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ISOSTATIC EFFECTS OF THE LAST GLACIATION IN THE PUGET LOWLAND, WASHINGTON

By

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ABSTRACT

During the Vashon Stade of the late Pleistocene Fraser Glaciation, the Puget lobe of the Cordilleran Ice Sheet advanced into western Washington, inundating the Puget lowland between the Cascade Range and Olympic Mountains. During the systematic recession of the Puget lobe between about 14,000 and 13,000 years B. P., a complex system of proglacial lakes formed in the troughs of the Puget Sound region. Initially the lakes drained south via the Chehalis River to the Pacific Ocean, but lake drainage was later directed northward to the Strait of Juan de Fuca. Following deglaciation of the lowland about 13,000 years ago, marine waters invaded the areas formerly occupied by proglacial lakes.

The surfaces of the proglacial lakes (water planes) that formerly occupied the Puget lowland can be reconstructed from lake spillways that controlled the height of the lakes, and from outwash deltas built at the lake surfaces. The marine limit in the northern Puget lowland is also a nearly isochronous water plane, and can be reconstructed from the distribution of raised marine deltas and glacial-marine drift and of meltwater channels. The ancient marine and lacustrine water planes were determined independently, yet both indicate a regional gradual northward increase in the amount of postglacial deformation, reaching a maximum of about 140 m in the northern Puget lowland. The major cause of postglacial deformation in the Puget lowland was a return to isostatic equilibrium following deglaciation. At least one, and possibly three, significant departures from the regional pattern of uplift near Seattle were probably caused by Holocene tectonic warping and (or) possibly by variations in the rate of glacier retreat.

Comparison of the predicted and observed isostatic responses indicates that the substratum below the Puget lowland responded rapidly to the mass imbalance caused by ice-sheet glaciation. As much as 35 to 70 percent of isostatic equilibrium apparently was attained at the glacial maximum, which occurred less than several thousand years after glaciers invaded the Puget lowland. This inferred rapid response is supported by relatively low estimates for the effective viscosity of the substratum obtained from minimum uplift rates following deglaciation (3 cm/yr). The large amount and rapid rate of isostatic rebound in the Puget lowland within the last 13,000 years may relate to regional crustal structure, Quaternary faulting, and recent seismicity.

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INTRODUCTION

Much attention is currently being focused on the tectonic behavior of western Washington. Recent studies have attempted to characterize the nature and rate of regional uplift, as well as the pattern of recent faulting and seismicity (Crosson, 1973, 1977; Gower, 1978; Moen, 1971). A thorough understanding of Quaternary deformation in the Puget lowland is needed, particularly where it affects the rates of sea-level change and the potential for causing damaging earthquakes. Such an understanding requires that the character, rates, and causes of postglacial deformation be determined as accurately as possible.

Glacial isostatic rebound has been a major cause of postglacial deformation in the Puget lowland. Broad uplift occurred in response to loading and unloading of the region by the advance and retreat of the Cordilleran Ice Sheet. Isostatic uplift may also have set up stresses large enough to have caused displacement or warping along major tectonic structures and along small faults in the surficial sediments. It is therefore imperative that studies of Quaternary tectonism take glacial isostatic effects into consideration.

The primary objective of this investigation is to document the pattern and rates of postglacial isostatic deformation in the Puget lowland. A further objective includes determining whether temporal or spatial irregularities in the pattern of isostatic recovery can be associated with known tectonic or seismic zones. Also, can the study of isostatic response provide insights into the nature of the underlying crust and upper mantle? Finally, can the inferred pattern of isostatic rebound provide new data on the mode of ice retreat during deglaciation?

The Puget lowland is a physiographically distinct region that includes Puget Sound and the lowland terrain between the Cascade Range and Olympic Mountains south of the Strait of Juan de Fuca (fig. 1). For the purpose of this study, the Puget lowland is considered to extend northward to the Stillaguamish River and to include Camano Island, southern Whidbey Island, and the Quimper and Miller Peninsulas east of Sequim. The study area is bordered on the south by the drainage divide between the Puget Sound and Chehalis River watersheds.

