Late Wisconsinan glaciomarine deposition and isostatic rebound, northern Puget Lowland, Washington

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ABSTRACT

The distribution and age of glaciomarine and marine sediment in the northern Puget Lowland, Washington, demonstrate that rapid retreat of continental ice, the Everson marine incursion, and high rates of isostatic rebound occurred between about 13 600 and 11 300 ^{14}C yr B.P. (11.3 ka). Glaciomarine and marine deposits are thickest in zones where retreating ice lobes grounded, in the northeast Puget Lowland, and near large drainages. Glaciomarine sediment was deposited mainly from (1) submarine outwash in ice-proximal zones; (2) turbid underflows, dispersed melt water, icebergs, and resedimentation in transitional zones; and (3) dispersed melt water and currents in ice-distal zones. Marine, estuarine, and emergence (intertidal and beach) facies accumulated in areas more than 10 km from ice margins, particularly near major rivers. Molluscan and foraminiferal assemblages in the glaciomarine and marine deposits indicate that turbid, cool, brackish water covered much of the Puget Lowland during the Everson interval. Water was generally shallower (<30 m) in the southern part of the area and deeper (15–60 m) to the north. Mineralogy and geochemical properties such as boron or sodium content of the gravel-free fraction do not clearly distinguish glaciomarine and marine deposits from terrestrial deposits.

Isostatic rebound rapidly lifted the glaciomarine and marine deposits through sea level between about 13.5 and 11.3 ka. The present altitudes of radiocarbon-dated shell and the marine limit show that initial rates of isostatic rebound exceeded 10 cm yr^{-1} in the northern Puget Lowland, but dropped to 2 cm yr^{-1} before 11 ka. The uplift gradient is about 0.6 m km^{-1} to the north and steepens locally to at least 1.3 m km^{-1}. The pattern of emergence in the northern Puget Lowland is anomalous locally, perhaps as a result of complex isostatic effects near the glacier margin, rapid rise of sea level, or tectonic deformation.

INTRODUCTION

This report describes the stratigraphic and geomorphic record of rapid, calving retreat of late Pleistocene Cordilleran ice, the incursion of marine waters, and glacioisostatic rebound in the northern Puget Lowland, Washington, between Everett and the International Boundary (Fig. 1). The Puget Lowland is a seismically active (Crosson, 1972; Yount and Crosson, 1983) topographic and structural trough bounded by the Olympic Mountains and Vancouver Island on the west, the Fraser Lowland to the north, and the Cascade Range on the east (Fig. 1). Hundreds of meters of Quaternary deposits (Hall and Othberg, 1974) blanket Tertiary bedrock in most of the Lowland, masking bedrock faults and structures identified using geophysical data (Rodgers, 1970; Wagner and Wiley, 1980, 1983; Gower et al., 1985; Johnson et al., 1994a). Detailed mapping, however, demonstrates that some late Pleistocene deposits in the central Puget Lowland are deformed near bedrock faults (Bucknam et al., 1992; Johnson et al., 1994b). Results we report here help to define stratigraphic and geomorphic relations of latest Pleistocene deposits and the record of glacioisostatic uplift in the northern Puget Lowland.

Landforms of the northern Puget Lowland are dominated by north-northwest–trending, streamlined hills and upland troughs, and by a number of major alluvial valleys. Elevations are generally <300 m, although several peaks in the San Juan Islands rise above 500 m adjacent to fjord-like troughs with average depths >125 m. In the eastern Strait of Juan de Fuca, deeper water is interrupted by a series of shallow banks and sills that extend as far west as Port Angeles (Chrzastowski, 1980). Submarine topography to the west of Port Angeles is smoother and slopes gradually toward the western margin of the Strait (Anderson, 1968).

We mapped and sampled glaciomarine and marine deposits of the Fraser (late Wisconsinan) glaciation in detail in the Port Townsend 1:100 000 quadrangle (Pessl et al., 1989) and studied adjacent areas as far west as Port Angeles and as far north as the San Juan Islands on a reconnaissance basis. Our study emphasized the distribution, stratigraphy, paleoecology, and geochemistry of glaciomarine and marine deposits, and geomorphic and ^{14}C evidence of ice recession, the marine incursion, and regional isostatic rebound. We used the altitude of strandlines, distribution of Everson glaciomarine and marine deposits (Easterbrook, 1976), and channels that terminate in Everson deposits (Thorson, 1980) to define the marine limit. Because some unfossiliferous glaciomarine deposits are difficult to distinguish from till, we performed mineralogic and geochemical analyses primarily to separate glaciomarine diamictons from till in the eastern Puget Lowland (Pevear et al., 1984).

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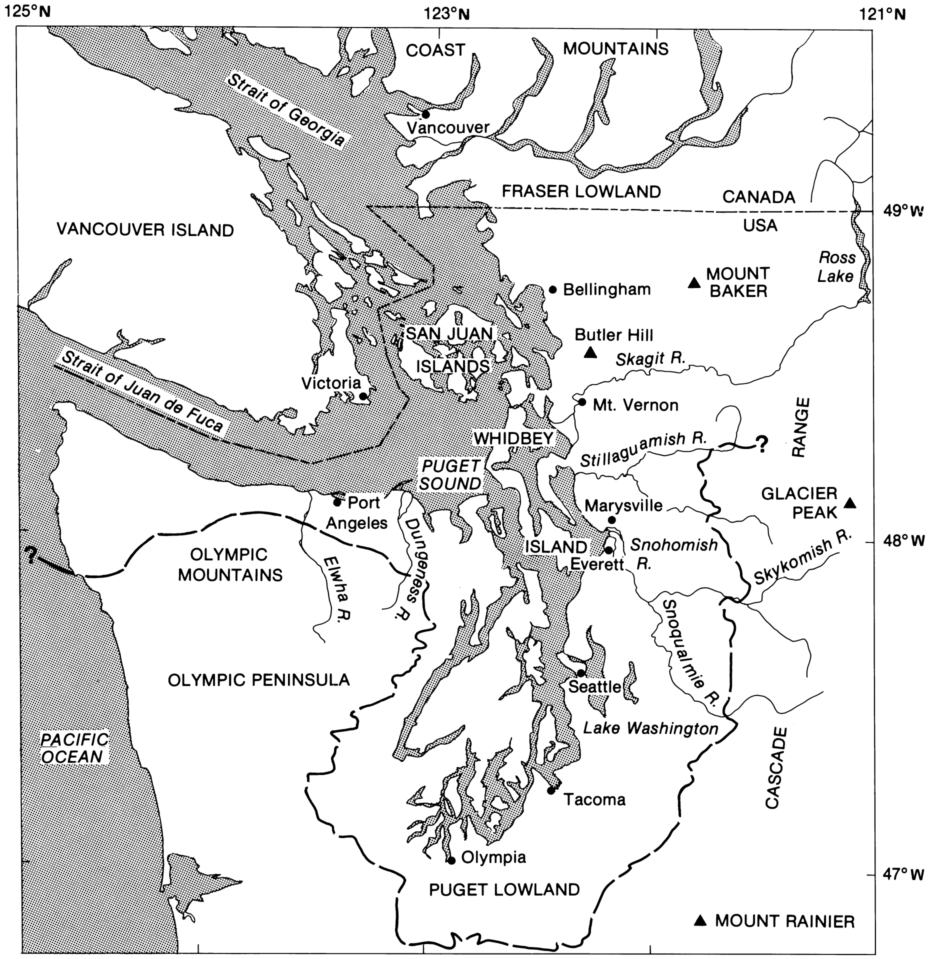


Figure 1. Sketch map of the Puget Lowland and adjacent areas in Washington and British Columbia. The northern Puget Lowland is considered to be that portion of the Puget Lowland north of the latitude of Everett, Washington (48°00′). Dashed line (from Waitt and Thorson, 1983) shows the approximate maximum extent of continental ice during the Fraser glaciation.

Our paleontological studies concentrated on the paleoecology of invertebrate assemblages (Balzarini, 1983) collected from published locations and from new sites discovered during regional mapping. We dated shell collected at sites where stratigraphic relations were particularly clear or where the change in relative sea level could be measured (Appendix Table A1). We report ^{14}C ages in years before 1950 (ka) without calibration to the sidereal scale (Stuiver and Reimer, 1993) or correction for marine reservoir effects.

Chronology of the Fraser Glaciation

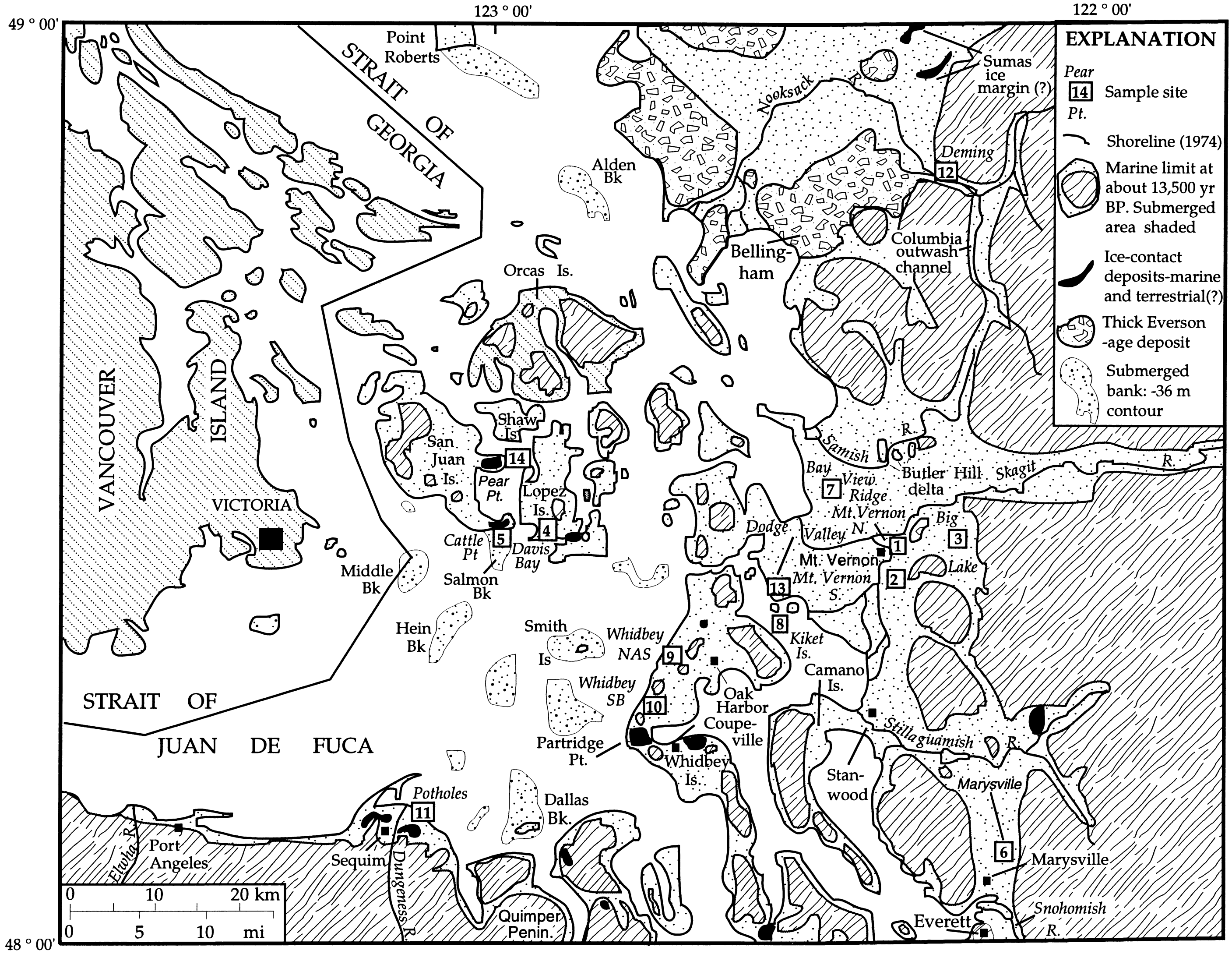
During the Fraser glaciation (25 to 10 ka), one lobe of Cordilleran ice (Juan de Fuca lobe) flowed west through the Strait of Juan de Fuca to a terminal position on the continental shelf. Another ice tongue, the Puget lobe, flowed through the Puget Lowland as far south as the low hills near Olympia. Continental ice attained an average thickness of about 1700 m at the International Boundary (Booth, 1987) and about 1200 m near Everett. Ice covered Vancouver, British Columbia, for about 5000 yr, whereas Everett was probably covered for <2000 yr (Waitt and Thorson, 1983).

