & Armstrong (1985)	30.0–25.0	^{14}C yr B.P.	Multiple locations	Hicock & Armstrong (1985)	Compilation of 52 dates
			26.0–17.8	^{14}C yr B.P.	Multiple locations	Clague (1980), Armstrong *et al.* (1985)	Equivalent to Evans Creek stade?
Evans Creek Drift, *Evans Creek stade*	Carbon River valley, near mouth of Evans Creek	Crandell (1963)	25.0–15.0	^{14}C yr B.P.	Type section	Armstrong *et al.* (1965)	Alpine glaciation in Cascade Range
		Armstrong *et al.* (1965) (Crandell)	na	na	na	na	na
Olympia interglaciation	Fort Lawton	Armstrong *et al.* (1965)	35.0–15.0	^{14}C yr B.P.	Fort Lawton and multiple locations in WA and BC	Armstrong *et al.* (1965), Troost (1999)	Compilation and new dates; may be partly equivalent to Quadra sediments at Point Grey in Vancouver (Armstrong & Brown, 1953)
			24.0–15.0	^{14}C yr B.P.	Fort Lawton and West Seattle	Mullineaux *et al.* (1965)	New dates
Olympia beds		Minard & Booth (1988)	>45–13.5	^{14}C yr B.P.	Multiple locations around Seattle and Tacoma	Troost (1999), Borden & Troost (2001)	New dates
Possession Drift	Possession Point, Whidbey Island	Easterbrook *et al.* (1967)	80	Amino acid	Multiple locations	Easterbrook & Rutter (1981)	New dates
Whidbey Formation	Double Bluff, Whidbey Island	Easterbrook *et al.* (1967)	107–96, avg = 100	Amino acid	Multiple locations	Easterbrook & Rutter (1981)	New dates
			151–102	Thermo–luminescence		Easterbrook (1994)	

Table 1. (Continued)

Name (Climatic Intervals in Italics)	Type Section Location	Reference for Nomenclature	Reported Age (in 10^3 Years)	Type of Date	Location	Reference for Age	Comment
Double Bluff Drift	Double Bluff, Whidbey Island	Easterbrook *et al.* (1967)	250–150	Amino acid	Type section	Easterbrook & Rutter (1982)	New dates
			178–111	Amino acid		Blunt *et al.* (1987)	na
			291–177	Thermo–luminescence		Easterbrook *et al.*, 1992	na
Salmon Springs Drift	Near Sumner	Crandell *et al.* (1958)	1000	Inferred, based on Lake Tapps	Type section	Easterbrook (1994)	Reversely magnetized (Easterbrook, 1986)
Lake Tapps Tephra	Near Sumner	Crandell (1963), Easterbrook & Briggs (1979)	840	Fission track	3 locations	Easterbrook & Briggs (1979)	Correlation of other locations to type section based on chemistry
			1000	Fission track	Multiple locations	Westgate *et al.* (1987)	na
Puyallup interglaciation, Puyallup Sand	Near Alderton	Willis (1898)	1690–1640	Laser–argon	Type section	Easterbrook *et al.* (1992)	New date; reversely magnetized (Easterbrook, 1986)
Puyallup Formation		Crandell *et al.* (1958)	na	na	na	na	na
Stuck Drift	Near Alderton	Crandell *et al.* (1958)	Close to 1600	Based on bounding ages	Type section	Easterbrook (1994)	Reversely magnetized (Easterbrook, 1986)
Alderton Formation	Near Alderton	Crandell *et al.* (1958)	2400–1000, avg = 1600	Laser–argon	Type section	Easterbrook (1994)	Reversely magnetized (Easterbrook, 1986)
Orting Gravel, Orting Drift	Orting	Willis (1898), Crandell *et al.* (1958)	2000 (?)	Inferred	Type section	Easterbrook (1986), Easterbrook *et al.* (1988)	Reversely magnetized (Easterbrook, 1986)

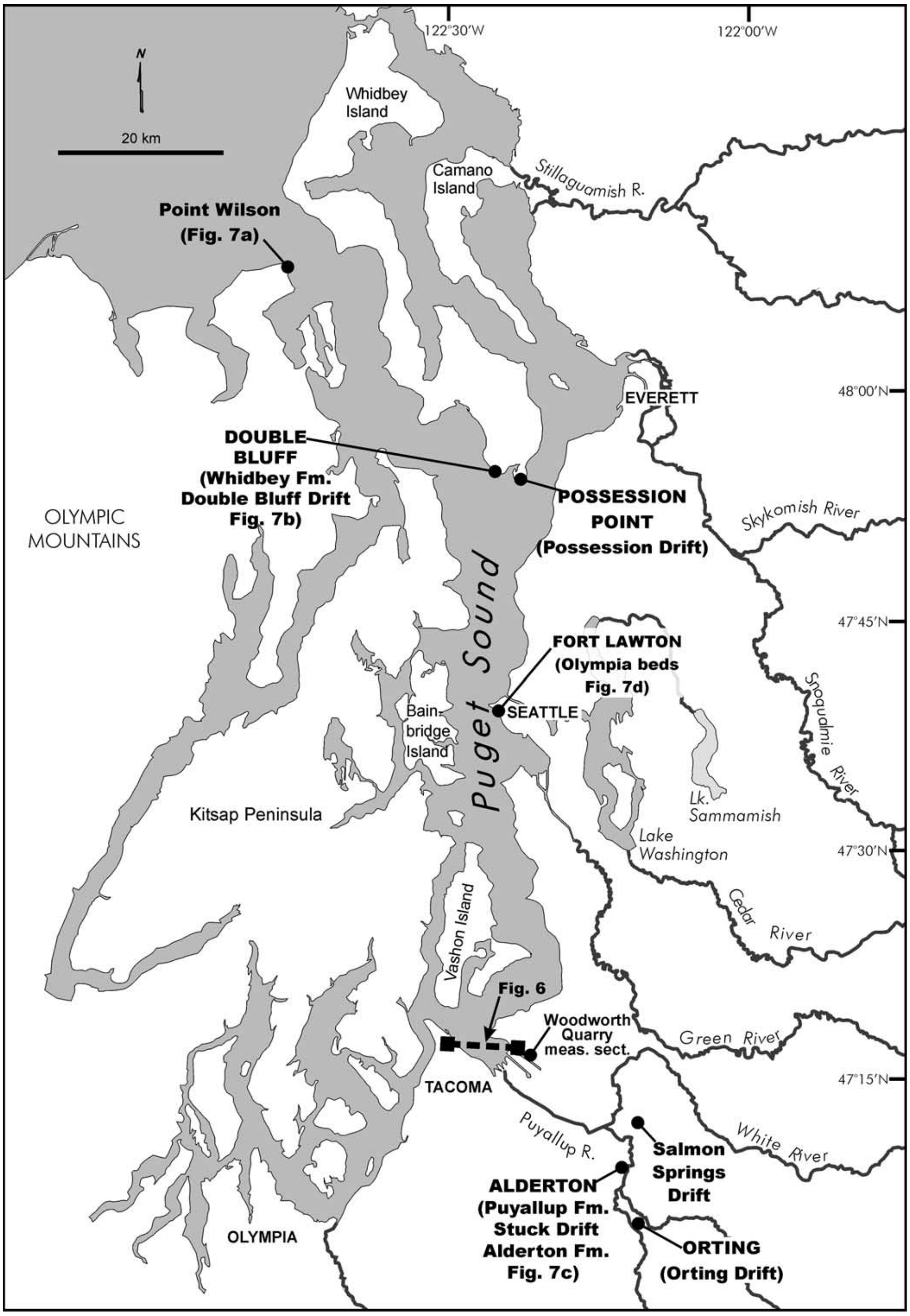


Fig. 5. Locations of type sections for the recognized pre-Fraser stratigraphic units in the Puget Lowland. Locations of cross section of Fig. 6 and measured sections in Fig. 7 are also shown. The Olympia nonglacial period was first defined by Armstrong et al. (1965) with its type section at Fort Lawton (Mullineaux et al., 1965). The Possession Drift, Whidbey Formation, and Double Bluff units were named and described by Easterbrook et al. (1967, 1981). The Salmon Springs and older drifts were first described by Willis (1898) and formally named by Crandell et al. (1958).