During late Pleistocene time lobes of the Cordilleran Ice Sheet that originated largely in the Coast Range of Canada repeatedly advanced southward into western Washington (Crandell and others, 1958). Near the San Juan Islands the south-flowing ice divided into two separate lobes. The Juan de Fuca Lobe advanced west to the Pacific Ocean through the Strait of Juan de Fuca. The portion of the ice sheet that extended south of the Strait of Juan de Fuca and the Stillaguamish River is referred to as the Puget lobe. It advanced southward along the length of the Puget lowland and terminated against the Black Hills south of Olympia (fig. 1).

During the Evans Creek Stade of the late Pleistocene Fraser Glaciation, the Puget lobe did not exist, but alpine glaciers that drained the adjoining Cascade

Range and Olympic Mountains reached their maximum extent (Armstrong and others, 1965; table 1). During the Vashon Stade the Puget lobe advanced southward across the Puget lowland, and dammed north-draining streams, creating a large proglacial lake that drained south via the Chehalis Valley to the Pacific Ocean (fig. 1). At its maximum extent, the Puget lobe extended across the Puget lowland between the Cascade and Olympic mountains. During the various stages of glacial retreat a complicated sequence of proglacial lakes evolved (Bretz, 1913); outwash transported along ice-marginal meltwater channels built large deltas into these lakes. According to Armstrong and others (1965), the Vashon Stade ended and the Everson Interstade began when the Puget lobe receded far enough north so that the sea reentered the Puget lowland (table 1). To comply with proper stratigraphic usage, the Everson Interstade should be extended forward in time to include the rapid recession of the Puget lobe from the southern Puget lowland. To avoid confusion with the existing literature, however, the more general terms Vashon recession and Everson interval will be used to avoid the restrictions imposed by use of the terms stade and interstade.

The age of the Vashon Stade in the Puget lowland is based on bracketing radiocarbon dates on pre- and postglacial sediments near Seattle (table 2; figs. 2 and 3). Maximum limiting dates of 15,000 ± 400 ^{14}C yr B.P. (W-1227) and 15,100 ± 600 ^{14}C yr B.P. (W-1305) (Mullineaux and others, 1965) indicate that the advancing Puget lobe had not yet reached the latitude of Seattle by about 15,000 yr B.P. A date of 13,570 ± 130 ^{14}C yr B.P. (UW-35) from near the eastern margin of the Puget lobe is interpreted as indicating that the Snoqualmie Valley was still dammed by Vashon ice at that time (Porter, 1976). A date of 13,650 + 550 ^{14}C yr B.P. (L-346a) for basal peat from Lake Washington indicates that the Vashon glacier had retreated to a position north of Seattle by that time (Rigg and Gould, 1957). Numerous dates for glacial-marine sediments in the northern Puget lowland between 12,500 and 13,100 yr B.P. (Easterbrook, 1968) indicate that rapid recession of the Puget lobe continued well north of the Seattle area during the Everson interval (figs. 2 and 3). No major stillstands or readvances during the Vashon recession have yet been documented in the Puget lowland.

Studies of glacial isostasy in the Puget lowland were initiated by Bretz (1913) who described uplifted marine sediments throughout the Puget Sound region. Sediments from at least some of the localities in the southern Puget lowland that Bretz described as marine in origin were later interpreted to be archeological shell middens (S.C. Porter, oral comm., 1979). From his detailed study of the Vashon recessional lake sequence, Bretz concluded that <u>no</u> significant differential isostatic rebound occurred in the Puget lowland. Subsequent geologic studies of Vashon glacier recession by Crandell (1963), Anderson (1965), Curran (1965), Rosengreen (1965), Noble and Wallace (1966), Knoll (1967), and Mullineaux (1970) did not report significant amounts of isostatic rebound in the south-central and east-central Puget lowland.

Glacial-marine sediments with unreworked shells were described on Whidbey Island (Easterbrook, 1968). These deposits range in age from about 12,000 to 13,000 yr B.P. and indicate that the northern Puget lowland was depressed isostatically below sea level at that time. Terrestrial peat deposits on Whidbey Island which have been submerged by rising sea level during the Holocene indicate