During retreat of Fraser ice, rapid glacioisostatic rebound commenced (Thorson, 1989), and rising marine waters invaded the Puget Lowland via the Strait of Juan de Fuca (Fig. 1). Ice retreat, the marine incursion, and locally deep troughs created highly embayed, calving margins on the ice lobes (Thorson, 1980) in the northern Puget Lowland. Disintegration of the Fraser ice took <1000, and perhaps only a few hundred, years (Booth, 1987). The marine incursion began before 13.6 ka and ended with emergence before 11.0 ka. During this period, glaciomarine and marine sediment accumulated over >10 000 km^2 of northwestern Washington and adjacent British Columbia (Easterbrook, 1992) as far south as Seattle. The period was named the Everson interstade and was considered "... an essentially nonglacial episode ..." by Armstrong et al. (1965), who defined the subsequent Sumas stade on the basis of ice-contact deposits overlying Everson-age sediment near the International Boundary.

Prior Studies of the Everson Interval

The earliest investigators of glacial deposits in the Puget Lowland (Willis, 1898; Bretz, 1913) and Fraser Lowland (Clapp, 1912, 1913; Johnston, 1921, 1923) noted marine shells in till-like sediment associated with the glacial drifts. Armstrong and Brown (p. 358, 1954) termed these diamicts "marine drift" or "of glacio-marine origin." Shell-bearing drift that blankets low-lying areas in much of northwest Washington was mapped by Easterbrook (1963, 1966, 1969, 1979), who suggested that it was deposited from a floating ice shelf, by rain out of sediment from icebergs, or by submarine landslides. Armstrong (1981) subdivided the interbedded marine, glaciomarine, and terrestrial sediments of the Fraser Lowland into a lower member, the Fort Langley Formation, and an upper member, the Capilano Sediments. He suggested that icebergs and melt water were the principal sources of glaciomarine sediment in both members. Domack's (1982, 1983, 1984) mapping, granulometry, and fabric studies on central Whidbey Island provided the first detailed analysis of the range of Everson sedimentary environments in the Puget Lowland. Deposits of Everson age have also been mapped on Vancouver Island (Mathews et al., 1970; Hicock et al., 1981), in the Strait of Juan de Fuca (Anderson, 1968), on the northern Olympic Peninsula (Othberg and Palmer, 1979), and elsewhere in the Fraser Lowland (cf., Armstrong and Hicock, 1980).

We use "Everson interval" to identify the period during deglaciation when marine waters occupied the northern Puget Lowland and adjacent Fraser Lowland and suggest that the Sumas advance may not have climatic significance sufficient to justify calling it a stade. Abundant icebergs in the northern Puget Lowland and locally coalescing alpine and piedmont glaciers in the eastern



Topographic base Victoria and Concrete 1:250,000 sheets

Figure 2. Map showing the marine limit at about 13.5 ka, the distribution of ice-marginal deposits, submerged banks and areas of glaciomarine deposits, and sample sites in the northern Puget Lowland. The eastern extent of marine waters in the Skagit River Valley is not well known. Ice contact-deposits near the International Boundary mark the approximate southern limit of Sumas ice (Easterbrook, 1963, 1966).

Fraser Lowland (Armstrong, 1981) suggest that the Everson interval records a late part of Fraser glaciation, rather than a separate, distinctive nonglacial episode.

LANDFORMS OF ICE RETREAT AND THE MARINE INCURSION

Ice Recession and Grounding Positions

The distribution and morphology of ice-contact deposits, thick glaciomarine deposits, and submerged banks (Fig. 2) provide the best evidence for the pattern of ice recession from the northern Puget Lowland. The Puget lobe thinned and retreated northward without significant stillstands until the ice front reached the northern Olympic Peninsula (Thorson, 1980). An irregular zone of ice-contact features and deltas, graded to sea level, extends eastward along the northern Olympic and Quimper Peninsulas to the western slope of the Cascade Range, roughly delimiting the ice margin at this time (Thorson, 1980; Booth, 1987).

Emerged glaciomarine fans and deltas and the submerged banks in the eastern Strait of Juan de Fuca record grounding positions during calving retreat of the combined Juan de Fuca and Puget lobe to the north and northeast of the Olympic Peninsula (Anderson, 1968; Chrzastowski, 1980; Wagner and Wiley, 1983). Cattle Point, San Juan Island, for instance, is a 100-m-high morainal embankment exposing mainly gravelly marine outwash (Fig. 2). Seismic reflection studies of Cattle Point's submerged extension, Salmon Bank, suggest that it is formed of south-dipping, stratified deposits at least 80 m thick; extensive thicknesses of stratified material lie beneath other banks as well (Wagner and Wiley, 1983). Banks are complex, probably composite features, but their distribution, relation to emerged features, and composition suggest that they

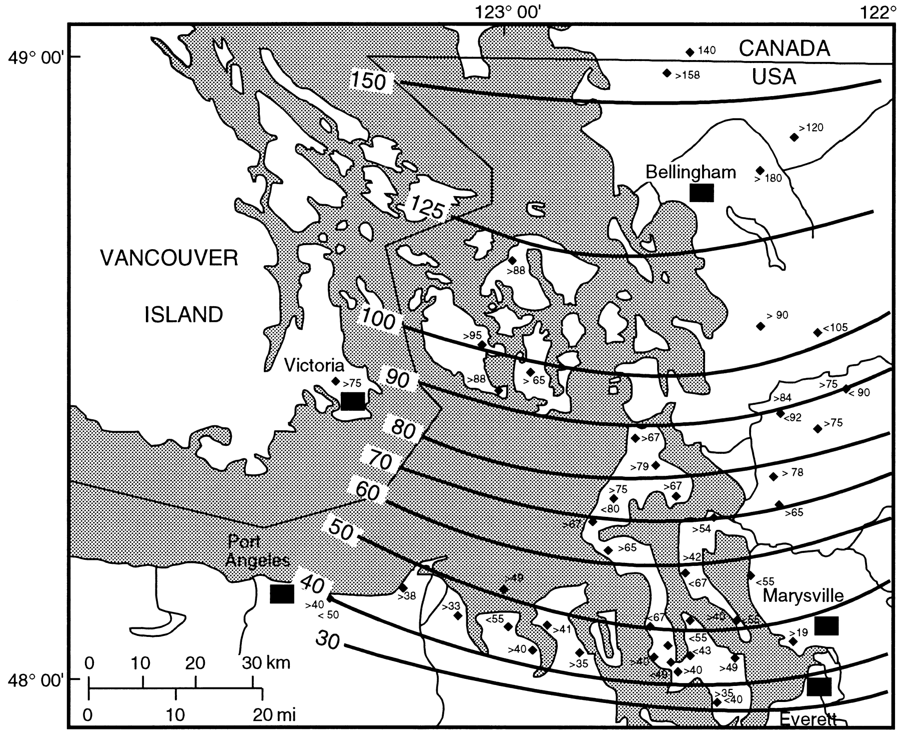


Figure 3. Map showing the altitude of the marine limit in the northern Puget Lowland, in meters. Plotted points are the highest local marine limit. Contour interval is 10 m south of the San Juan Islands, and 25 m north of that area. Contours do not reflect the point labeled >180 north-northeast of Bellingham, which represents the approximate upper limit of glaciomarine drift west of Deming (Easterbrook, 1963, 1976) (see text).

mark grounding positions of retreating ice lobes.

Vashon ice calved back rapidly north of the San Juan Islands, possibly constructing Alden Bank and an unnamed bank south of Point Roberts (Fig. 2) before grounding near the International Boundary before 13.2 ka (Easterbrook, 1976; Armstrong, 1981). The eastern margin of Fraser ice supported several valley glaciers that flowed into tidewater east of Vancouver (Armstrong, 1981) and contributed to the thick glaciomarine deposits north of Bellingham. Radiocarbon ages from wood show that ice finally disappeared from the Fraser Valley about 11.0 ka (Clague, 1980; Armstrong, 1981; Saunders et al., 1987).

Marine Limit

As Fraser ice retreated, marine waters invaded the northern Puget Lowland and adjacent areas of British Columbia, inundating most of the San Juan Islands and Whidbey Island, the lowland north of Bellingham, and extensive areas covered by Holocene alluvium of the Snohomish, Stillaguamish, Skagit, and Samish Rivers. The marine limit is approximate in the eastern Skagit Valley and northeast of Bellingham (Fig. 2) but as shown probably represents the minimum land area covered by the Everson sea in Washington. We have used field and published data for the highest local marine limit to plot Figure 3, which shows contours for those data in the northern Puget Lowland only. The elevation of the marine limit is not well constrained north of the Skagit Valley, but the point near Deming (>180) is anomalously high. The marine limit near Vancouver, some 50 km northwest of Deming, for instance, is about 180 m.

EVERSON SEDIMENTARY ENVIRONMENTS

Everson-age deposits thicker than 5 m are preserved in narrow zones near where ice grounded and north of Bellingham, as extensive fills along river valleys, and as a widespread but thin (<5 m) mantle of pebbly silt (Pessl et al., 1989). Deposits range from matrix-supported, cobble-rich diamictons to laminated silty clay in a variety of contact relations; sharp lithologic changes over short distances dominate most thick (>5 m) exposures.

Representative Stratigraphic Sections

A few sites in the northern Puget Lowland expose glaciomarine deposits that formed at the margin of grounded Fraser ice. Deformed glaciomarine sediments within kettled topography on the northern Olympic Peninsula (Fig. 4, site 11) and at Penn Cove on central Whidbey Island (Easterbrook, 1992) accumulated in an ice contact environment (Domack, 1982, 1984) where glaciomarine deposition was contemporaneous with melting of ice masses more than 100 m long and 30 m thick. Raised marine fans and deltas (?) on San Juan Island (Fig. 4, sites 5 and 14), on the northern Kitsap Peninsula (Thorson, 1980), and on Whidbey Island (Fig. 4, site 9) were deposited beside ice.