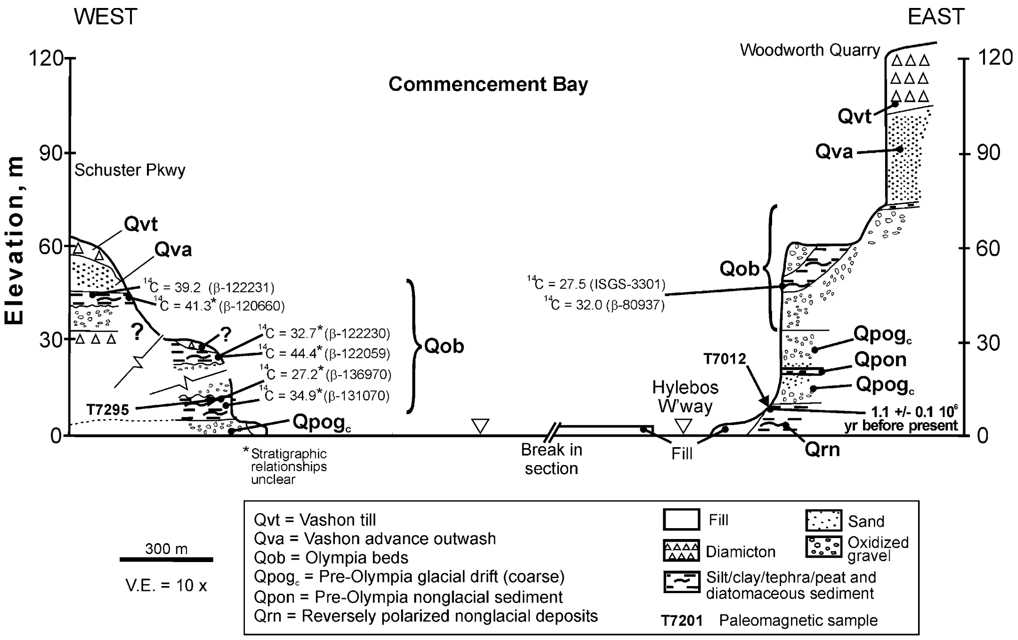


Fig. 6. East-west cross section through Commencement Bay near Tacoma, showing radiocarbon dates and topographic relief within the Olympia beds (unit Qob). Reversely magnetized nonglacial volcanic-rich deposits yield a zircon fission-track age of 1.1 × 10^6 years (modified from Troost et al., 2003). Unlabeled numbers are ^{14}C ages in 10^3 yr B.P.

tephra, lahar, peat, and diatomite layers (Troost *et al.*, 2003). At least five discontinuous Olympia-age tephras and lahars have been identified near Tacoma, with source areas including Mt. St. Helens and Mt. Rainier. Freshwater diatomites and in situ tree roots reveal lacustrine and forested environments across the lowland. Mastodons, mammoths, and bison roamed the Puget and Fraser lowlands during this nonglacial interval (Barton, 2002; Harrington *et al.*, 1996; Plouffe & Jette, 1997).

The next-oldest Pleistocene sediment in the Puget Lowland is the Possession Drift, probably related to glaciation during MIS 4 (Easterbrook, 1994) (Fig. 7a). The ice sheet responsible for this drift may have been less extensive than during MIS 2, according to reconstructions of global temperature. Away from the type section on Whidbey Island, pre-Fraser glacial deposits cannot be uniquely correlated with Possession Drift without age control. Thermoluminesence dating may prove most useful in this age range (Easterbrook, 1994), with preliminary results suggesting localities of Possession-age outwash south of the type section (Easterbrook, 1994; Mahan *et al.*, 2000).

The Whidbey Formation and its counterpart in British Columbia, the Muir Point Formation, correlate with MIS 5, the youngest full interglacial interval of the Pleistocene record. Climate was similar to that of today, with sea level perhaps slightly above today's level (Easterbrook, 1994; Easterbrook *et al.*, 1967). At its type section (Easterbrook, 1994) (Fig. 7b), the Whidbey Formation includes silt, sand, gravel, ash, and diatomite. On Whidbey Island, extensive sand deposits at the type section may be deltaic in origin. Like deposits of the Olympia nonglacial interval, sedimentary layers surely vary in thickness and composition over short distances; relief on the upper surface of the Whidbey Formation probably resembles today's landscape relief. Difficulties in dating sediments of this age, however, provide few constraints on the paleotopography from this time.

Still older mid- and early-Pleistocene deposits in the Puget Lowland include the Double Bluff Drift (Easterbrook, 1994) (Fig. 7b) and various unnamed glacial and interglacial deposits in the interval from 250,000 to 780,000 years ago, the existence of which are anticipated from climatic fluctuations expressed by the marine isotope record. Recent chronological and stratigraphic correlation efforts have begun to identify deposits in this age range and to confirm the presence of pre-Fraser deposits at locations away from their

Fig. 7. Measured sections at pre-Fraser localities on and near Whidbey Island and in the Puyallup Rivervalley (reproduced from Easterbrook, 1994), and at Fort Lawton in Seattle. Fig. 7a depicts both the Possession Drift and the Whidbey Formation at Point Wilson. Fig. 7b shows the lithologies noted at the type locality of the Double Bluff Drift. Fig. 7c depicts the stratigraphic relationships between the Puyallup Formation, Stuck Drift, and Alderton Formation at the Alderton type locality; black dots depict reversely polarized samples. Fig. 7d shows the modern exposure at the type locality for deposits of the Olympia nonglacial interval, and for the Lawton Clay Member and Esperance Sand Member (the latter now generally mapped as Vashon advance outwash) of the Vashon Drift (Mullineaux et al., 1965).

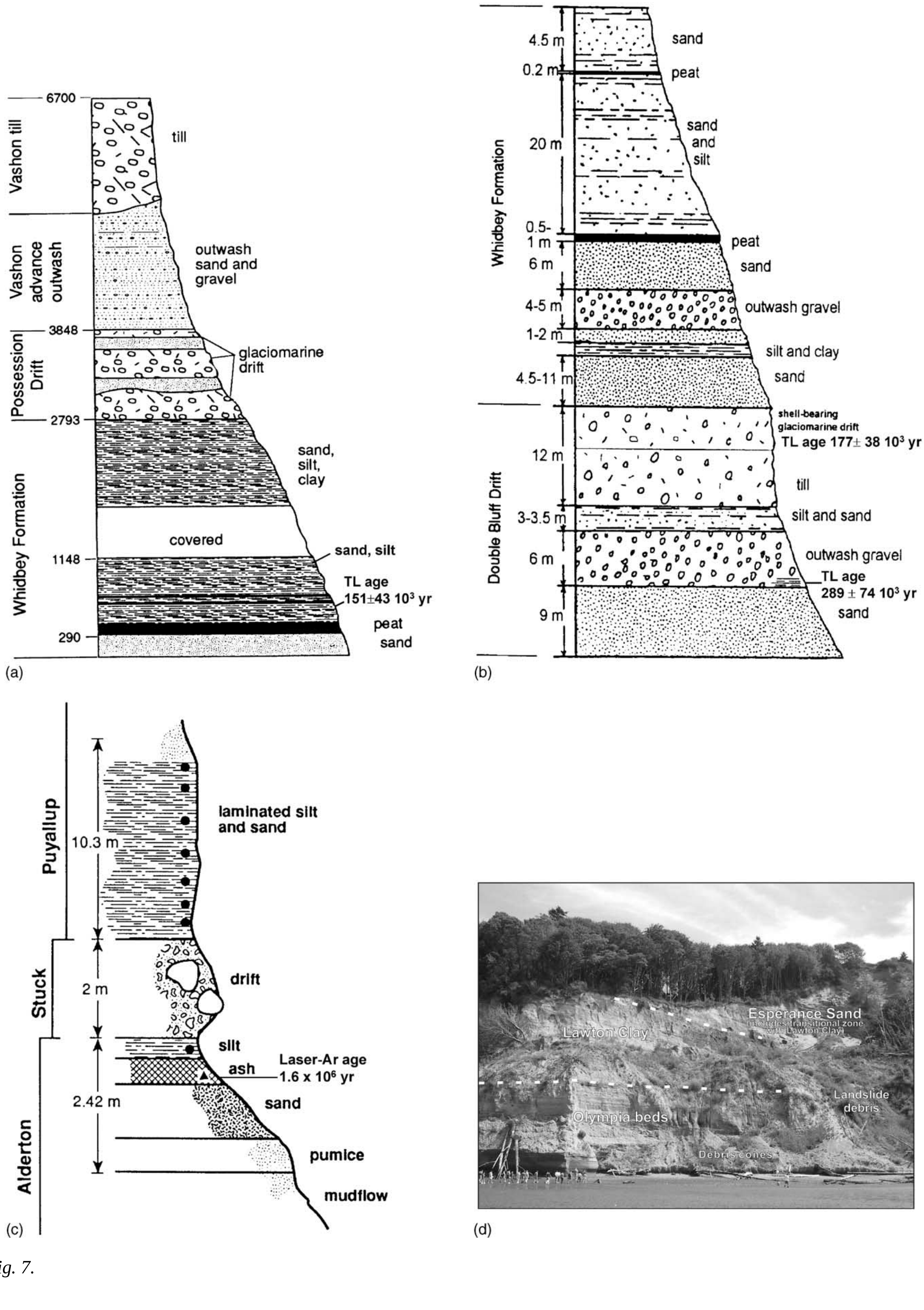

6700
Vashon till
till
Vashon advance outwash
outwash sand and gravel
3848
Possession Drift
glaciomarine drift
2793
sand, silt, clay
Whidbey Formation
covered
1148
sand, silt
TL age 151±43 10³ yr
peat
290
sand
(a)
4.5 m
sand
0.2 m
peat
20 m
sand and silt
Whidbey Formation
0.5-1 m
peat
6 m
sand
4-5 m
outwash gravel
1-2 m
silt and clay
4.5-11 m
sand
shell-bearing glaciomarine drift
TL age 177± 38 10³ yr
12 m
till
Double Bluff Drift
3-3.5 m
silt and sand
6 m
outwash gravel
TL age 289 ± 74 10³ yr
9 m
sand
(b)
Puyallup
10.3 m
laminated silt and sand
Stuck
2 m
drift
silt
Laser-Ar age 1.6 x 10⁶ yr
ash
2.42 m
sand
Alderton
pumice
mudflow
(c)
Esperance Sand
(includes transitional zone with Lawton Clay)
Lawton Clay
Landslide debris
Olympia beds
Debris cones
(d)

Fig. 7.

type sections (Hagstrum *et al.*, 2002; Mahan *et al.*, 2000; Troost *et al.*, 2003). The oldest pre-Fraser deposits, about 1 million years old and older, are the Salmon Springs Drift, Puyallup Formation, Stuck Drift, Alderton Formation, and Orting Drift (Crandell, 1963; Westgate *et al.*, 1987) (Fig. 7c).