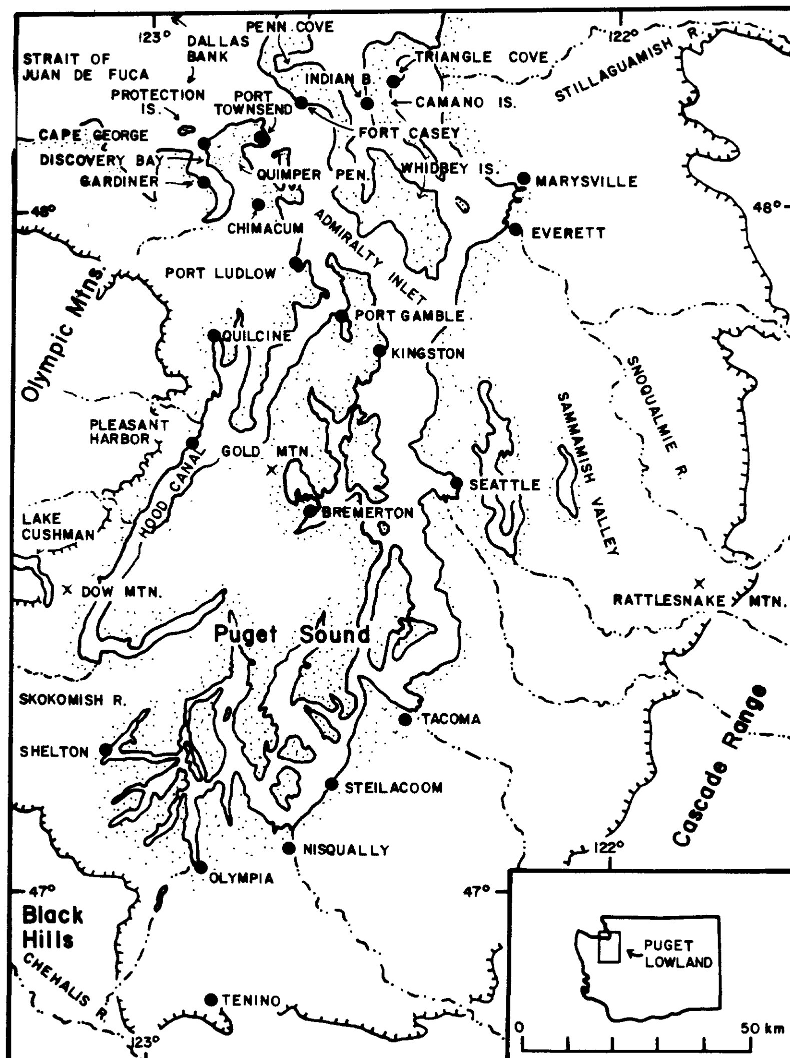

Figure 1.--Map of the Puget lowland, Washington, showing localities mentioned in the text and the maximum extent of the Puget lobe (hachured line) during the Vashon Stade.

TABLE 1.-- LATE QUATERNARY GEOLOGIC-CLIMATE UNITS FOR WESTERN WASHINGTON (MODIFIED FROM ARMSTRONG AND OTHERS 1965).

Glaciation or Interglaciation	Stade or Interstade
Holocene Interglaciation	
Fraser Glaciation	Sumas Stade Everson Interstade Vashon Stade Unnamed Interstade Evans Creek Stade
Olympia Interglaciation	
Salmon Springs Glaciation	

that isostatic recovery in the northern Puget lowland was largely complete by about 6000 years ago (D.D. Biederman, unpublished report, University of Washington, 1967).

North of the Puget lowland in Whatcom County, Easterbrook (1963 p. 1480) presented evidence that he interpreted as indicating that "...250 feet of emergence, 500-700 feet of submergence, and emergence of 500-700 feet occurred in approximately 1000-1500 years following the Vashon glaciation." Mathews and others (1970), working in the Fraser lowland of southwestern British Columbia, also found evidence for post-Vashon resubmergence of hundreds of feet (about 100 m). Neither of the authors satisfactorily explained the inferred extremely large amounts and rapid rates of emergence, submergence, and re-emergence. Because these inferred fluctuations of sea level are based on stratigraphic interpretations, the pattern and rates of postglacial isostatic recovery north of the Puget lowland have been controversial.

TABLE 2.-- RADIOCARBON DATES PERTAINING TO GLACIATION OF THE PUGET LOWLAND DURING THE VASHON STADE.