Exposures of Everson-age massive to laminated marine silt and fine sand <3 m thick are common in the Mount Vernon–Conway area (Fig. 4, site 2), north of Marysville (Fig. 4, site 6), on central Whidbey Island (Domack, 1984), north and east of Bellingham (Easterbrook, 1976), and in the western Fraser Lowland (Armstrong, 1981). Two sections are of particular interest because they record transitions from glaciomarine or marine sedimentation to lacustrine or lagoonal conditions. The Big Lake site (Fig. 4, site 3) exposes fossiliferous pebbly sand and gravel that lie beneath silt containing freshwater gastropods. The site was probably a beach at about 13.3 ka. At Davis Bay (Fig. 4, site 4), fossiliferous lagoonal deposits lie on shell-rich intertidal deposits; ages for the two units overlap at about 12.6 ka.

Wave-cut surfaces, strandlines, and beach deposits are preserved on the moraine at Cattle Point (Fig. 4, site 5), on a delta at Butler Hill (Easterbrook, 1979, 1992), and at other locations where an extended fetch, southerly aspect, and easily eroded sediment produced a record of emergence through the littoral zone. We refer to the thin, coarse, weakly stratified sediment as emergence deposits. Extensive sequences of strandlines visible on air photos are marked at many field sites, however, by a gravel lag only a few decimeters thick.

The Everson type section near Deming, Washington (Fig. 4, site 12) records a complex transition from glaciomarine to subaerial sedimentation at about the same time as the Sumas readvance to the International Boundary (Saunders et al., 1987). At the type section, fluvial or deltaic sediment that buries rooted stumps is exposed between massive units of pebbly marine silt (Easterbrook, 1963). This section suggests glaciomarine, then fluvial-deltaic, then glaciomarine, and finally terrestrial conditions at about 12 ka, providing the best documented evidence of a local, temporary reversal in rapid emergence of the northern Puget Lowland and nearby areas.

Paleoecology

Invertebrate fossils are relatively abundant in Everson-age glaciomarine and marine sediments in the northern Puget Lowland (Appendix Table A3). They represent cold-water species that have strong cool-temperate to Arctic affinities (Balzarini, 1981, 1983). Foraminifera are representative of high-latitude, cold, shallow-water assemblages presently found in southeast Alaska (Smith, 1970), and macrofossils suggest a marine environment comparable in temperature to that of the northern Gulf of Alaska (Wagner, 1959; Mallory et al., 1972; Shaw, 1972; Armstrong, 1981; Balzarini, 1981). Everson-age marine waters were cold (annual temperature range of 0 to 15 °C) and shallow (0 to 30 m) south of Lummi Peninsula, indicated by abundant elphidiid and *Bucella* foraminifera and by articulated intertidal and shallow subtidal bivalves. Waters were slightly deeper to the north (15 to 30 m, locally up to 60 m), indicated by abundant cassidulinid and elphidiid foraminifera and by the bivalve *Nuculana* in growth position. Assemblages suggest persistently brackish water (25‰ to 30‰) south of Bellingham and near normal salinity (30‰ to 35‰) to the north, where circulation was more open. The environment was one of a bay or estuary, suggested by the presence of many species adapted to fluctuating salinity. Low faunal abundance and well-preserved fossils suggest that sedimentation rates were high.

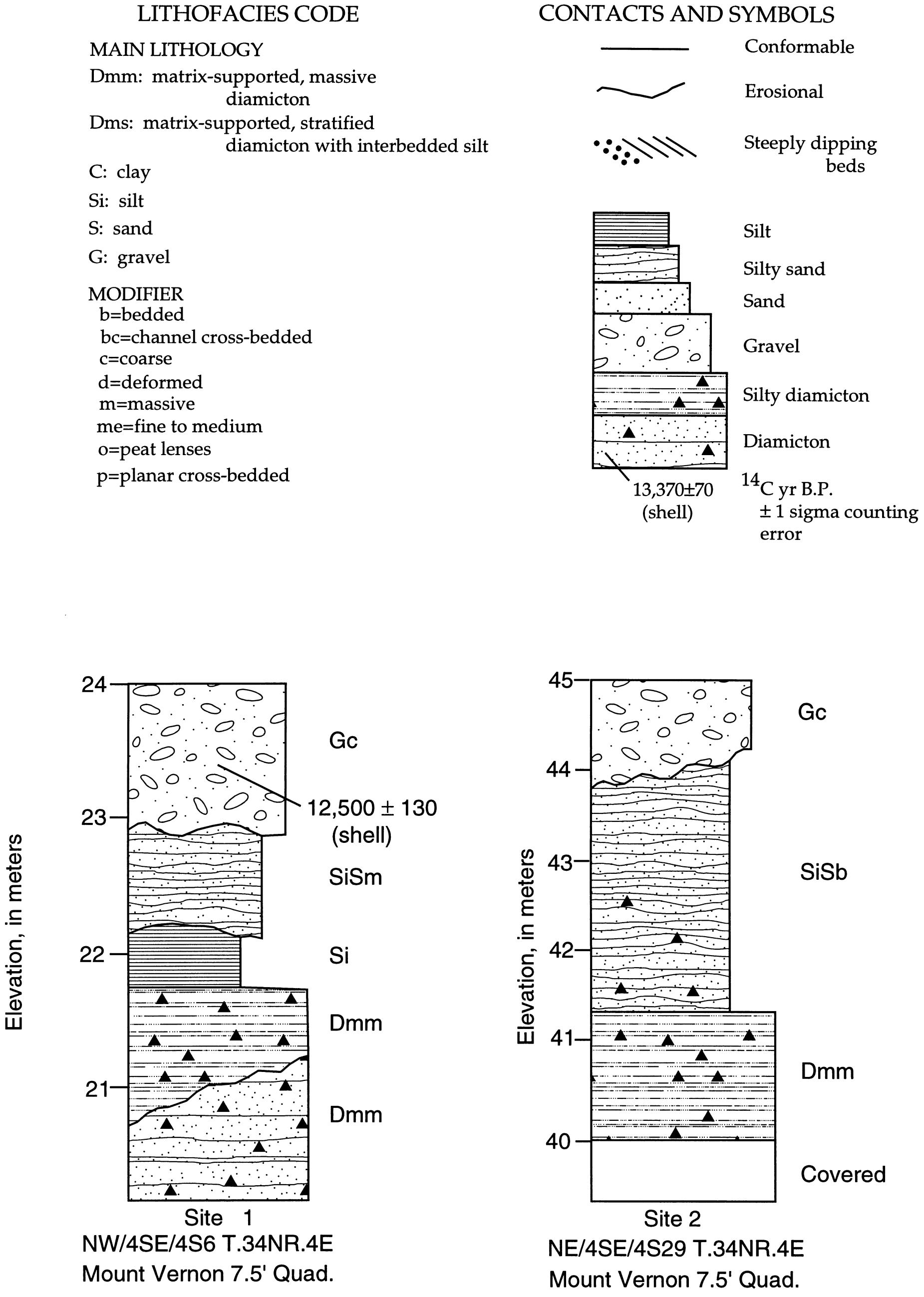


Figure 4. Stratigraphic sections of Everson-age deposits at selected sites, northern Puget Lowland. Site 1, Mount Vernon N., and site 2, Mount Vernon S.

Mineralogy and Geochemistry

The mineralogy and geochemistry of glaciomarine sediment from the Puget and Fraser Lowlands does not clearly distinguish glaciomarine deposits from terrestrial glacial drift. Bulk glaciomarine samples and till both are dominated by quartz and feldspar in a ratio of about 3:2; minor amounts of amphibole, mica, and chlorite also are present (Pevear et al., 1984). Differences in clay mineralogy between individual samples of till and glaciomarine sediments are typically small, suggesting homogenization by erosion and deposition processes. The clay (<2 μm) fraction of glaciomarine sediment is dominated by smectite, chlorite, and mica (muscovite and biotite) and minor amounts of kaolinite, amphibole, feldspar, and quartz; vermiculite and mixed-layer intergrades of mica, chlorite, and expandable clays are present in samples from weathered outcrops (Pevear et al., 1984). Most glaciomarine drift is a rock flour produced by abrasion of bedrock.

Neither boron nor sodium content clearly discriminates glaciomarine from nonmarine samples (Appendix Tables A2 and A3). Collectively, the data suggest that boron either was removed by postdepositional leaching or was never absorbed due to low depositional temperatures (Walker, 1975) or low salinity. Although some Pleistocene glaciomarine sediment is enriched in Na com-

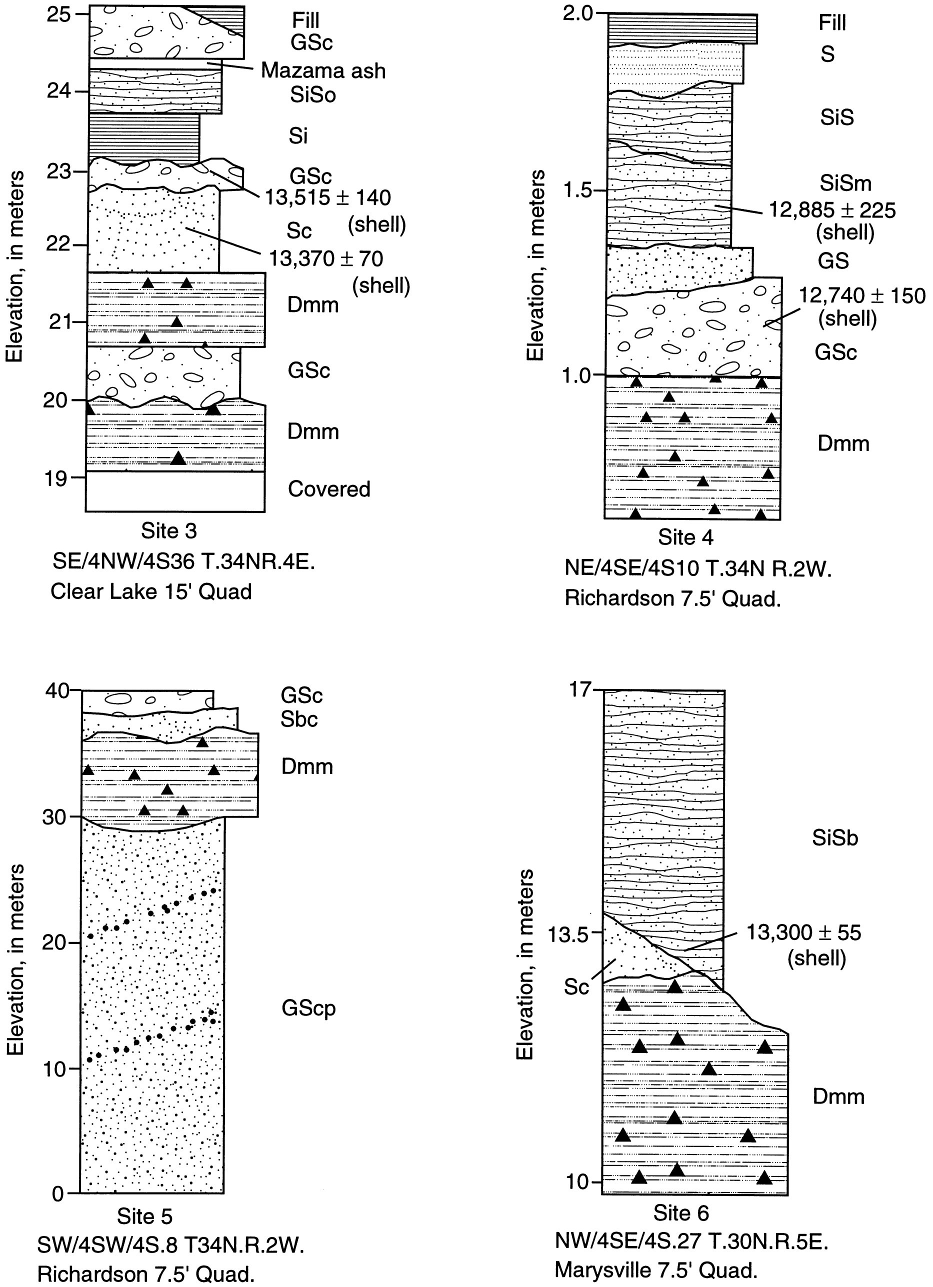


Figure 4. (*Continued*). Site 3, Big Lake; site 4, Davis Bay; site 5, Cattle Point; and site 6, Marysville.

pared with till, many samples evidently have been leached of original Na by ground water rich in Ca and Mg. Exchangeable sodium thus is not a reliable indicator of glaciomarine deposits in the Puget Lowland.