Eastern Washington

Discontinuous drift extending beyond the limits of Fraser-age drift in the Pend Oreille, Columbia, and Little Spokane valleys has stones that are highly weathered or deeply penetrated by cracks, has a slightly argillic soil, and overlies granite and gneiss bedrock that is highly decayed, even to grus. These characteristics indicate that the drift is pre-Fraser in age. Direct dating of pre-Fraser sediments is poor, but radiocarbon dates in Canada have been interpreted as denying the existence of an ice sheet between 65,000 and 25,000 yr B.P. (Clague, 1980), consistent with nonglacial conditions west of Cascade Range during this time. The weathering of the drift and surrounding bedrock in places is so strong as to suggest an age very much older than late Wisconsin – equivalent to MIS stage 6 (160,000–130,000 years ago) or older. In northeastern Washington and adjacent Idaho, however, there is no objective basis for Richmond's (1986, Chart 1) assignment of any of these deposits to particular time intervals.

Probably there were several pre-Wisconsin Cordilleran ice-sheet glaciations in eastern Washington and farther east in Idaho and Montana. Glacial Lake Missoula and great floods from it are possible only when the Purcell Trench lobe advances far enough south (to 48°10′ N) to dam the Clark Fork of the Columbia. In southern Washington, deposits resembling Fraser-age Missoula-flood gravel bars but thickly capped by calcrete deeply underlie some of these Fraser deposits. One such gravel was dated to between 200,000 and 400,000 Th/U yr ago and another to before 780,000 Th/U yr ago (Bjornstad *et al.*, 2001).

Chronology of the Fraser Glaciation

The Cordilleran ice sheet most recently advanced out of the mountains of British Columbia about 25,000 ^{14}C yr B.P. It flowed west onto the continental shelf, east into the intermontaine valleys of British Columbia where it probably merged with the western edge of the Laurentide ice sheet, and south into the lowlands of Washington State (Fig. 8, Table 1). In southern British Columbia and western Washington the Puget lobe filled the Fraser Lowland and the Puget Lowland between the Olympic Mountains and Cascade Range. The Juan de Fuca lobe extended east along the Strait of Juan de Fuca to termine some 100 km west of Washington's present coast. Several ice lobes east of the Cascade Range expanded south down the Okanogan Valley and down other valleys farther east. The Fraser-age ice-sheet maximum on both sides of the Cascade Range was broadly synchronous (Waitt & Thorson, 1983). It approximately coincided with the maximum advance of some parts of the Laurentide ice sheet in central North America at about 14,000 ^{14}C yr B.P. but lagged several thousand years behind the culminating advance of the most of the Laurentide ice sheet (Lowell *et al.*, 1999; Mickelson *et al.*, 1983; Prest, 1969).

Northern Puget Lowland/Southern Fraser Lowlands

The Fraser glaciation began about 25,000 ^{14}C yr B.P. with an expansion of alpine glaciers in the Coast Mountains of British

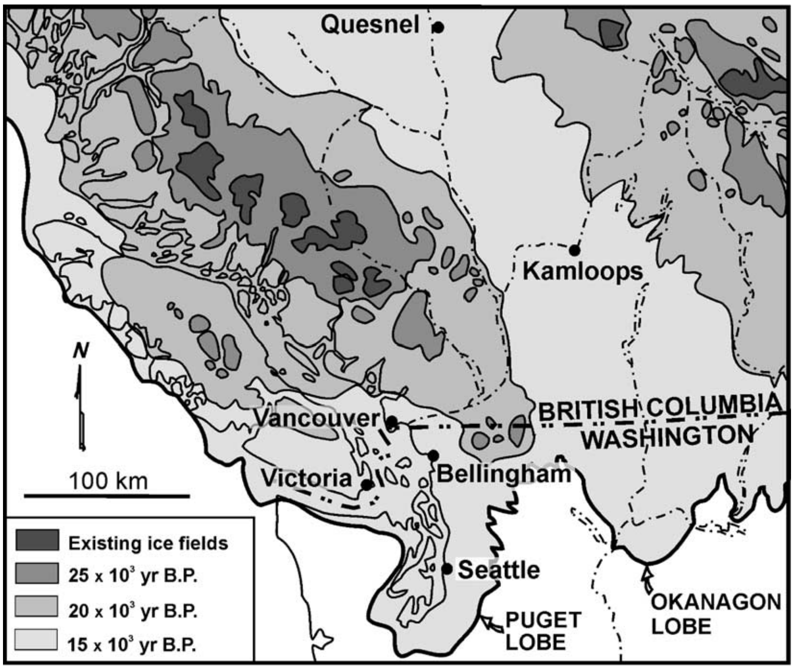


Fig. 8. Growth of the Cordilleran ice sheet during the Fraser Glaciation (from Clague, 1981).

Columbia, the Olympic Mountains, and the Cascade Range of Washington. Glaciers in the Coast Mountains coalesced to form piedmont ice lobes that reached the Fraser Lowland of British Columbia about 21,000 ^{14}C yr B.P. during the Coquitlam Stade (Hicock & Armstrong, 1981). The Coquitlam Stade correlates with the Evans Creek Stade of Washington, an early alpine phase of the Fraser Glaciation in the Cascade Range (Armstrong *et al.*, 1965).

The Coquitlam Stade was followed by a period of climatic amelioration that lasted from about 19,000 to 18,000 ^{14}C yr B.P. – the Port Moody Interstade of Hicock & Armstrong (1985). The Port Moody Interstade was in turn followed by the late Wisconsin advance of the Cordilleran ice sheet during the Vashon Stade (Armstrong *et al.*, 1965). The Puget lobe advanced into northern Washington about 17,000 yr B.P. (Clague, 1981; Easterbrook, 1986) and retreated rapidly from its maximum position around 14,000 yr B.P. (Clague, 1981; Easterbrook, 1986; Porter & Swanson, 1998).

The Vashon Stade was followed by a period of rapid and extensive glacier retreat (Everson Interstade) that ended with a resurgence of the southwestern margin of the Cordilleran ice sheet in the Fraser Lowland about 12,000 ^{14}C yr B.P. (Sumas Stade) (Clague *et al.*, 1997; Kovanen, 2002; Kovanen & Easterbrook, 2001). Several advances separated by brief periods of retreat apparently marked the Sumas Stade. The final advance(s) occurred 11,000 ^{14}C yr B.P. or shortly thereafter. Soon after 10,500 ^{14}C yr B.P., the Cordilleran ice sheet rapidly disappeared from the lowlands.

Central Puget Lowland

Rates of ice-sheet advance and retreat are well constrained in the central Puget Lowland. The Puget lobe advanced to the latitude of Seattle by about 14,500 ^{14}C yr B.P. (17,590 cal yr B.P.) and to its maximum by 14,000 ^{14}C yr B.P. (16,950 cal yr B.P.) (Porter & Swanson, 1998). The ice apparently remained near its maximum position only a few hundred years and then rapidly retreated. It retreated past Seattle by 13,600 ^{14}C yr B.P. (16,575 cal yr B.P.) (Porter & Swanson, 1998) (Fig. 9). Glacial lakes, including Lake Russell, formed south of the retreating ice front, draining through a spillway to the Chehalis River (Bretz, 1913). The lakes coalesced into one lake, Lake Bretz (Lake Leland of Thorson, 1980), which enlarged northward as the ice front retreated until a northern spillway was uncovered. Further backwasting allowed sea water to enter the lowland from the Strait of Juan de Fuca. Glaciomarine drift and other marine deposits accumulated in the northern lowland where land had not yet rebounded from isostatic depression. This interstade – named the Everson by Armstrong *et al.* (1965) – ended about 12,000 ^{14}C yr B.P. Isostatic rebound raised the glaciomarine and marine deposits above sea level between about 13,500 and 11,300 ^{14}C yr B.P. (Dethier *et al.*, 1995).

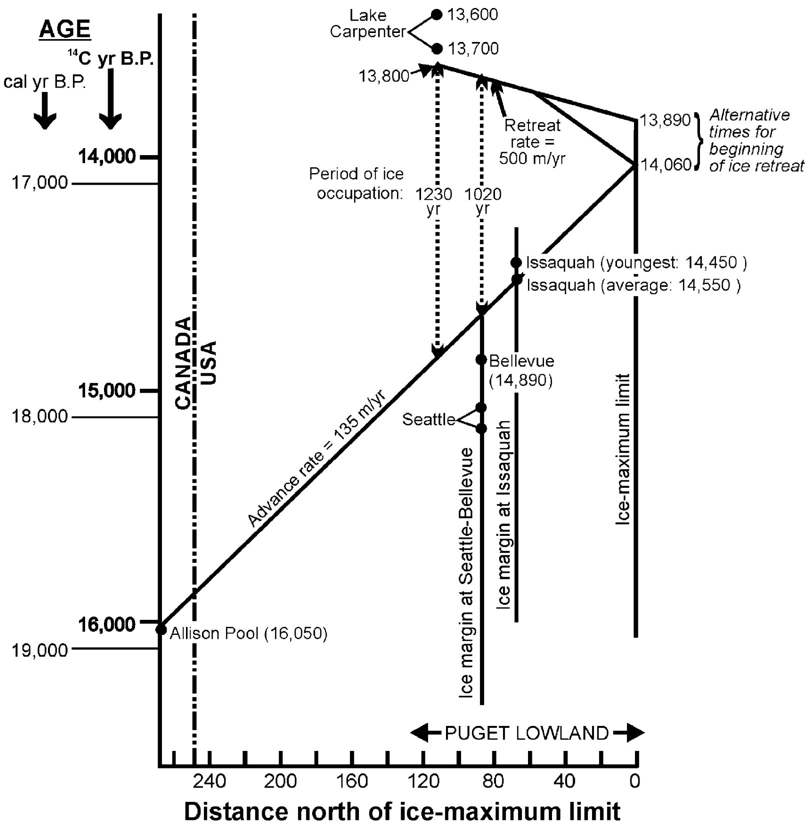


Fig. 9. Rates of Puget lobe advance and retreat in the Puget Lowland during the Vashon Stade (modified from Porter & Swanson, 1998). Rapid advance and retreat are required to honor the limiting radiocarbon dates from Lake Carpenter, Seattle, Bellevue, and Issaquah. Maximum ice-sheet extent could have persisted at most a few hundred years.