Lab Number	Age ^{14}C yr B.P.	Location	Material
		Maximum Limiting Dates	
W-1892	18,920 ± 600	North Seattle	wood
W-1186	18,100 ± 700	Discovery Park, Seattle	wood
I-2282	18,000 ± 400	North Marrowstone Island	peat
W-2125	16,070 ± 600	Bellevue	peat
W-1035	15,100 ± 600	Seattle	wood
W-1227	15,000 ± 400	Seattle	wood
		Minimum Limiting Dates	
L-330	14,000 ± 980	Lake Washington	peat (?)
L-346a	13,650 ± 550	Lake Washington	peat
UW-35	13,570 ± 130	Snoqualmie Valley	wood
UW-32	13,100 ± 170	Penn Cove	marine shells
W-398	12,900 ± 330	Sedro Wooley	basal peat
GS-945	12,720 ± 160	Palmer, B.C.	peat & marine shells
UW-146a	12,700 ± 160	Simpson Lake (Shelton)	wood
GS-246	12,660 ± 160	Victoria, B.C.	marine shells
UW-147	12,620 ± 150	Simpson Lake (Shelton)	wood
I-1881	12,600 ± 190	Orcas Island	marine shells
I-1079	12,535 ± 300	Point Partridge	marine shells
GS-390	12,440 ± 230	Brentwood Bay, B.C.	marine shells
UW-146b	12,430 ± 160	Simpson Lake (Shelton)	wood
I-2286	12,400 ± 190	Fidalgo Island	marine shells
I-1469	12,350 ± 330	Strait of Juan de Fuca	marine shells
UW-8	12,300 ± 200	Skagit County	peat
I-2154	12,300 ± 180	Whidbey Island	marine shells

TABLE 2.-- RADIOCARBON DATES PERTAINING TO GLACIATION OF THE PUGET LOWLAND DURING THE VASHON STADE -- Continued.

Lab Number	Age ^{14}C yr B.P.	Location	Material
I-1470	12,160 ± 290	San Juan Island	marine shells
GSC-1114	12,100 ± 160	Victoria, B.C.	marine shells
L-269a	11,900 ± 350	Snoqualmie Valley	peat
I-2156	11,900 ± 170	San Juan Island	marine shells
I-1448	11,850 ± 240	Penn Cove	marine shells
UW-394	11,500 ± 300	Olympia	peat
GSC-1131	11,500 ± 160	Victoria, B.C.	organic silt
GSC-945	11,400 ± 190	Esquimalt, B.C.	organic silt
GSC-1142	11,200 ± 190	Victoria, B.C.	organic silt
GSC-1130	11,200 ± 170	Victoria, B.C.	peat