EVERSON FACIES RELATIONSHIPS

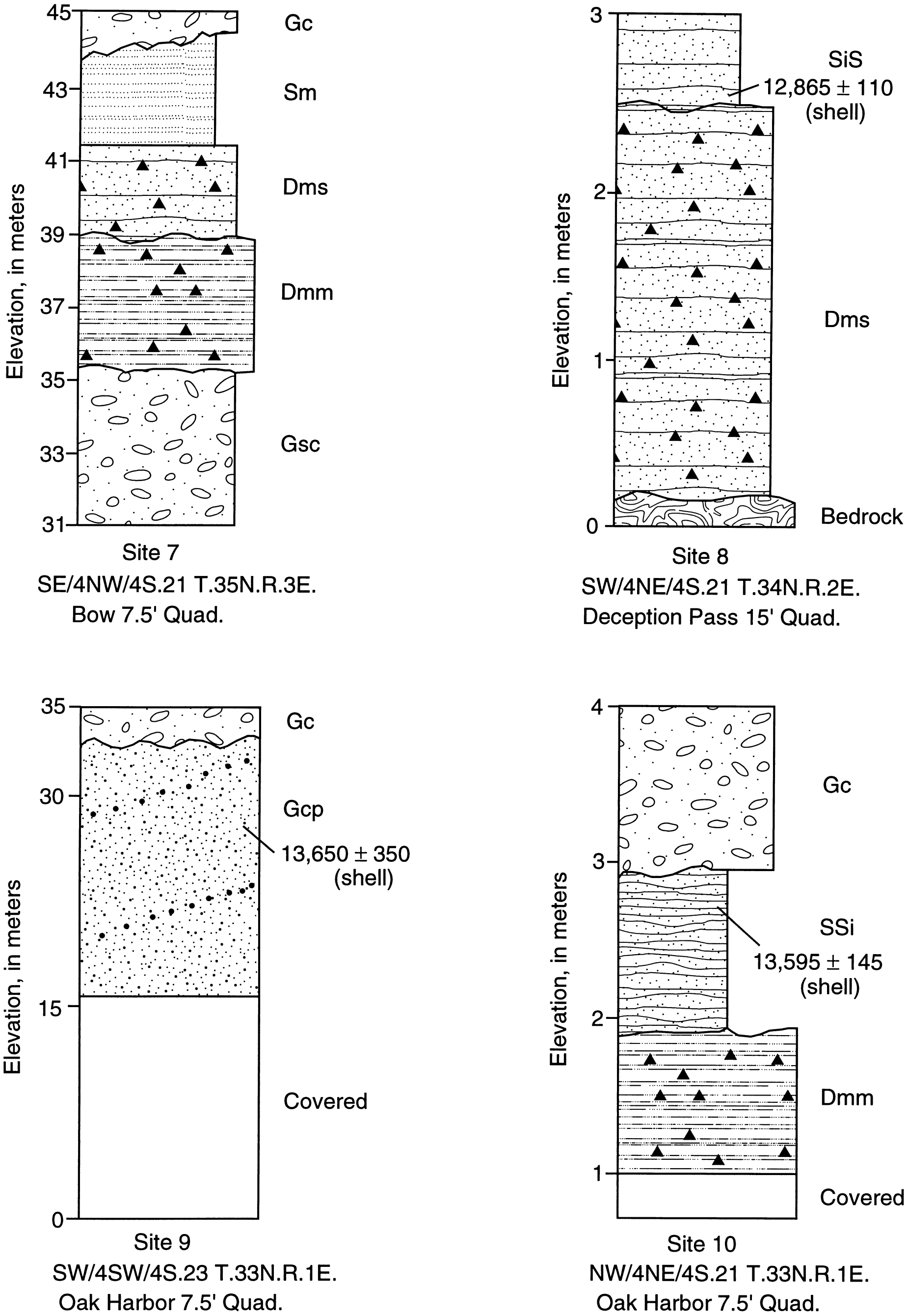
Everson sedimentary facies record depositional and erosional processes near the retreating ice margin, in ice-free marine and estuarine zones, and in the wave zone when isostatic rebound lifted older deposits above sea level. We subdivided the glaciomarine facies based on (1) field evidence, including distance from the probable ice margin (Fig. 5); (2) processes inferred from late Pleistocene glaciomarine deposits on Whidbey Island (Domack, 1982) and in eastern North America (Thomas, 1977; Thompson et al., 1983; Ashley et al., 1991) and from ancient glaciomarine deposits (Eyles et al., 1985); and (3) analogy to depositional environments at modern tide-water glaciers in southeastern Alaska (Powell and Molnia, 1989; Powell, 1991). Glaciomarine depositional processes and facies in Glacier Bay, although complex, can be grouped into several overlapping zones that extend from beneath the ice to tens of kilometers from the ice margin (Powell, 1984; Powell and Molnia, 1989). The ice-proximal zone extends a few hundred meters from the ice front; sedimentation is dominated by subaqueous outwash and debris flows, turbid underflows, and subglacial processes (Fig. 5). In a transitional zone 0.1 to 10 km from the ice margin, deposition occurs from turbid underflows, dispersed melt water and icebergs, and resedimentation. In the distal zone, more than 10 (?) km from the ice margin, sedimentation is controlled by dispersed melt water, currents, resedimentation, and icebergs. Sediment flux per unit time decreases logarithmically with distance away from grounded tide-water fronts (Powell and Molnia, 1989). We use these generalizations to guide our interpretation of Everson-age glaciomarine deposits from the northern Puget Lowland.

Sediment deposited in glaciomarine and marine facies include planar- and cross-bedded gravel, gravelly diamicton, pebbly silt, cross-bedded silty sand, and massive silt (Fig. 6). Each glaciomarine facies includes distinctive lithologic types (Table 1) produced by processes ranging from proximal to more than 10 km from the grounded ice margin. We also recognize marine, estuarine, and emergence facies. These facies were not directly influenced by the retreating ice margin or icebergs, lack dropstones or lenses of coarse sediment, and were derived from rivers and older deposits that were remobilized by mass movements, currents, and waves. Emergence facies formed during the rapid fall of local sea level after about 13.0 ka.

Proximal Facies

Coarse rocks that dominate proximal facies (Table 1) are exposed in morainal ridges or kettled topography. Extensive zones of such deposits (Fig. 2) occur on southern San Juan Island, on central Whidbey Island, on the Olympic Peninsula near Sequim, in the Fraser Lowland (Armstrong, 1981), and near Victoria on Vancouver Island (Hicock et al., 1981). On San Juan Island, the moraines are composed primarily of planar and cross-bedded gravel, coarse diamicton, and channel-filling, silt-rich diamicton. Sorted deposits generally become finer upward and from north to south in the moraines, where preserved thicknesses are as much as 100 m. Deposits at Cattle Point and at Pear Point (before gravel mining) each has a volume between 30 × 10^6 and 80 × 10^6 m^3. Several of the submerged banks south of San Juan Island contain as much as 1 km^3 of sorted deposits. The elongate shape of some banks may record slow retreat of actively calving margins (Powell, 1981), as well as extensive modification by waves and currents during Holocene time.

Gravel-rich proximal facies are exposed in thick sections near Sequim, Coupeville, and at Partridge Point (Fig. 2), where large ice blocks apparently were buried in gravels that subsequently collapsed, producing a kettled landscape. In some places, planar-bedded, pebble-poor marine silt overlies deformed sand and gravel, indicating that marine sediment was deposited after melting of the ice blocks. Our proximal facies include facies mapped by Domack (1983) in the Oak Harbor area mainly as (1) ice-marginal sediment flow, (2) proximal melt-water fan, and (3) turbidite channel.



Site 7
SE/4NW/4S.21 T.35N.R.3E.
Bow 7.5' Quad.

Site 8
SW/4NE/4S.21 T.34N.R.2E.
Deception Pass 15' Quad.

Site 9
SW/4SW/4S.23 T.33N.R.1E.
Oak Harbor 7.5' Quad.

Site 10
NW/4NE/4S.21 T.33N.R.1E.
Oak Harbor 7.5' Quad.

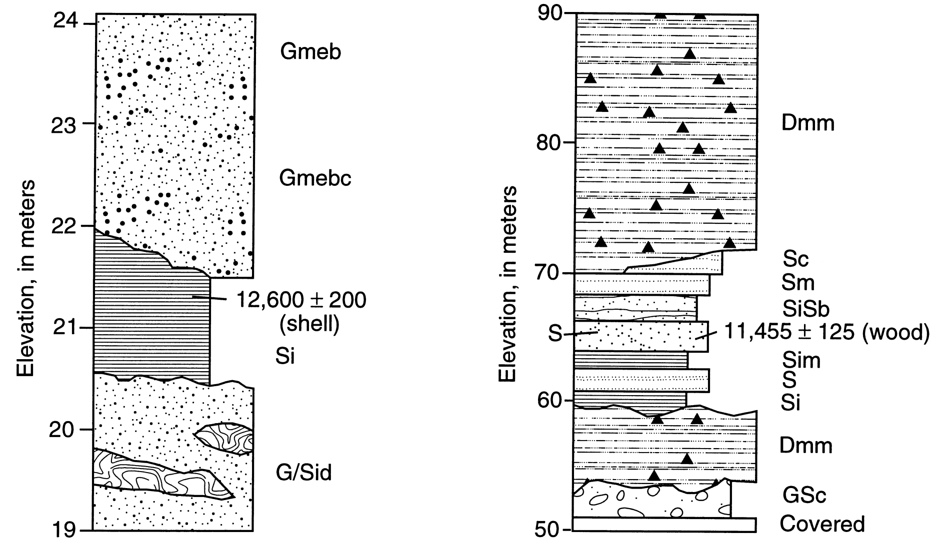
Figure 4. (*Continued*). Site 7, Bay View Ridge; site 8, Kiket Island; site 9, Whidbey Island–Naval Air Station; and site 10, Whidbey Island–SB.

Fossils are rare in most proximal sediment. *Hiatella arctica* and *Mytilus* sp. shells from gravel in a proximal gravel fan-delta 4 km northwest of Oak Harbor (Fig. 4, site 9) were dated at 13.6 ± 0.3 ka (Beta-1319). Marine bivalves have also been found at Cattle Point (Easterbrook, 1969), and *Mytilus* spp., *Hiatella arctica*, and *Balanus* sp. dated at 13.2 ± 0.1 ka (Beta-70971) occur in coarse gravel at Pear Point. Marine organisms lived in the ice-proximal environment of the northern Puget Lowland and Fraser Lowland (Wagner, 1959), much as they do near the margin of calving glaciers in south-southeast Alaska (Powell, 1981; Powell and Molnia, 1989). Light requirements of many of the fossil marine organisms show that they did not live beneath shelf ice (Crandall, 1979; Armstrong, 1981; Balzarini, 1981).