Eastern Washington

In contrast to the tight age constraints west of the Cascade Range, limits on the Fraser maximum east of the Cascades and Coast Mountains are broad. They include a date of 17,240 ^{14}C yr B.P. for proglacial advance outwash, 100 km north of the ice limit, followed by advance to the glacier maximum, then a retreat of at least 80 km by 11,250 ^{14}C yr B.P., judged partly on the distribution of Glacier Peak tephra layer G (Clague *et al.*, 1980; Mehringer *et al.*, 1984; Porter, 1978). Lake Missoula flood deposits, interbedded with varves of glacial Lake Columbia that contain detrital wood dated 14,490 ^{14}C yr B.P., suggest that the Purcell Trench lobe blocked the Clark Fork for 2,000–3,000 yr and reached its maximum extent about 15,000 ^{14}C yr B.P. (Atwater, 1986).

Sea-Level Record

Changing sea levels greatly altered the shorelines of the Pacific Northwest. Variations in relative sea level, ranging from 200 m above present sea level to more than 100 m below, are the integrated result of eustasy, isostasy, and tectonism. These phenomena are difficult to assess separately, however, because eustasy and isostasy are interdependent and because the eustatic component has proven particularly difficult to quantify.

Eustasy

Global Record

Eustatic sea-level changes are global and are caused mainly by changes in volume of ocean water. Fluctuating continental glaciers are the most important cause of eustatic sea-level change on the time scale of concern here – sea level falls when ice sheets grow and rises when they shrink. Seawater also decreases in volume as it cools, which further lowers sea level during glaciations.

The growth and decay of large ice sheets during the Pleistocene caused sea level to fluctuate by 120–140 m (Fairbanks, 1989; Lambeck *et al.*, 2000, 2002; Peltier, 2002; Yokoyama *et al.*, 2000). Estimates of sea-level lowering during the last glaciation (MIS 2) derive from fossil corals in Barbados, New Guinea, and Tahiti (Bard *et al.*, 1990a, b, 1993, 1996; Chappell & Polach, 1991; Fairbanks, 1989) and from more-recent sediment cores taken from the Sunda Shelf (Hanebuth *et al.*, 2000) and Northwest Shelf of Australia (Yokoyama *et al.*, 2000). Eustatic sea-level changes have also been estimated from variations in the oxygen-isotope composition of air in bubbles trapped in the Greenland and Antarctica ice sheets (Dansgaard *et al.*, 1971; Epstein *et al.*, 1970; Grootes *et al.*, 1993; Johnsen *et al.*, 1972; Jouzel *et al.*, 2002; Lorius *et al.*, 1985; Petit *et al.*, 1999) and in foraminifera in ocean sediment (Chapman & Shackleton, 1999; Chappell & Shackleton, 1986; Lea *et al.*, 2002; Shackleton, 1987; Waelbroeck *et al.*, 2002). Numeric modeling and geologic data (summaries in Clark & Mix, 2002) provide equivalent sea-level lowering of 118–130 m for the volume of ice locked in glaciers at the last glacial maximum.

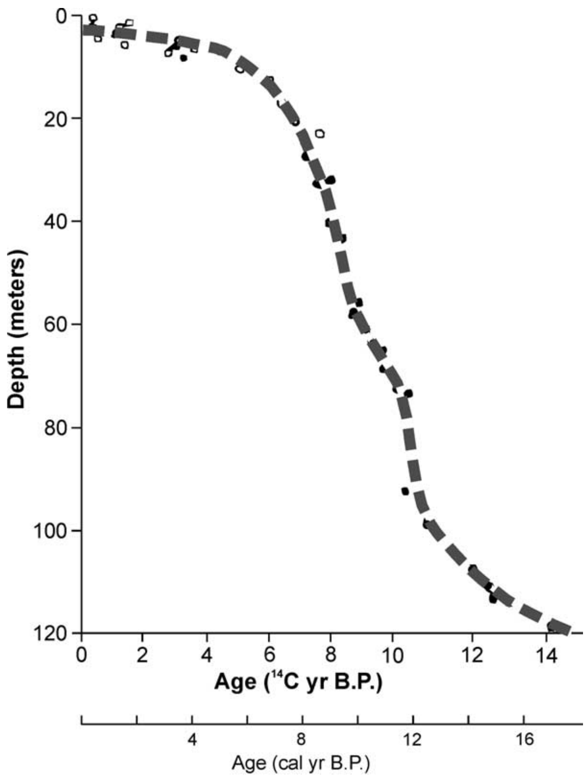


Fig. 10. Eustatic sea-level curve based on dating of shallow-water corals at Barbados (after Fairbanks, 1989).

Eustatic sea level rose after about 18,000 ^{14}C yr B.P. as ice sheets in the Northern Hemisphere began to decay. Sea-level rise accelerated after about 15,000 ^{14}C yr B.P. and remained high until about 7,000 ^{14}C yr B.P. when the Laurentide ice sheet had largely disappeared (Fig. 10; Fairbanks, 1989). Rates of eustatic sea-level rise were exceptionally high between about 11,000 and 10,500 ^{14}C yr B.P. and between 9,000 and 8,000 ^{14}C yr B.P. After 7,000 ^{14}C yr B.P., the rate of eustatic sea-level rise sharply decreased, and by 4,000 ^{14}C yr B.P. sea level was within 5 m of the present datum.

Regional Expression

It is difficult to disentangle the eustatic and glacio-isostatic components of the sea-level record in Washington and British Columbia. Isostatic depression and rebound dominate the late Pleistocene sea-level record in peripheral areas of the former Cordilleran ice sheet, but these effects decrease with distance beyond the ice margin. Estuaries in southwestern Washington record a mostly eustatic response, with the river valleys in this area drowned by rising sea level when ice sheets melted. In southwestern British Columbia and the northern Puget Lowland, in contrast, relative sea level is lower at present than during deglaciation because these areas were isostatically depressed during the last glaciation.

Isostasy

Global Record

The growth and decay of ice sheets, and thus changes in global sea level, redistributed mass on the Earth's surface. Ice sheets depressed the crust beneath them, but just beyond their margins the crust warped as a "forebulge" (Walcott, 1970). Melting ice sheets reversed the process: the forebulge migrated back towards the former center of loading to cause uplift there.

Water transfer from oceans to ice sheets unloaded the seafloor, and vice-versa during deglaciation. These hydro-isostatic adjustments opposed the direction of glacio-isostatic adjustments. Continental shelves rose when seawater was removed and they subsided again when melting ice sheets returned water to the oceans.

Regional Expression

Expanding glaciers during the early part of the Fraser glaciation progressively depressed the land surface of southwestern British Columbia and northwestern Washington (Clague, 1983). This depression started beneath the Coast Mountains, where glaciers first grew. As glaciers continued to advance, the area of crustal subsidence expanded beneath coastal areas. Subsidence probably exceeded the eustatic fall in sea level as ice sheets grew between 25,000 and 15,000 ^{14}C yr B.P. (Chappell *et al.*, 1996; Lambeck *et al.*, 2002; Shackleton, 1987; Waelbroeck *et al.*, 2002). If so, relative sea level in the region rose during this period. The relative rise in sea level controlled deposition of thick bodies of advance outwash (the Quadra Sand in British Columbia and the Esperance Sand in western Washington) on braided floodplains and deltas, and in littoral environments (Clague, 1976). As the Puget lobe reached its limit near the city of Olympia, the region to the north was isostatically depressed. The depression was greatest beneath the Strait of Georgia and Fraser Lowland and decreased south along the Puget Lowland.

The height of the uppermost shorelines that formed during deglaciation gives some limits on isostatic depression. Marine deltas near Vancouver lie 200 m above present sea level (Clague *et al.*, 1982). With eustatic sea level −100 m at the time the highest shorelines formed (Fairbanks, 1989), local glacio-isostatic depression must have exceeded 300 m. The depression was actually larger, because the Cordilleran ice sheet had thinned before the highest shorelines formed, and thus rebound had started earlier.

The modern altitudes of the late-glacial marine limit display the variable isostatic influence of the Cordilleran ice sheet. The marine limit is highest around the Strait of Georgia and in the Canadian part of the Fraser lowland and it declines west and south (Clague *et al.*, 1982; Dethier *et al.*, 1995; Mathews *et al.*, 1970). From about 125 to 150 m above sea level (asl) near Bellingham, it drops to 70 m asl west of Victoria, below 50 m asl on the west coast of Vancouver Island at Tofino, and probably below 50 m asl near the entrance to Juan de Fuca Strait. The marine limit decreases south of Bellingham to about 35 m asl at Everett. At the heads of the British Columbia mainland fiords to the north, the marine limit is fairly low because these areas remained ice-covered until isostatic rebound was well along (Clague & James, 2002; Friele & Clague, 2002).