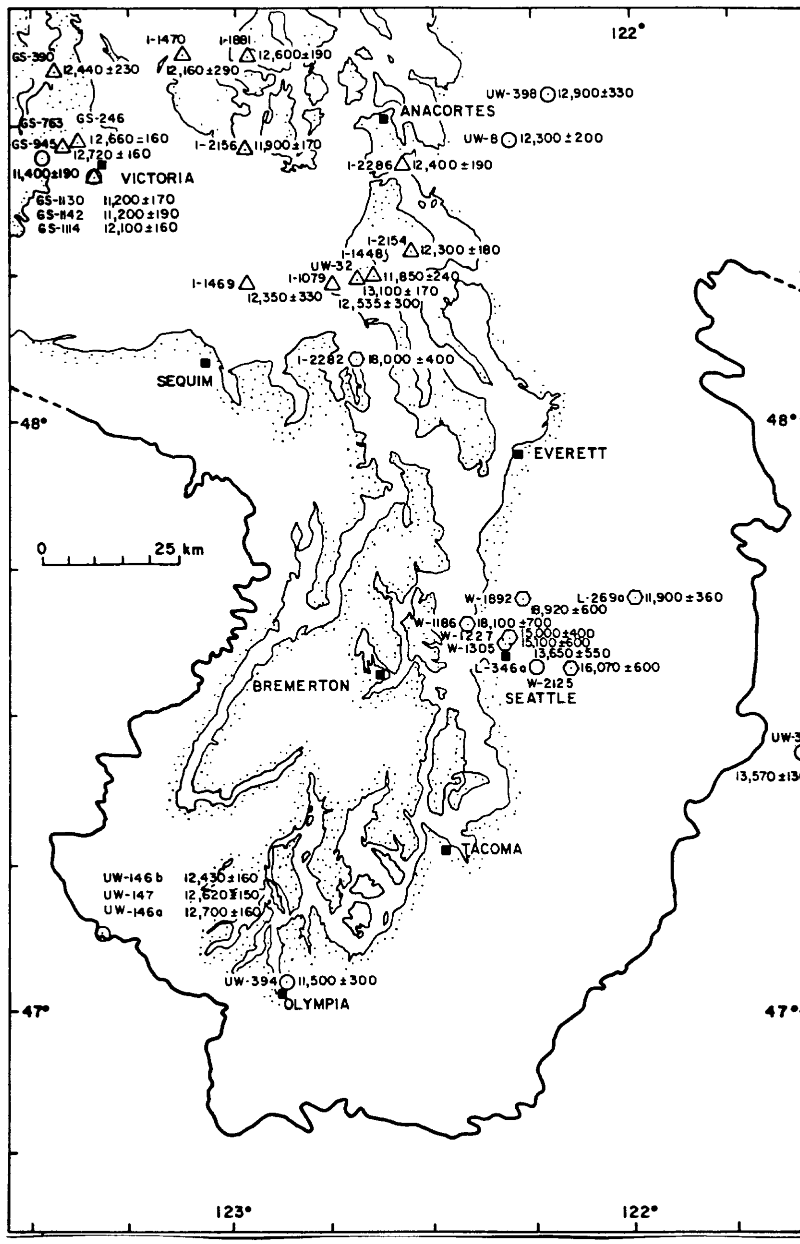

Figure 2.--Radiocarbon dates pertaining to Fraser-age glaciation of the Puget lowland, Washington. Locations of maximum limiting dates derived from wood and peat (hexagons), and minimum limiting dates derived from wood and peat (circles) and marine shells (triangles) shown. Heavy solid line indicates maximum extent of the Puget lobe during the Vashon Stade.