Transitional Facies

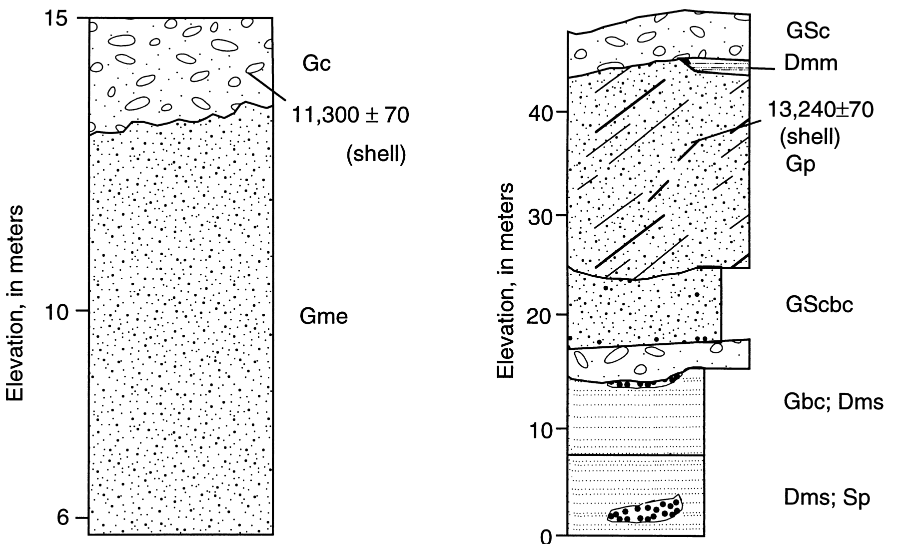
Transitional facies are finer grained, richer in dropstones, and contain a more diverse fauna than proximal facies (Table 1). Deposits were derived from ice margins located 1 km to perhaps 10 km away (Domack, 1983). Thick sections of pebbly silt are common in the northeast Puget Lowland and adjacent Fraser Lowland but generally are <3 m thick south of Bellingham, except within a few kilometers of ice-contact zones, where they range up to 10 m thick. The substantial local thicknesses of Capilano Sediments (Armstrong, 1981) and the Bellingham glaciomarine drift (Easterbrook, 1963) record more than a millennium of sediment deposition when the ice margin lay near the International Boundary after about 13.5 ka.

Transitional facies are locally rich in fossils of invertebrates (particularly *Clinocardium ciliatum*) that lived in environments similar in depth, salinity, and temperature to those inhabited by the sparse fauna of the proximal facies. At site 10 (Fig. 4), massive sandy silt containing sparse pebbles, bivalves in growth position, and abundant shallow marine ostracodes was dated at 13.6 ± 0.15 ka (Beta-1716). This age is equivalent to that of shell in an ice-contact delta 5 km east, demonstrating that proximal and transitional facies accumulated near one another at about the same time.



Site 11
NE/4NW/4S.1 T.30N.R.4W.
Dungeness 7.5' Quad.

Site 12
NW/4NE/4S.34 T.39N.R.4E.
Lynden 15' Quad.



Site 13
SW/4NW/4S.9 T.33N.R.3E.
Utsalady 7.5' Quad.

Site 14
NE/4SE/4S13 T.35N. R.3W.
Friday Harbor 7.5' Quad

Figure 4. (*Continued*). Site 11, Potholes; site 12, Deming; site 13, Dodge Valley; and site 14, Pear Point.

Distal Facies

Deposits of the distal facies interfinger with and overlie pebbly glaciomarine silt of the transitional facies. Deposits are thickest (locally >10 m) from Stanwood north to Butler Hill, north of Bellingham, and locally on the northern Olympic Peninsula and central Whidbey Island. Isolated dropstones and gravel lenses allowed us to separate distal facies from marine and estuarine deposits (see Table 1) in some areas. Near the mouths of major rivers such as the Skagit and Stillaguamish, however, we were unable to map consistently a contact between these facies. Absence of dropstones in silty sands suggests that floating ice was uncommon by 13.3 ka near Marysville (Fig. 4, site 6) and did not reach Mount Vernon (Fig. 4, site 1) after 12.5 ka.

Marine and Estuarine Facies

Marine and estuarine deposits in most surface exposures (Fig. 4, sites 1, 2, 4, and 7) are fine grained and thin (<3 m). At least some sandy deposits included with this facies accumulated in shallow (<10 m) water. Texture and sedimentary structures demonstrate a shallow, subtidal origin for sandy sediment overlying pebbly silt (Siegfried, 1978) at Bay View Ridge (Fig. 4, site 7) and for deposits south of Mount Vernon (Fig. 4, site 2), the upper two to three meters of deposits north of Stanwood, and the fossiliferous deposits on Kiket Island (Fig. 4, site 8). Microfauna at the three latter sites indicate tidal to shallow subtidal environments (Balzarini, 1983).

Rivers, dispersed melt water, and waves or currents evidently transported most sediment of the distal, marine and estuarine facies. The modern Skagit, Stillaguamish, Snohomish, Dungeness, and Elwha Rivers transport sediment equivalent to a thickness >0.5 mm yr^{-1} into northern Puget Sound (estimated from Nelson, 1971). These rivers would have delivered higher loads during Everson time as fluvial systems transported sediment removed from unvegetated, unconsolidated deposits freshly bared by ice retreat. In addition, sediment-rich water derived from Fraser ice near the International Boundary and from meteoric sources entered the Skagit Valley along the Columbia outwash channel during Everson time (Easterbrook, 1992, Fig. 20).

Emergence Facies

Sandy deposits containing in situ bivalves, boulder lags, and finer deposits record shallowing and emergence of land from the Everson sea (Table 1). Fossiliferous lagoonal and lacustrine deposits lie directly above shell-rich intertidal or beach deposits at several locations, dating precisely the emergence of the land (Fig. 4, sites 3 and 4). Intertidal pebbly sand and sand also have been reported by Armstrong (1981) and are doubtless widespread but difficult to identify without diagnostic fossils. Wave-cut surfaces are prominent at Cattle Point, on Lummi and central Whidbey Islands, and north of Bellingham (Easterbrook, 1979; Armstrong, 1981). At Butler Hill, successively younger deltas and terraces at the mouth of the Columbia Outwash channel were graded to a falling sea level (Easterbrook, 1966).

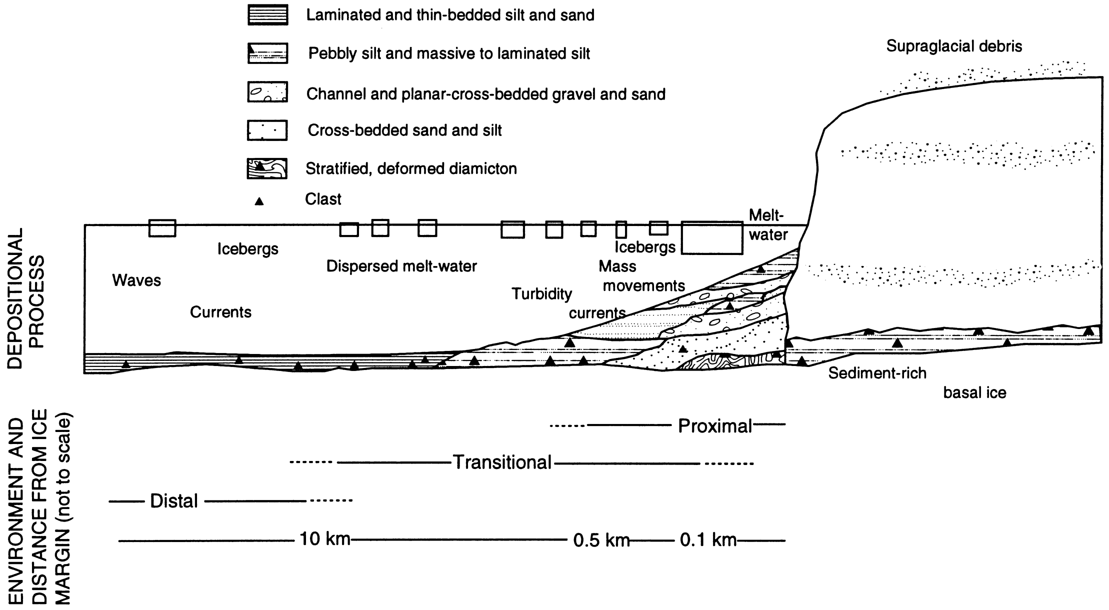


Figure 5. Relationship of glaciomarine deposits to inferred depositional mechanisms and environments during retreat of Vashon ice from the northern Puget Lowland (modified from Nelson, 1981; Molnia, 1983; and Powell, 1983, who described modern analogs on Baffin Island and coastal Alaska).

AGE OF EVERSON DEPOSITS

Published ^{14}C ages from material (mostly shell) deposited during deglaciation of the northern Puget Lowland (Easterbrook, 1992; Yount et al., 1980) and adjacent British Columbia (Clague, 1980, 1981; Armstrong, 1981) and new ages reported here (Appendix Table A1) demonstrate that the Everson interval lasted from about 13.6 to about 11.3 ka. The oldest ages reported for shell from glaciomarine sediment in the Puget Lowland are 13.8 ± 0.3 west of Seattle (Anundsen et al., 1994), 13.6 ± 0.14 ka on Whidbey Island (this report), and 13.5 ± 0.22 ka in British Columbia (Armstrong, 1981). Basal peat ages of 13.6 ka from the Seattle area (Leopold et al., 1982), 13.1 ± 0.13 ka from Vancouver Island (Alley and Chatwin, 1979), and 12.9 ± 0.13 ka north of Sedro Woolley (Rubin and Alexander, 1958) give minimum ages for deglaciation of these areas. The array of ^{14}C ages suggests that the entire Puget Lowland was deglaciated by 13.5 ka.

The precision and accuracy of the ages in Table A1 are important for the chronology of ice retreat and for our estimates of uplift rates. Our oldest shell ages (13.6 to 13.5 ka) demonstrate that the Everson sea had invaded the Puget Lowland as far north as northern Whidbey Island nearly 500 yr earlier than had been known previously. We discuss the possibility of errors from contamination, reworking, and old carbon in sea water below.