Isostatic uplift rates can be inferred from a variety of shoreline data. Proglacial lakes covered southern and central Puget Lowland during deglaciation (Fig. 11), which were

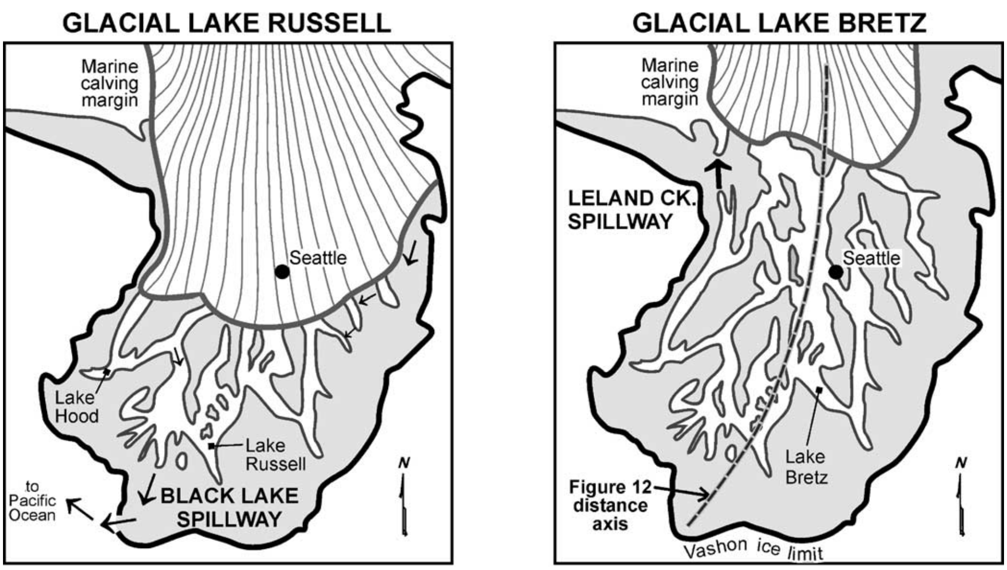


Fig. 11. Paleogeographic maps showing the maximum extents of Lake Russell and Lake Bretz (modified from Thorson, 1989, Fig. 2). Arrows show spillway locations controlling local and regional lake altitudes.

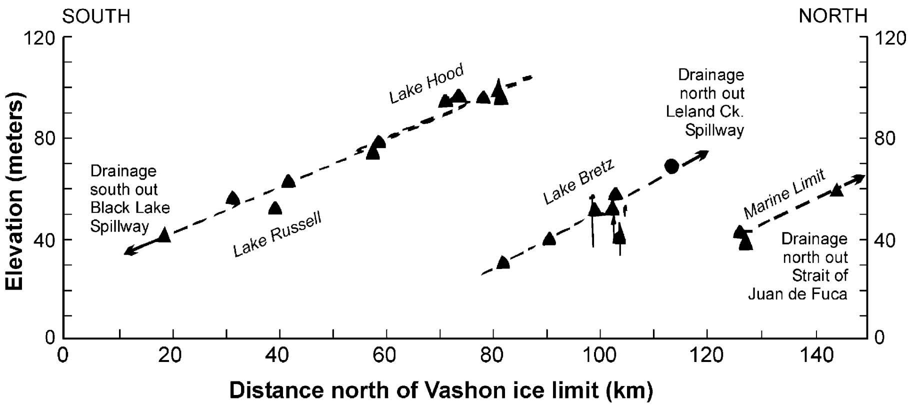


Fig. 12. North-south profile of shoreline features (delta tops) associated with Lake Russell, Lake Hood (confluent with Lake Russell), Lake Bretz, and the marine limit near Discovery Bay on the south side of the Strait of Juan de Fuca (modified from Thorson, 1989, Fig. 5; with additional data from Dethier et al., 1995).

dammed to the north by the retreating Puget lobe. The last lake drained when the Puget lobe retreated north of Port Townsend and marine waters entered Puget Sound. Differential isostatic rebound warped the shorelines of these proglacial lakes (Fig. 12) – shorelines of Lake Russell-Hood are tilted up to the north at 0.85 m/km; the tilt of Lake Bretz shorelines is 1.15 m/km (Thorson, 1989). Most uplift in the Fraser Lowland and on eastern Vancouver Island occurred in less than 1,000 years (Clague *et al.*, 1982; Mathews *et al.*, 1970), as inferred from the shoreline tilt data and relative sea-level observations. These data underlie a postglacial rebound model of the Cordilleran ice sheet (Clague & James, 2002; James *et al.*, 2000) that predicts low mantle viscosities ($<10^{20}$ Pa s).

Besides rapid rebound, low mantle viscosities in this region are responsible for nearly complete glacio-isostatic uplift by the early Holocene (Clague, 1983). Relative sea level was lower 8,000–9,000 ^{14}C yr B.P. than it is today, by at least 15 m at Vancouver (Mathews *et al.*, 1970) and by perhaps as much as 50 m in Juan de Fuca Strait (Hewitt & Mosher, 2001; Linden & Schurer, 1988). Evidence for lower sea levels includes submerged spits, deltas, and wave-truncated surfaces on the floor of Juan de Fuca Strait, and buried terrestrial peats found well below sea level in the Fraser Lowland. Sea level seems to have tracked global eustatic sea-level rise thereafter (Clague *et al.*, 1982; Mathews *et al.*, 1970), except on the west coast of Vancouver Island where sea level was several meters higher in the middle Holocene than now (Clague *et al.*, 1982; Friele & Hutchinson, 1993). Tectonic uplift probably caused this anomaly (see below).

Isostatic uplift occurred at different times in southwestern British Columbia and northwestern Washington as the Cordilleran ice sheet retreated. Regions that deglaciated first rebounded earlier than those deglaciated later (Fig. 13). Glacio-isostatic response to deglaciation varied across the region, showing that the lithosphere responded non-uniformly as the ice sheet decayed (Clague, 1983).

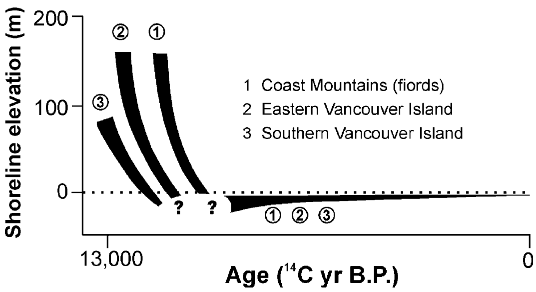


Fig. 13. Generalized patterns of sea-level change on south coast of British Columbia since the end of the last glaciation (Clague & James, 2002, Fig. 8; modified from Muhs et al., 1987, Fig. 10). Deglaciation and isostatic rebound occurred later in the southern Coast Mountains than on Vancouver Island. Line widths display range of uncertainty.

Tectonics

Trends in elevations of the late-glacial marine limit and the patterns of sea-level change summarized above show that much of the crustal deformation is isostatic. Slippage on reactivated faults, however, may have caused some of the observed deformation, analogous to recognized movement on some faults in the Puget Lowland later in the Holocene (Bucknam *et al.*, 1992; Johnson *et al.*, 1996, 2001). As yet, no such late-glacial or early postglacial fault movements have been documented unequivocally.

Late Quaternary sea-level change in the coastal Pacific Northwest also includes a component of aseismic tectonic deformation, but the rates of such vertical motions are at least an order-of-magnitude less than those of late-glacial eustatic and glacio-isostatic sea-level change and so cannot be isolated from those signals. However, the much slower changes in late Holocene sea level may include a significant component of aseismic tectonic deformation, which may partly explain the late Holocene regression on the west coast of Vancouver Island (Clague *et al.*, 1982; Friele & Hutchinson, 1993).

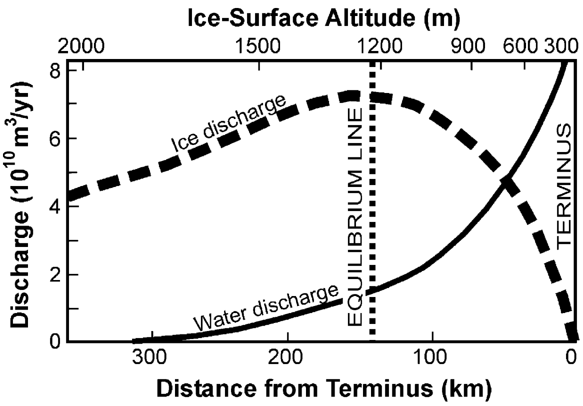


Fig. 14. Pattern of ice and water fluxes along the Puget lobe, reconstructed at ice-maximum conditions (from Booth, 1986).

Physical Behavior of the Cordilleran Ice Sheet

The Puget lobe of the Cordilleran ice sheet during the last glaciation provides an exceptional opportunity to examine the connection between glacier physics and the geomorphic products of the glacier system. Such an approach to interpreting the deposits and landforms of glaciated terrain has been widely applied only in the last several decades. The Puget lobe is not necessarily typical of every continental ice lobe, having a strong maritime influence. However, it is particularly well-constrained, with good age control, clearly recognized boundaries, moderately definitive source area, and good expression of its topographic effects and sedimentary deposits.

Ice-Sheet Reconstruction

By applying a height-mass balance curve (Porter *et al.*, 1983) over the reconstructed boundaries and surface altitudes of the Puget lobe, an ELA between 1,200 and 1,250 m balances the ice sheet (Booth, 1986). The ice flux peaks at the ELA, while meltwater flow increases monotonically downglacier (Fig. 14). The contribution to ice velocity from internal ice deformation (Paterson, 1981), based on reconstructed ice thickness and surface slope, is less than 2% of the total flux (Booth, 1986). Thus, basal sliding must account for virtually all of the predicted motion, several hundred meters per year over nearly the entire area of the lobe. From lobe dimensions, the calculated basal shear stress of the ice ranges between 40 and 50 kPa (Booth, 1986; Brown *et al.*, 1987). This value is low by the standards of modern valley glaciers but typical of ice streams and large modern ice lobes (Blankenship *et al.*, 1987; Mathews, 1974; Paterson, 1981), whose sliding velocities are also hundreds of meters per year. The system was thus one of rapid mass transport under a rather low driving stress across a bed of mainly unconsolidated sediment.

Average pore-water pressures across the glacier bed closely approached the ice overburden, because so much water cannot be quickly discharged (Booth, 1991a), even through an extensive subglacial tunnel system. Thus, the ice loading of bed sediments was low except near the margins, and the strength of the sediments correspondingly poor; shearing and streamlining would have been widespread. The modern landscape amply testifies to these processes (Fig. 15).

Meltwater

The Puget Lowland basin became a closed depression once the ice advanced south past the entrance of the Strait of Juan de Fuca and blocked the only sea-level drainage route. Lacustrine sediment (e.g. Lawton Clay Member of the Vashon Drift; Mullineaux *et al.*, 1965) accumulated in ice-dammed lakes, followed by fluvial outwash (Esperance Sand Member of the Vashon Drift) that spread across nearly all of the Puget

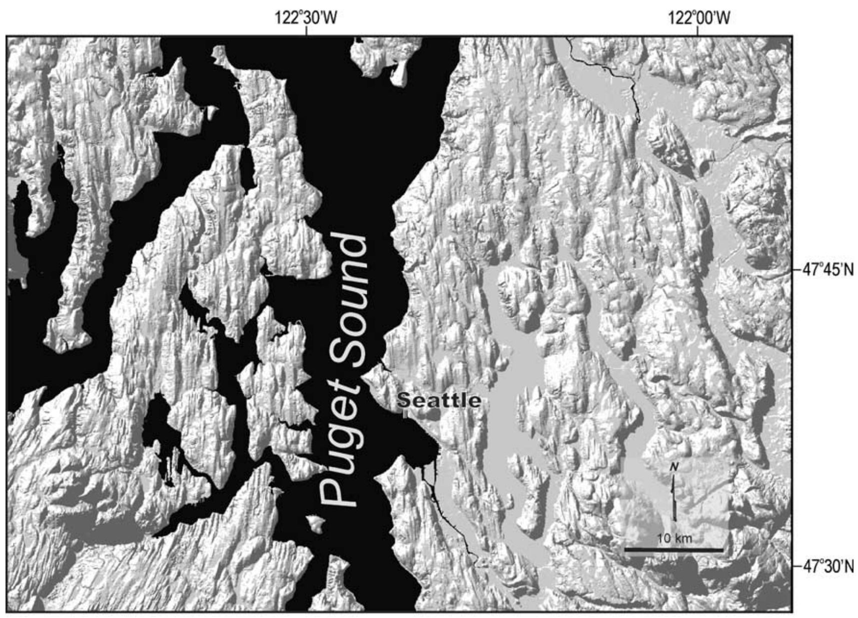


Fig. 15. Shaded topographic view of the central Puget Lowland, showing strongly streamlined landforms from the passage of the Puget lobe ice sheet during the Vashon Stade. Modern marine waters of Puget Sound in black; city of Seattle is in the south-central part of the view. Nearly all streamlined topography is underlain by deposits of the last glaciation.

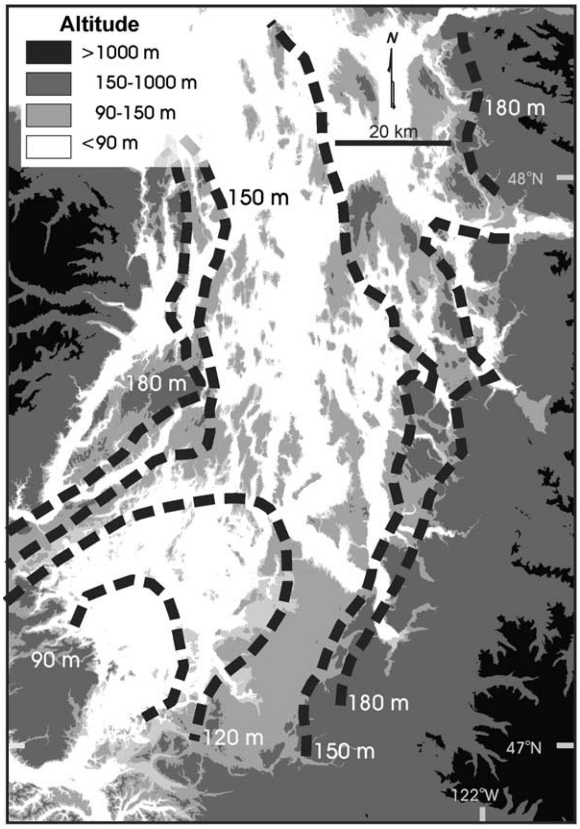


Fig. 16. Topographic shading of the Puget Lowland, from U.S. Geological Survey 10-m digital elevation model. Contours show generalized topography of the great Lowland fill (modified from Booth, 1994), with a modern altitude between 120 and 150 m across most of the lowland, as reconstructed from the altitude of modern drumlin tops, and subsequently incised by both subglacial channels and modern river valleys.

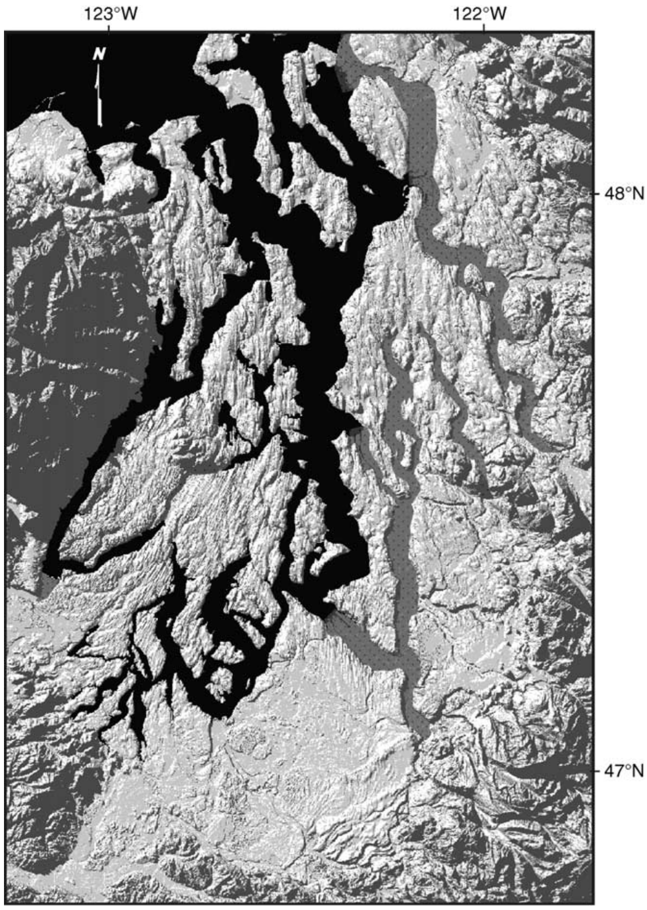


Fig. 17. Shaded topography of the Puget Lowland from U.S. Geological Survey 10-m digital elevation model, displaying the major subglacial drainage channels of the Puget lobe. Most are now filled by marine waters (black), with others by late-glacial and Holocene alluvium and mudflows (dark gray stipple).

Lowland. The outwash must have prograded as deltas like those that formed during ice recession (Thorson, 1980). With the greater time available during ice advance, however, sediment bodies coalesced into an extensive outwash plain in front of the ice sheet (e.g. Boothroyd & Ashley, 1975), named the "great Lowland fill" by Booth (1994) (Fig. 16). With continued ice-sheet advance and outwash deposition, this surface ultimately would have graded to the basin outlet in the southern Puget Lowland. Crandell *et al.* (1966) first suggested that this deposit might have been continuous across the modern arms of Puget Sound; Clague (1976) inferred a correlative deposit (Quadra Sand) filled the Georgia Depression farther north.

The fill's depositional history lasted 2,000–3,000 years. Outwash of the ice-sheet advance did not inundate the Seattle area until shortly before 15,000 ^{14}C yr B.P. (Mullineaux *et al.*, 1965). Deposition may have begun a few thousand years earlier, but accumulation would have been slow until advancing ice blocked drainage out of the Strait of Juan de Fuca (about 16,000 ^{14}C yr B.P.). Although late in starting, deposition across the entire lowland must have been complete before the ice maximum at about 14,000 ^{14}C yr B.P. (Porter & Swanson, 1998) because basal till of the overriding ice sheet caps the great Lowland fill almost everywhere.