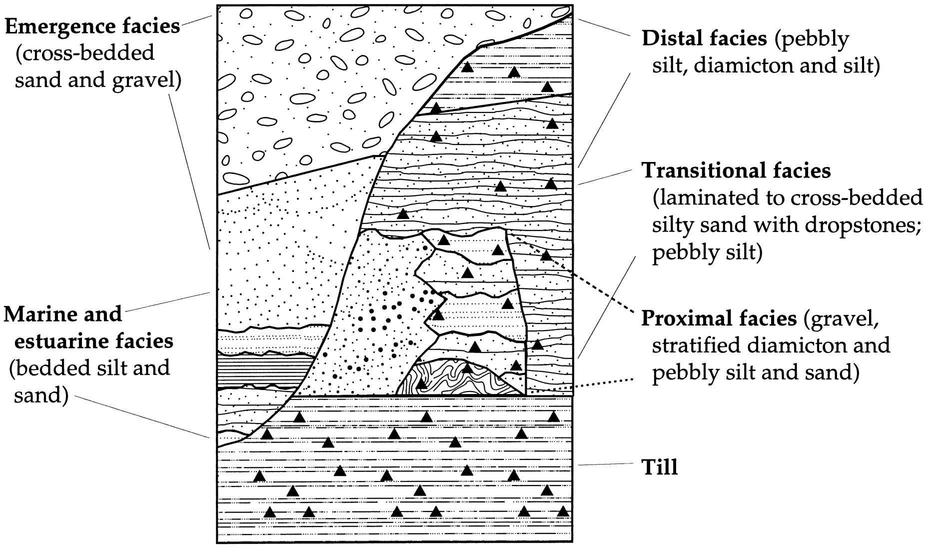


Figure 6. Sketch showing stratigraphic and lithologic relations among Everson-age deposits.

Contamination of samples with "young" or "old" material is unlikely to have produced large (>500 yr) errors in any of the ages listed in Table A1. With the exception of leached shell from Pear Point, none of the material submitted in this study appeared more than slightly altered. Contamination of samples by younger material is a possibility, but zero-age material would have to make up more than 5% of a sample to alter significantly the age, and this is unlikely for the clean shells dated in this study. Shells from two separate samples collected at sites 1, 3, and 4 (Fig. 4) and dated by two different laboratories gave ages that overlapped at the 1σ level. These data suggest that the ^{14}C ages are accurate and that contamination was not a problem at these localities.

Shells from Mount Vernon North, Dodge Valley, and Whidbey Island Naval Air Station (Fig. 2) were not in growth position and could have been reworked from older deposits. However, the Whidbey and Big Lake ages of 13 650 ± 350 yr and 13 370 ± 70 yr, respectively, are among the oldest that have been obtained and demonstrate that mollusks inhabited the area about 13.5 ka. The Dodge Valley age (11 330 ± 70 yr) is

TABLE 1. SEDIMENTOLOGY, PALEONTOLOGY, AND INFERRED DEPOSITIONAL ENVIRONMENT OF GLACIOMARINE, SHALLOW MARINE, AND ESTUARINE AND EMERGENCE FACIES DEPOSITED DURING THE EVERSON INTERVAL

Facies	Lithology and sedimentary structures	Fossils	Stratigraphic context	Inferred paleoenvironment
Proximal	Gravel, gravelly diamicton; locally sand and silty sand. Moderately well stratified to nonstratified. Gravelly deposits locally cross-bedded to planar-bedded with dips as steep as 32°. Contorted beds, faults, water-escape structures, and sharp lateral changes in lithology common. Generally 5 to 35 m thick	Sparse bivalves (*Mya truncata*; *Hiatella arctica*) at Whidbey Island Naval Air Station. *Mytilus* sp., *Macoma* spp., and *Balanus* at Pear Point	Lies unconformably above Vashon till, Vashon recessional outwash or ice-contact deposits, or pre-Vashon sediment. Interfingers with or lies beneath other glaciomarine and marine facies	Grounding zone subaqueous fans, debris flows, deltas, turbid underflows and subglacial deposition
Transitional	Interbedded sand and silt, pebbly silt, and massive silt enclosing widely scattered lenses of sandy gravel. Sand and silt poorly to locally well stratified and trough cross-bedded at some locations; pebbly silt massive to faintly stratified. Silt displays local, discontinuous laminations. Contacts of stratified units sharp and irregular to gradational with pebbly silt. Generally 2 to 10 m thick	Sparse to abundant bivalves (*Hiatella arctica*, *Nuculana* sp., and *Clinocardium ciliatum*)	Lies above proximal glaciomarine deposits or Vashon till. Lies beneath distal glaciomarine or emergence deposits	Deposition from turbid underflows, dispersed melt water and icebergs, and redeposition 0.1 to 10(?) km from grounded ice
Distal	Interbedded to laminated silt and clay, and massive silt containing sparse pebbles and local sandy partings. Weakly to well stratified. Intrafacies contacts gradational. Generally less than 5 m thick	Locally abundant bivalves (*Nuculana fossa, Hiatella arctica,* and *Macoma* spp.)	Generally lies above proximal or transitional glaciomarine deposits, and beneath thin marine or emergence deposits	Deposition mainly from dispersed melt water, resedimentation, currents, and icebergs more than 10(?) km from grounded ice
Shallow marine and estuarine	Sandy silt, silt, and clay. Thin bedded to laminated; locally massive with thin laminations. Local water (and gas) escape structures. Generally less than 5 m thick; thicker near major rivers	Bivalves (*Saxidomus giganteus*, *Mya, arenaria,* and *Macoma* spp.) and worm tubes (*Serpula vermicularis*) locally abundant	Lies above Vashon till and glaciomarine deposits and beneath emergence deposits	Deposition from material suspended by currents and waves, and from fluvial suspended load
Emergence: intertidal beach and lagoon	Sand and silty sand with sparse gravel, and interbedded sand and silty sand (low intertidal to beach); gravelly sand and gravel (upper beach and back beach); silt, organic-rich clayey silt enclosing sparse lenses of sand and fresh-water mollusks (lagoonal); and organic-rich, massive silt and coarse sand (eolian). Intertidal deposits well-stratified, rippled to trough cross-bedded. Beach deposits poorly to moderately well sorted, locally cross-bedded. Lagoonal deposits massive to laminated. Eolian silt massive; eolian sand planar cross-bedded	Sparse in gravelly deposits. Barnacles (*Balanus* spp.) attached to rocks, bivalves (*Saxidomus giganteus*) locally abundant in intertidal deposits. Brackish and fresh-water gastropods common at Big Lake and Davis Bay	Lies unconformably on Vashon and glaciomarine deposits. Generally forms the surface layer	Deposition by wave and current action, from suspension in lagoons, and by eolian activity

younger than most reported Everson dates, so reworking is not a significant problem.

Shell ages reported here and in the Everson literature are probably 600 to 800 yr older than those from wood that grew at the same time. The isotopic composition of local marine water and melt water is not known for late Pleistocene time, but shells probably formed from marine waters out of equilibrium with atmospheric carbon dioxide. Modern shells in the area from northern California to Alaska are too old by 700 to 800 yr (Robinson and Thompson, 1981). Most ages from paired, late Pleistocene wood and shell samples from the region (Clague, 1980; Armstrong, 1981; Anundsen et al., 1994) diverge by 400 to 800 ^{14}C yr. It is likely that reservoir effects account for much of this disparity. In the discussion below we assume that local differences in the reservoir correction for marine shells are small compared to other age uncertainties such as the amount of ^{14}C in the latest Pleistocene atmosphere.

ISOSTATIC REBOUND IN THE NORTHERN PUGET LOWLAND

Everson deposits record isostatic rebound of the northern Puget Lowland after about 13.5 ka. When Cordilleran ice retreated into

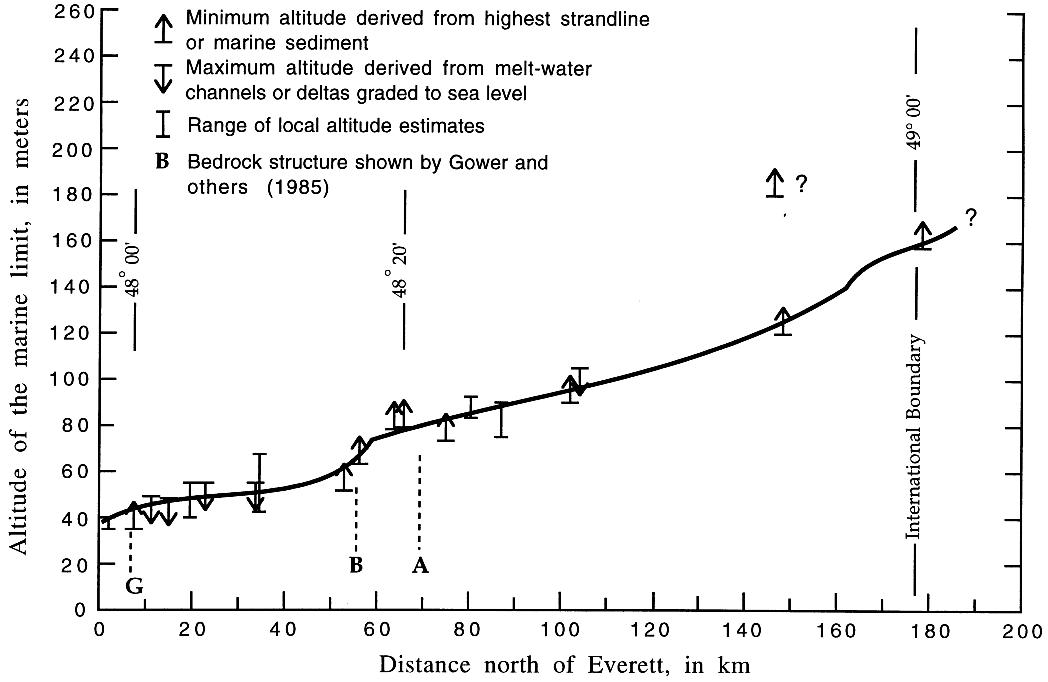


Figure 7. Maximum altitude of the marine limit within 20 km of long 122°15′00″W from near Everett, Washington, to the international boundary.

the Fraser Lowland, most of the present land area south of Everett lay above sea level, whereas present-day land to the north was covered by 40 to >150 m of marine water. Rebound was rapid. Most of the uplift at Victoria, British Columbia, for instance, had occurred by 10.0 ka (Mathews et al., 1970). The rate of sea-level rise has exceeded rebound rates since before 7.0 ka in the vicinity of southern Whidbey Island (Thorson, 1980; Eronen et al., 1987) and in the Fraser Lowland (Armstrong, 1981; Clague, 1983). The episode of relative sea level fall thus persisted only 3000 to 6000 yr.