Incised up to 400 m into the fill (and the overlying till) are prominent subparallel troughs (Fig. 17), today forming one of the world's great estuarine systems. These troughs were once thought to result from ice tongues occupying a preglacial drainage system (Willis, 1898), preserving or enhancing a topography of fluvial origin. This scenario is impossible, however, because impounded proglacial lakes would have floated the ice tongues and precluded any bed contact or ice erosion. Incision by subaerial channels is impossible because the lowest trough bottoms almost 300 m *below* the southern outlet of the Puget Lowland basin, and Holmes *et al.* (1988) report seismic-reflection data that suggest that the troughs were excavated during ice occupation to more than twice their current depth. Thus, troughs must have been excavated after deposition of the great lowland fill. Yet the troughs must predate subaerial exposure of the glacier bed during ice recession, because many of the eroded troughs

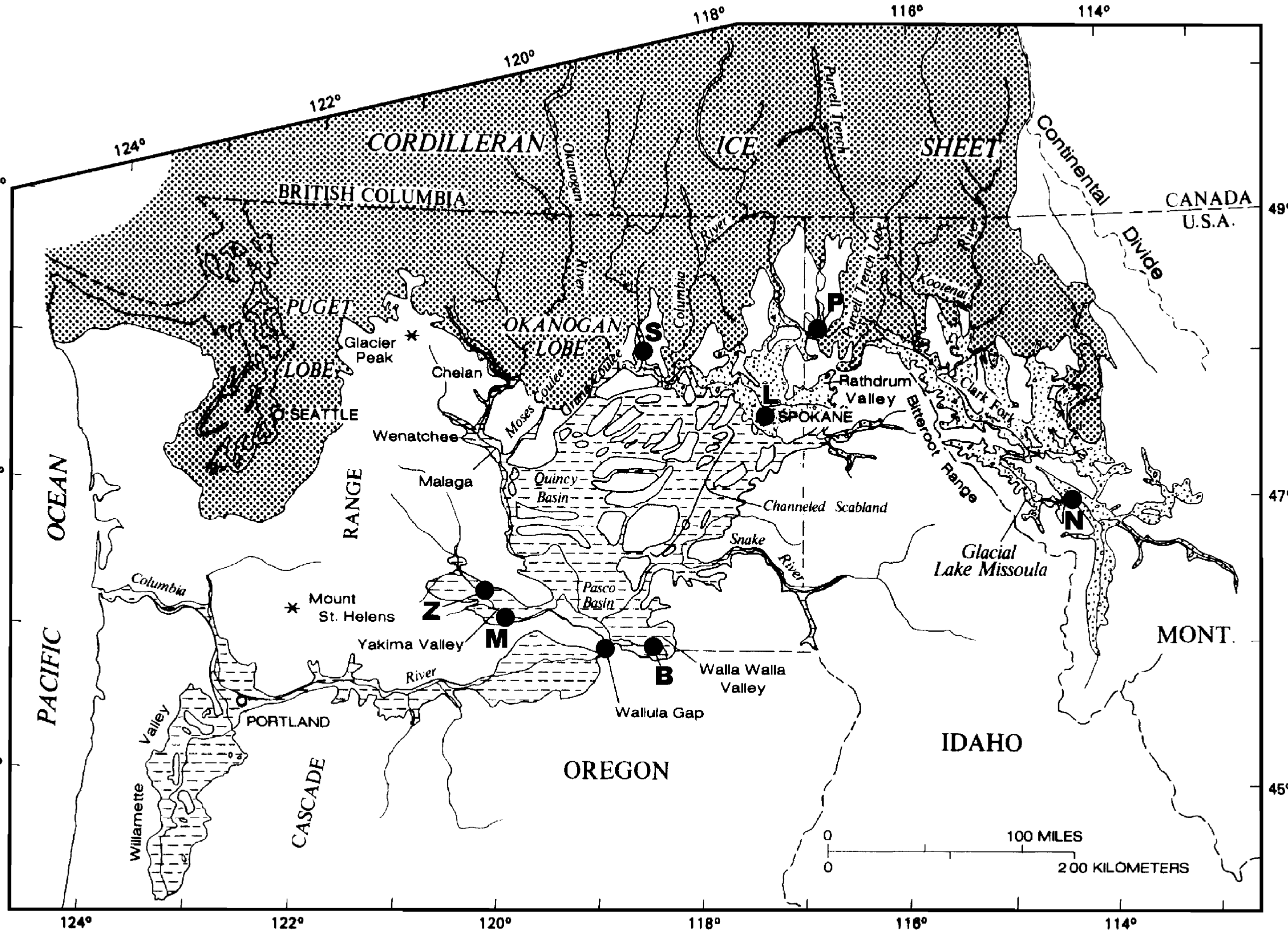


Fig. 18. Map of Columbia River valley and tributaries. Dark cross-hatching shows maximum extent of Cordilleran ice sheet; fine stipple pattern shows maximum area of glacial Lake Missoula east of Purcell Trench ice lobe and maximum extent of glacial Lake Columbia east of Okanogan lobe. Dashed-line pattern shows area that was swept by the Missoula floods in addition to these lakes. Large dots indicate key localities: B, Burlingame ravine; L, Latah Creek; M, Mabton; N, Ninemile Creek; P, Priest valley; S, Sanpoil valley; Z, Zillah. From Waitt (1985, Fig. 1). Relations at sites B, P, and N shown schematically on Fig. 21.

are still mantled on their flanks with basal till (e.g. Booth, 1991b) and filled with deposits of recessional-age lakes (Thorson, 1989). Thus, the troughs were formed primarily (or exclusively) by subglacial processes and probably throughout the period of ice occupation. They were carved primarily by subglacial meltwater (Booth & Hallet, 1993). A similar inference explains Pleistocene glacier-occupied troughs and tunnel valleys of similar dimensions and relief elsewhere in the Northern Hemisphere: Germany (Ehlers, 1981), Nova Scotia (Boyd *et al.*, 1988), New York (Mullins & Hinchey, 1989), Ontario (Shaw & Gilbert, 1990), and Minnesota (Patterson, 1994).

Missoula Floods

During several glaciations in the late Pleistocene, the Cordilleran ice sheet invaded Columbia River drainage and temporarily deranged it. The Purcell Trench lobe thwarted the Clark Fork of the Columbia to dam glacial Lake Missoula (Fig. 18) with volumes of as much as 2500 km^3 – as much water as Great Lakes Erie and Ontario together contain today. Stupendous floods from the lake swept the north and central part of the Columbia Plateau to carve a plexus of scabland channels as large as river valleys.

In the 1920s, J. Harlen Bretz argued an astonishing idea: the Channeled Scabland originated by enormous flood (Bretz, 1923, 1925, 1928a, b, 1929, 1932). His scablands evidence included gigantic water-carved channels, great dry cataracts (Fig. 19), overtopped drainage divides, and huge gravel bars. But with no known water source, skeptics in the 1930s–1940s tried to account for the scabland channels by mechanisms short of cataclysmic flood, such as by sequential small floods around many huge ice jams. Then Pardee (1942) revealed giant current dunes and other proof of a colossal outburst of glacial Lake Missoula. Thus, a source for Bretz's great flood had been found. In the 1950s, Bretz himself vindicated his old story (Bretz *et al.*, 1956). Baker (1973) showed that

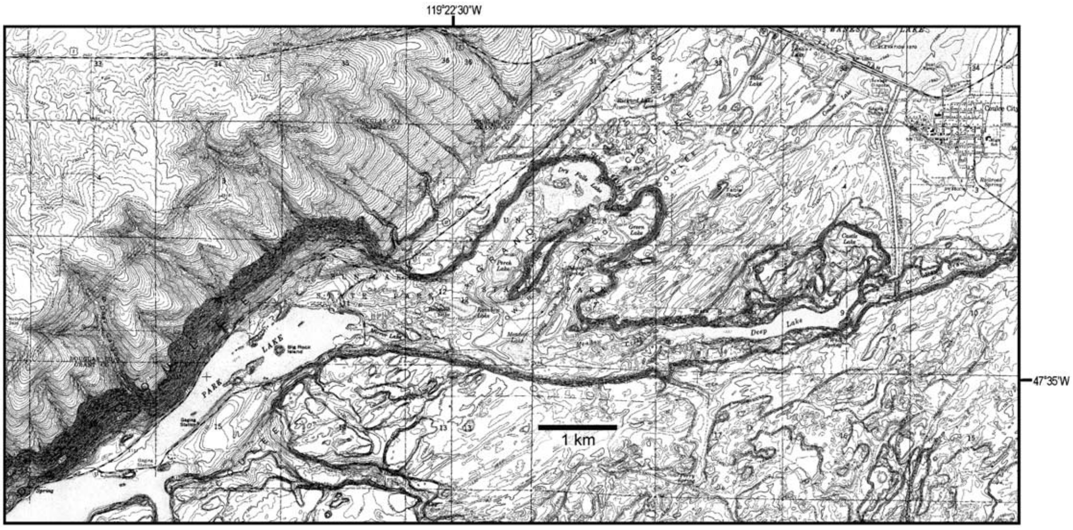


Fig. 19. Topographic map of Great Cataract Group, including Dry Falls in Grand Coulee (center of map). From U.S. Geological Survey 7.5-minute Park Lake and Coulee City quadrangles. Contour interval 10 ft. Land-grid squares (Township sections) are 1 mile (1.6 km) on a side. Top is north.

Bretz's observations were in accord with principles of open-channel hydraulics. Bretz's old heresy now wore respectable clothes.

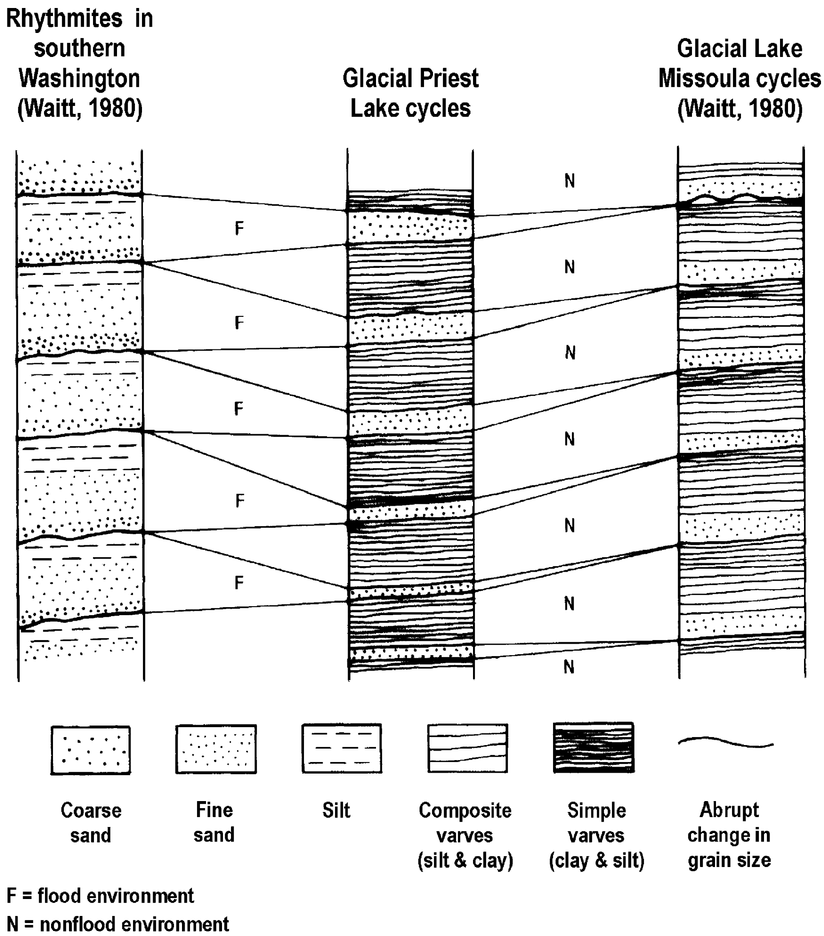
In the high-velocity, high-energy scabland reaches, one great flood eroded evidence of any earlier ones. But the waters also backflooded up tributary valleys and quietly deposited suspended load there in transitory hydraulic ponds. Within stacks of rhythmic beds in southern Washington (Fig. 20), the Mount St. Helens "set-S" ash couplet (14,000 ^{14}C yr B.P.) lies atop a floodlaid bed identical to other beds in these sections. This, and other evidence, shows that each graded bed is the deposit of a separate great flood. Numerous sites across the region tell a similar story of scores of separate floods (Atwater, 1984, 1986; Waitt, 1980, 1984, 1985, 1994). All together there were probably 95–100 Missoula floods during the last glaciation.

In northeastern Washington and Idaho, glacial lakes dammed along the Cordilleran ice margin (Fig. 18) accumulated sand-silt-clay varves. These beds are interrupted by many thick, coarse floodlaid beds. The numbers of varves indicate periods of 6 decades to a few years between successive floods (Atwater, 1984, 1986; Waitt, 1984, 1985). The only water body big and high enough to flood these glacial lakes was Lake Missoula. The sediment of Lake Missoula itself comprises dozens of fining-upward varve sequences, each the record of a gradually deepening then swiftly emptying lake (Chambers, 1971; Waitt, 1980). Fig. 21 relates the deposits across the region. The rhythmic beds of southern Washington record the floods, Lake Missoula bottom sediment records interflood periods, and the northern lake deposits record both.

East of the Cascade Range, the Fraser-age Cordilleran ice sheet is bracketed in time by preglacial dates as young as 17,200 ^{14}C yr B.P. and postglacial dates as old as 11,000 ^{14}C yr B.P. in southern British Columbia, 150–300 km north of the ice limit (Clague, 1981, 1989). Dammed at the ice terminus, Lake Missoula existed less than half this period. Fewer than 2,500 varves are known from Lake Missoula bottom sediment or between Missoula-flood beds in other glacial lakes (Atwater, 1986; Chambers, 1971).

Fig. 20. Rhythmically bedded Missoula-backflood deposits at Burlingame ravine, Walla Walla valley (site B of Fig. 18). Each graded bed is the deposit of a separate flood.

Fig. 21. Inferred relations between rhythmites in southern Washington, northern Washington and Idaho, and western Montana (Lake Missoula). These columns schematically represent sedimentary motif at sites B, P, and N of Fig. 18. From Waitt (1985, Fig. 17).



Radiocarbon ages and proxy ages further limit the age of the floods. Atwater (1986, Fig. 17) dated a wood fragment at 14,490 ^{14}C yr B.P. in the lower-middle of the Missoula-flood sequence in Sanpoil Valley. In Snake Valley, 21 Missoula-backflood couplets (Waitt, 1985) overlie gravel of the great flood from Lake Bonneville (ca. 14,500 ^{14}C yr B.P.; Oviatt *et al.*, 1992). The 14,000-yr-B.P. Mount St. Helens ash couplet overlies at least 28 giant-flood rhythmites in southern Washington and underlies 11 (Waitt, 1980, 1985). After these giant floods came several dozen smaller Missoula floods (Waitt, 1994). Organic matter within and below Missoula flood deposits in the Columbia gorge yielded three dates between 15,000 and 13,700 ^{14}C yr B.P. (O'Connor & Waitt, 1995). The 11,250 ^{14}C yr B.P. Glacier Peak tephra (Mehringer *et al.*, 1984) postdates ice-sheet retreat in northern Washington and Montana (Waitt & Thorson, 1983). These various limits suggest that glacial Lake Missoula existed for 2000 years or so during the period 15,700–13,500 ^{14}C yr B.P.

The controlling Purcell Trench ice dam became progressively thinner during deglaciation. Shallower lake levels were required to destabilize the smaller ice dam. Floods from the lake thus became smaller and more frequent. The average period between floods indicated by varves is about 30 years. At the glacial maximum it was much longer, and late during deglaciation it was much shorter. Atwater's varve counts (1986) detail a near-continuous record of the Missoula floods. The period between floods was about 50 years at the glacial maximum and during deglaciation decreased successively to 30, 20, and fewer than 10 years.

A recurring discharge every few decades or years suggests that glacial Lake Missoula emptied by a recurring hydraulic instability that causes glacier-outburst floods (jökulhlaups) from modern Icelandic glaciers (Waitt, 1980). As the water deepens against the ice dam, it buoys the lakeward end of the dam. Subglacial drainage occurs when the hydrostatic pressure of water from the lake exceeds the ice overburden pressure at the glacier bed (Bjornsson, 1974; Clarke *et al.*, 1984; Waitt, 1985). Drainage begins, and ice tunnels enlarge swiftly. The tunnel roof collapses; the whole lake drains. Glacier flow then repairs the damage, and within months the lake basin begins to refill.

The peak flow of Missoula floodwater down 10-km-wide Rathdrum Valley as modeled by O'Connor & Baker (1992, Figs 7 and 8) was at least 17 million m^3/s. More recent modeling suggests peak discharge almost twice that (Waitt *et al.*, 2000). During deglaciation the thinning ice dam fails at progressively shallower lake levels. Calculations suggest the Missoula floods ranged in peak discharge from as much as 30 million to as little as 200,000 m^3/s (Waitt, 1994). The largest were the Earth's grandest freshwater floods. Even a lake volume only one-third of maximum sufficed for a mighty flood down the Channeled Scabland and Columbia valley.

Summary

Advances in both global and regional understanding of Quaternary history, deposits, and geomorphic processes have brought new information and new techniques to characterize the growth, decay, and products of the Cordilleran ice sheet during the Pleistocene. Ice has advanced south into western Washington at least six times, but the marine-isotope record suggests that these are but a fraction of the total that entered the region in the last 2.5 million years. Several glacial and interglacial deposits are likely in the interval from 780,000 to 250,000 years ago but are not yet formally recognized. Growth and decay of large ice sheets during the Pleistocene have also caused sea level to fall and rise about 120–140 m, with strong influence on the tidewater margins of the Cordilleran ice sheet, as did progressive depression of the land surface as glaciers expanded during each glaciation. During the most recent (Fraser) glacier advance, local glacio-isostatic depression exceeded 270 m. Subsequent postglacial rebound of the Earth's crust, recorded in detail by proglacial lake shorelines, was rapid.

Reconstruction of the Puget lobe of the Cordilleran ice sheet during the last glacial maximum requires basal sliding at rates of several hundred meters per year, with pore-water pressures nearly that of the ice overburden. Landforms produced during glaciation include an extensive low-gradient outwash plain in front of the advancing ice sheet, a prominent system of subparallel troughs deeply incised into that plain and carved mainly by subglacial meltwater, and widespread streamlined landforms. At the southeastern limit of the ice sheet, the Purcell Trench lobe dammed glacial Lake Missoula to volumes as much as 2500 km^3, which episodically discharged as much as 30 million m^3/s. Scores of great floods swept across the Channeled Scablands of eastern Washington at intervals of typically a few decades, carving scabland channels as large as great river valleys. Modern geomorphic analysis of them confirms one of the region's early theories of wholesale development of landscape by the Cordilleran ice sheet.

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