Patterns and Rates of Uplift

The elevation of the marine limit, which rises gradually to the north (Figs. 2 and 3), ^{14}C ages, and estimates of eustatic sea level (Fairbanks, 1989) permitted us to analyze isostatic rebound in the northern Puget Lowland. We made the simplifying assumption that deglaciation was sufficiently rapid and the rise of sea level sufficiently slow that the marine limit at 13.5 ka defined a single horizontal plane. The assumption is reasonable because <500 yr (2σ differences) separate the earliest postglacial ^{14}C ages from Seattle, Whidbey Island, and the Fraser Lowland (Table A1; Armstrong, 1981; Booth, 1987). Eustatic sea level was ~100 m at 13.5 ka and rose about 10 m between 13.5 and 12.5 ka and about 20 m between 12.5

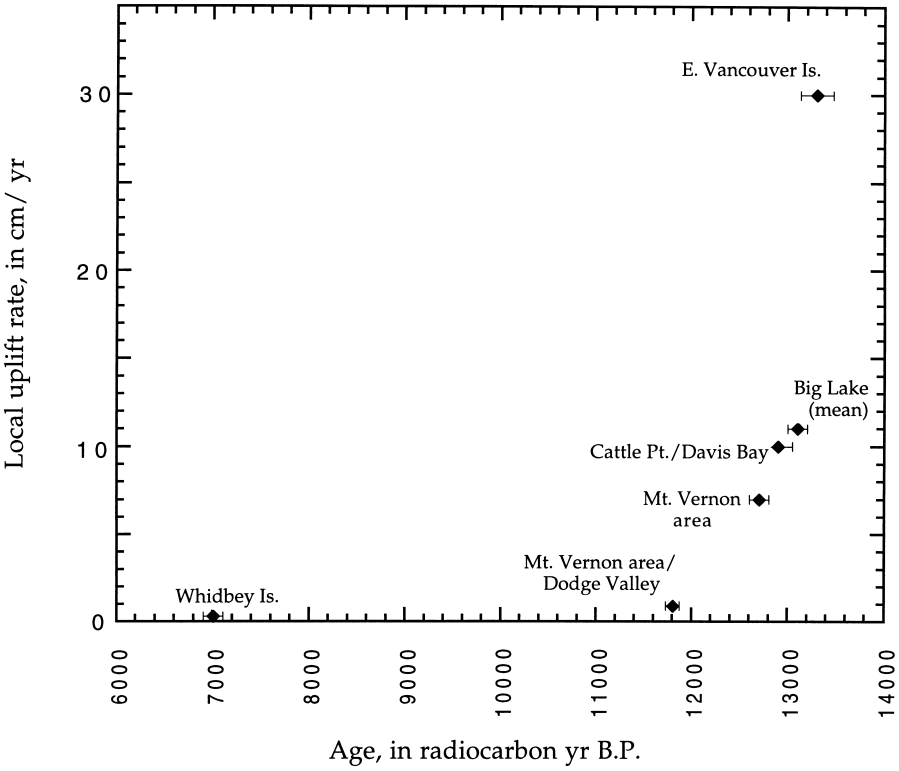


Figure 8. Approximate local rates of uplift in the northern Puget Lowland during the latest Pleistocene and early Holocene. Point for eastern Vancouver Island is from Mathews et al. (1970); Whidbey Island value is from Biederman (1967). Other localities are described in text. Error bars show 1σ uncertainty for ^{14}C ages.

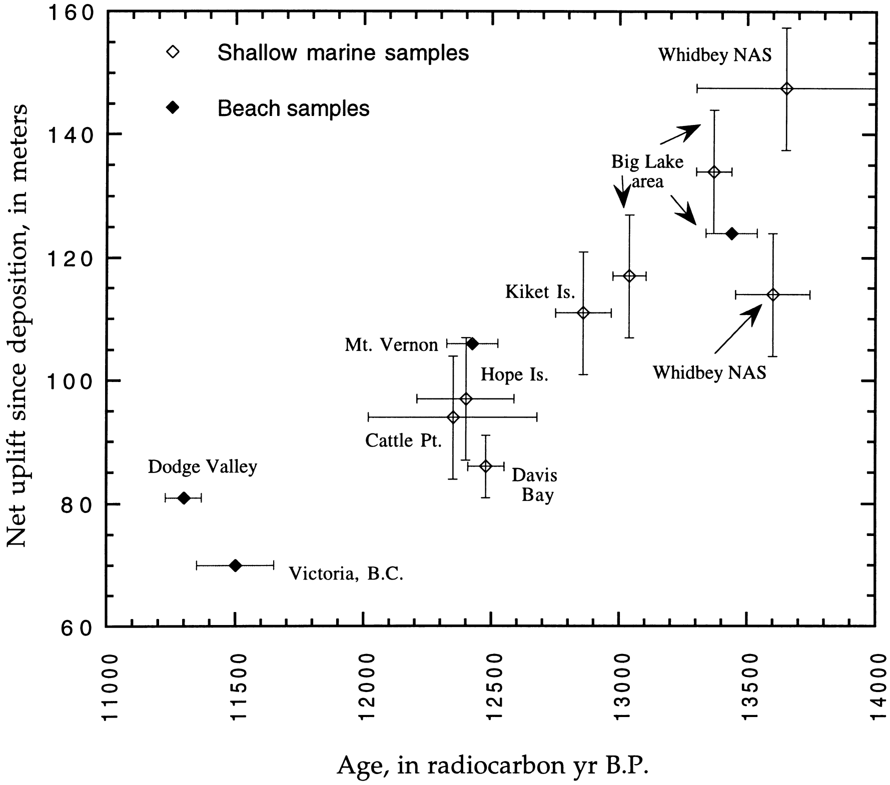


Figure 9. Net uplift of glaciomarine and marine sediment deposited within 15 km of latitude 48°25′30″N between 13.65 and 11.30 ka, calculated using the late Pleistocene eustatic sea-level data of Fairbanks (1989). Shallow marine samples are plotted as present elevation + 8 m, except for Davis Bay, which is a composite age of a shallow marine sample and a lagoonal sample and is plotted at +2 m. Elevation uncertainty for shallow marine samples is ±10 m (see text) except for Davis Bay, which is ±4 m. Elevation uncertainty of beach samples (±2 m) is shown by symbol height. Age uncertainty is plotted as ±1σ.

and 12.0 ka (Fairbanks, 1989). We calculated net uplift since the marine incursion at a site from its present elevation and the estimated elevation of Everson sea level. In a few areas, abundant data permit calculation of local isostatic rebound rates.

The marine limit, plotted approximately parallel to a flow line of Vashon ice (Waitt and Thorson, 1983), rises to the north at about 0.6 m km^{-1}, steepening locally to >1.3 m km^{-1} (Fig. 7). Thorson (1989) calculated an average uplift gradient of 0.84 m km^{-1} for the area south of Everett. Total postglacial uplift exceeds the relative values plotted in Figure 7 by the sea-level rise since 13.5 ka plus "restrained rebound," the isostatic uplift that occurred as continental ice thinned prior to invasion by the sea. Minimum isostatic uplift ranged from about 140 m near Everett (marine limit of 40 m + 100 m sea-level rise) to more than 250 m at the International Boundary. Total isostatic uplift could be as much as twice the minimum values (Thorson, 1989), but the amount of restrained rebound cannot be calculated from local geologic evidence.

The rate of isostatic rebound exceeded the rate of local sea-level rise throughout the study area after about 13.5 ka, rapidly raising marine and glaciomarine deposits above sea level. We do not have sufficient radiocarbon or sea-level control to characterize the uplift curve for a single site. Data in Table A1 and the literature (Biederman, 1967; Mathews et al., 1970), however, permit us to construct an approximate regional curve using local rebound rates measured at different times from closely spaced deposits whose relation to sea level is well defined (Fig. 8). For instance, if we assume that marine waters were 90 m deep at Davis Bay at 13.5 ka (see Fig. 3) and that the site emerged at about 12.6 ka (Table A1), the uplift rate for this period is 10 cm yr^{-1}. Rise of eustatic sea level, however, during the period makes 10 cm yr^{-1} a lower limit. Assuming that each of the areas plotted in Figure 8 was deglaciated at about 13.5 ka, the values suggest an exponential decrease in uplift rates with time, consistent with rates of local sea-level fall in the Fraser Lowland and eastern Vancouver Island. Data from British Columbia (Mathews et al., 1970; Armstrong, 1981; Clague et al., 1982) and the study area (Croll, 1980; Thorson, 1980) indicate that emergence in the first millennium following deglaciation occurred at rates >10 cm yr^{-1}, including an initial rate (first several hundred years) of more than 30 cm yr^{-1} (Mathews et al., 1970). After about 12.5 ka, the rate of uplift was 3–7 cm yr^{-1}. The 6 cm yr^{-1} rise of eustatic sea-level at about 12 ka (Blanchon and Shaw, 1995) might thus have exceeded rates of uplift for tens of years.

Anomalies

Inflections in the northward rise of marine limit (Fig. 7) and resubmergence at about 12.3 ka at several sites, best exemplified by the Deming locality (Mathews et al., 1970; Clague, 1981; Easterbrook, 1992), suggest complexities in the generally smooth record of emergence (Figs. 2 and 8). Inflections in the marine limit (Fig. 7) may reflect postglacial tectonism, stillstands during ice retreat, or incomplete data. For instance, at least two inflections in uplift curves for the southern and central Puget Lowland coincide with probable faults in pre-Quaternary rocks (Thorson, 1980, 1989). The best-determined change in gradient in the northern Puget Lowland, near 48°20′N, corresponds with fault zone A of Gower et al. (1985).

The Deming locality (Figs. 2 and 4, site 12; Easterbrook, 1963) provides the best stratigraphic evidence for anomalous submergence. Exposures and ^{14}C ages suggest that a zone of unknown areal extent emerged, submerged, and re-emerged in a period of <1000 yr following the initial marine incursion about 13.5 ka. Since substantial compaction of the lower glaciomarine unit and underlying till at Deming is unlikely, an oscillation of at least 35 m in local sea level is required to explain the thickness and sequence of deposits. The marine limit near Deming (>180 m; Easterbrook, 1976) is also anomalously high compared with areas to the northwest (Fig. 3). No Quaternary faults have been mapped near Deming.

The distribution of altitudes and ^{14}C ages of Everson deposits on a transect approximately normal to ice flow lines (Booth, 1987) (Fig. 9) suggests additional complexities in the pattern of isostatic rebound. If ice thickness was constant and ice retreated past each site at about the same time, uplift since deposition should be an exponential function of time. Uplift and deposit age are correlated (Fig. 9), but there are substantial deviations from a simple relationship. For example, deposits at Mount Vernon North (Fig. 4, site 1) and the marine/fresh water transition recorded at Davis Bay (Fig. 4, site 4) differ in elevation by 22 m, despite similar ^{14}C ages and latitude. At least six dated deposits (Easterbrook, 1969, 1992) lay below sea level at 12.2 ka, yet most of the sites in Figure 9 (at the same latitude) had emerged by about 12.5 ka and apparently have remained above sea level since then. Disparate local sea-level changes may be artifacts of (1) dating errors resulting from ^{14}C reservoir effects, (2) errors in water depths assumed for shell assemblages, (3) complex patterns of isostatic rebound, possibly including depression near the margin of Sumas ice, (4) tectonism, or (5) rapid changes in the rise of eustatic sea level.

The scatter in the uplift relationship (Fig. 9) probably does not result from artifacts in the data. There is no evidence that marine reservoir ages differed locally. Shell in intertidal or beach deposits records sea level within a meter or two. Water depth assignments for dated shallow subtidal shells, mainly *Saxidomus giganteus*, are more difficult to evaluate. *Saxidomus* is generally found in water <18 m deep, and associated microfossil assemblages are typical of water <30 m deep. Error bars in Figure 9 portray uncertainty in estimated sea level (thus net uplift) for shallow marine samples as ± 10 m.

Various workers (cf. Andrews, 1991) have suggested that an area depressed by glacial loading must be flanked by an outboard bulge forced up by asthenospheric flowage of material away from the depressed zone. Formation, collapse, and migration of a foreland bulge tens of meters high may help to explain the complex pattern of relative uplift most clearly recorded at Deming, as well as in adjacent British Columbia (Mathews et al., 1970; Clague, 1981, 1983). Booth (1987) concluded that evidence from geophysical models suggested that the magnitude of a marginal bulge was too small to explain the field relations at Deming. Rapid sea-level rise of nearly 100 m on the central continental shelf of British Columbia, however, was attributed mainly to collapse and migration of a bulge marginal to the retreating Cordilleran ice sheet at about 10 ka (Luternauer et al., 1989). Bulge migration and collapse also are supported by field evidence along the western margin of the Cordilleran Ice Sheet in British Columbia (Clague et al., 1982) and by similar evidence from eastern Canada (Quinlan and Beaumont, 1981; Scott et al., 1984). Sea-level rise of 13.5 m at about 12 ka (Blanchon and Shaw, 1995) may account for some of the differences in net uplift portrayed in Figure 9. We did not recognize evidence for resubmergence at Big Lake or Davis Bay, but Shaw (1972) described a site on Shaw Island (Fig. 2) where an Everson lagoonal deposit is covered by sediment containing subtidal macrofossils.

Passage of a foreland bulge, a sudden rise in global sea level, and a tectonic component of uplift, superimposed on regional isostatic adjustments, may have each played a part in determining the age and present elevation of uplifted glaciomarine sediment. No single process seems likely to account for the observed patterns of uplift, and recognition of interactions among multiple processes that rapidly change local base level is limited by the accuracy of shell ^{14}C ages. Precise dating of additional marine and glaciomarine deposits with specific altitude control will help to improve our understanding of spatial and temporal variation in the pattern of uplift.

SUMMARY

Fraser ice retreated rapidly from the northern Puget Lowland at about 13.5 ka, and marine waters briefly invaded about 10 000 km^2 of Washington and adjacent British Columbia during a period of rapid isostatic uplift. Locally thick deposits of glaciomarine facies mark grounding zones of the retreating ice, but many Everson-age deposits accumulated when the ice margin was tens of kilometers distant or formed in marine or estuarine environments. The marine limit formed during the Everson interval rises northward at about 0.6 m km^{-1}, reaching an elevation of >150 m at the International Boundary. Isostatic uplift at rates of 5 to >10 cm yr^{-1} raised all sites in this study above sea level before 11.3 ka, but at least one locality, at Deming, resubmerged before about 12 ka. Local changes in gradient of the marine limit and anomalous resubmergence may result from passage of a foreland bulge, a period of rapid sea-level rise, or a tectonic component of uplift.

ACKNOWLEDGMENTS

This report resulted from field studies by the Earth-Sciences Applications Project of the U.S. Geological Survey in the late 1970s and early 1980s. Jim Minard (U.S. Geological Survey) spent many days helping us refine our field observations and V. Standish Mallory (University of Washington) supervised the paleontologic analyses. The thoughtful review comments of Brian Atwater, Bernard Hallet, Derek Booth, John Clague, Robert Thorson, and James Yount greatly improved earlier versions of this paper.

APPENDIX

Tables A1, A2, and A3 follow.

TABLE A1. SELECTED NEW RADIOCARBON AGES FOR EVERSON-AGE DEPOSITS, NORTHERN PUGET LOWLAND, WASHINGTON[*]

Laboratory number	Location	Description[†]	Age, (^{14}C yr B.P.)	Remarks
BETA-1324	Everson (locality 12) 48°49.87'N, 122°16.39'W	Stump (*Pinus contorta)* in sand	11,455 ± 125	Elevation 66 m; nonmarine unit lies between glaciomarine units
BETA-1717	Davis Bay (locality 4) 48°27.64'N, 122°55.69'W	Shell (*Saxidomus* sp.) in intertidal sand and gravel	12,740 ± 150	Elevation 1 m
USGS-1234	Davis Bay (locality 4) 48°27.64'N, 122°55.69'W	Shell (*Saxidomus* sp.) in intertidal sand and gravel	12,480 ± 70	Shells adjacent to BETA-1717
BETA-1723	Davis Bay (locality 4) 48°27.64'N, 122°55.69'W	Fresh-water gastropods in lagoonal (?) silty sand	12,885 ± 225	Collected 30 cm above BETA-1717
BETA-1321	Mount Vernon North (locality 1) 48°26.56'N, 122°19.33'W	Shell (*Saxidomus* sp.) in beach gravel	12,500 ± 130	Elevation 23 m
USGS-1236	Mount Vernon North (locality 1) 48°26.56'N, 122°19.33'W	Shell (*Saxidomus* sp.) in beach gravel	12,350 ± 70	Shells adjacent to BETA-1321
USGS-782	Big Lake (locality 3) 48°23.71'N, 122°14.54'W	Shell (*Saxidomus* sp.) in silty sand	13,370 ± 70	Elevation 23 m
BETA-1322	Big Lake (locality 3) 48°23.84'N, 122°14.44'W	Shell (*Saxidomus* sp.) in sand and gravel (beach?)	13,515 ± 140	Elevation 24 m, 300 m NE of USGS-782
USGS-1235	Big Lake (locality 3) 48°23.84'N, 122°14.44'W	Shell (*Saxidomus* sp.) in sand and gravel (beach?)	13,180 ± 70	Shell adjacent to BETA-1322
USGS-787	Big Lake South 48°22.35'N, 122°12.80'W	Shell (*Saxidomus* sp.) in silty sand	13,040 ± 65	Elevation 24 m
BETA-1715	Kiket Island (locality 8) 48°25.27'N, 122°33.61'W	Shell (*Saxidomus* sp.) in silty sand	12,865 ± 110	Elevation 2 m
USGS-124	Dodge Valley (locality 13) 48°21.88'N, 122°26.27'W	Shell in beach gravel	11,330 ± 70	Elevation 14 m, Siegfried (1978)
BETA-1716	Whidbey Is. Naval Air Station South Beach (locality 10) 48°20.36'N, 122°41.20'N	Shell *(Clinocardium nuttalli)* in sandy silt	13,595 ± 145	Elevation 2 m
BETA-1319	Whidbey Is. Naval Air Station (locality 9) 48°19.42'N, 122°39.15'W	Shell *(Hiatella arctica)* in fan and deltaic gravel	13,650 ± 350	Elevation 24 m
BETA-1323	Potholes (locality 11) 48°07.71'N, 123°08.04'W	Shell (*Tresus* sp.) in silt	12,600 ± 200	Elevation 25 m
USGS-808	Marysville locality 6) 48°03.22'N, 122°09.13'W	Shell (*Saxidomus* sp.) in silt	13,300 ± 55	Elevation 13 m. Collected by J. P. Minard
USGS-1304	Coupeville Boat launch 48°13.24'N, 122°40.66'W	Shell in silt-rich diamicton	12,640 ± 150	Elevation 1 m. Collected by E. W. Domack
BETA-70791	Pear Point gravel pit (locality 14) 48°31.42'N, 123°00.41'W	Shell (*Hiatella arctica; Balanus* sp.) in fan gravel	13,240 ± 70	Elevation 40 m

[*] Ages from this study except as noted.

[†] Shell from marine to brackish water except where noted. See Table A3.

TABLE A2. SELECTED GEOCHEMICAL CHARACTERISTICS OF GLACIOMARINE, MARINE, AND NONMARINE SAMPLES

Sample number	Environment of deposition*	Cations[†]	X_{Na}	X_K	X_{Ca}	X_{Mg}	B[#]	N**	Sample location
			Equivalent percent[§]						
39-36	M	62(28)[††]	41	25	5	30	50 ± 3	2	Modern glaciomarine,
39-291	M	52(28)	32	19	24	25	42 ± 1	2	Icy Bay, Alaska
39-566	M	56(28)	33	19	23	24	50 ± 6	2	
45-45	M	135(28)	5	5	64	26	43 ± 1	2	
RM-1	M	115	45	5	15	35	120 ± 6	5	Recent marine,
RM-2	M	118	31	7	20	42	122 ± 10	5	Bellingham Bay, Washington
LW-4	NM	37	2	3	76	18	26 ± 6	2	Recent lacustrine, Lake
LW-5	NM	35	2	3	74	20	23 ± 4	2	Whatcom, Washington
GO-2A	MO	42(37)	2	2	45	52	31 ± 4	7	Everson-age
GO-2A	MO	39(37)	2	3	51	43	N.D.[§§]	N.D.	glaciomarine,
GO-2B	MO	48(37)	3	4	48	45	37 ± 6	8	N. Puget Lowland,
GU-2B	M	48(37)	1	3	58	38	29 ± 3	4	Washington
GU-2B	M	42(37)	1	3	66	30	N.D.	N.D.	
GU-1A	M	238(37)	4	2	43	51	43 ± 10	3	
GU-1B	M	190(37)	6	5	67	22	37 ± 2	5	
GO-3A	MO	47(37)	3	3	26	62	45 ± 5	18	
GO-4A	MO	64	17	2	25	56	21 ± 1	4	
GO-4A	MO	59	14	2	27	57	N.D.	N.D.	
								N.D.	
TU-1A	NM	31	1	4	55	41	24 ± 2	4	Vashon till,
TU-2A	NM	33	2	3	80	16	N.D.	N.D.	N. Puget Lowland,
TU-2A	NM	35	2	3	77	19	N.D.	N.D.	Washington
TU-2A	NM	34	1	3	84	12	N.D.	N.D.	
TU-3A	NM	51	1	4	62	33	34 ± 5	5	

*M = marine, including brackish water; NM = non-marine; O = from oxidized outcrop.

[†] Sum of Na, K, Ca, and Mg extracted with NH_4^+, in meq/100 g of clay.

[§] Exchangable cations (NH_4^+) in mg/100 g of clay normalized to 100% (adjusted to CAT = 100).

[#] Mean boron content and standard deviation (ppm).

**N = number of replicate boron analyses.

[††] Numbers in parentheses are values corrected for Ca dissolved from $CaCO_3$ during extraction.

[§] N.D. = no data.

TABLE A3. PALEOECOLOGICAL DATA AND INFERRED PALEOENVIRONMENTS FOR DOMINANT EVERSON-AGE FORAMINIFERAL AND MICROFOSSIL SPECIES FROM THE NORTHERN PUGET LOWLAND*

Species	Distrib.[†]	Depth range	Salinity	Habitat; substrate
Foraminifera				
Elphidium clavatum (Cushman)	N,S	<30 m	Normal marine (35‰ to 15‰)	Bay and estuarine
Protelphidium orbiculare (Brady)[§]	S	<10 m	Brackish (30‰ to 18‰)	Intertidal, estuarine
Buccella frigida (Cushman) and *Buccella tenerrima* (Bandy)	S	10 to 50 m	Normal marine (35‰ to 20‰)	Bay and estuarine
Cassidulina teretis Tappan[§]	N	<15 m to upper bathyal zone	Normal marine to reduced	Shallow bay to open marine
Cassidulina islandica Nørvang*	N	40 to 70 m	Normal marine	Nearshore to open marine
Macrofossils				
Bivalves				
Nuculana fossa (Baird) and *Nuculana minuta* (Fabricius)	N	15 to 90 m	Normal marine	Offshore; mud and sand
Saxidomus giganteus (Deshayes)	S	Intertidal to 18 m	Normal marine to brackish	Bay and estuarine; sand and gravel
Mya truncata (Linne)	S	Intertidal to 64 m	Normal marine to brackish	Bay; mud and sand
Hiatella arctica (Linne)	S	Intertidal to 100 m	Normal marine to reduced	In mud or attached to rocks
Worms				
Serpula vermicularis (Linne)	S	Intertidal to 36 m	Normal marine to brackish	Bay and estuarine, on rocks
Barnacles				
Balanus sp.	S	Intertidal	Normal marine to brackish	Attached to rocks or shells

*From Balzarini, 1981.

[†] Distribution of species north (N) or south (S) of Lummi Peninsula.

[§] Species have been reported only from the Arctic. All other species range from the Arctic south into Puget Sound.

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