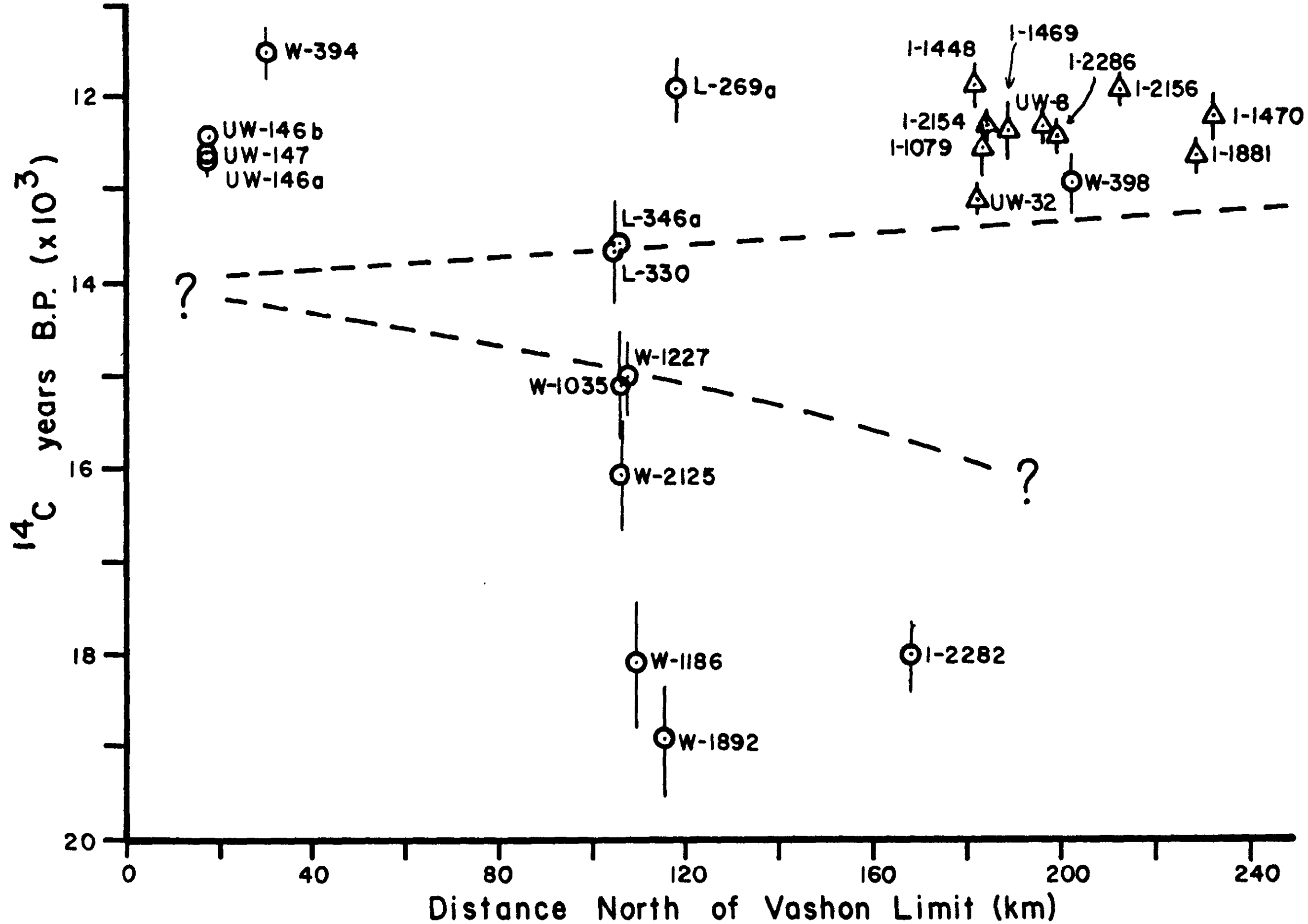

Figure 3.--Diagram showing the maximum and minimum ages for advance and retreat of the Puget lobe during the Vashon Stade and Everson Interstade. Triangles represent dates on marine shells. Circles represent dates on terrestrial organic material. Refer to Table 2 for descriptions of dates. Dates are projected to axial flowline "E" of figure 4.

RECONSTRUCTION OF THE PUGET LOBE DURING THE VASHON MAXIMUM

Reconstructions of the geometry and history of the Puget lobe during the Vashon stade are essential to the study of late Quaternary isostasy in the Puget lowland. Changes in the distribution of mass over the region, which set up an isostatic disequilibrium, were caused almost entirely by the invasion of the Cordilleran Ice Sheet. Therefore, the Puget lobe must be accurately reconstructed so that the predicted and observed isostatic responses can be compared and interpreted. A reconstruction of the terminal positions and direction of flow of the Puget lobe at its Vashon maximum is essential to an understanding of the shape of the glacier margin during retreat.

Three critical parameters necessary to reconstruct accurately the geometry of the Puget lobe are the maximum horizontal extent of the ice, its altitude against the confining mountain ranges, and the flowlines at the base of the glacier. The nature and distribution of rock and sediment types at the glacier bed may also play a significant role in geometric reconstruction of the ice sheet. The maximum extent of the Puget lobe is known accurately because the limit of Vashon drift has been mapped extensively (fig. 4). Within the entire Puget lowland, only a small part of Rattlesnake Mountain west of Seattle and Dow Mountain near Lake Cushman apparently were exposed as nunataks above the ice surface during the maximum stand. The altitude of the ice limit along the Cascade Range and Olympic Mountains has been determined by various investigators from the height of glacial drift, erratics, and glacier-erosional features (fig. 4). Glacier-altitude data from the literature were supplemented by interpretations of the maximum altitude of meltwater channels and erosion features of Vashon age.

The glacier flowlines can be inferred from the orientation of landforms that indicate the direction of ice flow at the base of the glacier. Throughout much of the Puget lowland, bedrock and pre-Vashon sediments were molded and streamlined by the eroding glacier. The Vashon till plain commonly exhibits broad areas of well-developed drumlins and other elongate hills. These features were mapped in detail in order to reconstruct accurately the flowlines at the base of the Puget lobe. Each feature was mapped originally at the largest available map scale, then compiled at a scale of 1:250,000. Glacial striations and till fabric at various localities (Curran, 1965; Rosengreen, 1965) are subparallel to these streamlined features. These four types of ice-flow indicators are approximately the same age, and provide excellent data on the direction of flow at the base of the Puget lobe. Continuous flowlines can be drawn by interpolation between individual flow features (fig. 4). If it is assumed that almost all the glacier flow takes place at the base of the glacier whether by basal slip or shear in the lowest layers (Weertman, 1964), then the flowlines at the glacier bed should parallel those at the ice surface.

The flowline reconstruction indicates that the terminus of the Puget lobe divided into two sublobes (fig. 4). An eastern sublobe, deflected westward by the Cascade foothills, terminated near Tenino south of Olympia. A western sublobe terminated against the Black Hills several kilometers south of Shelton. The

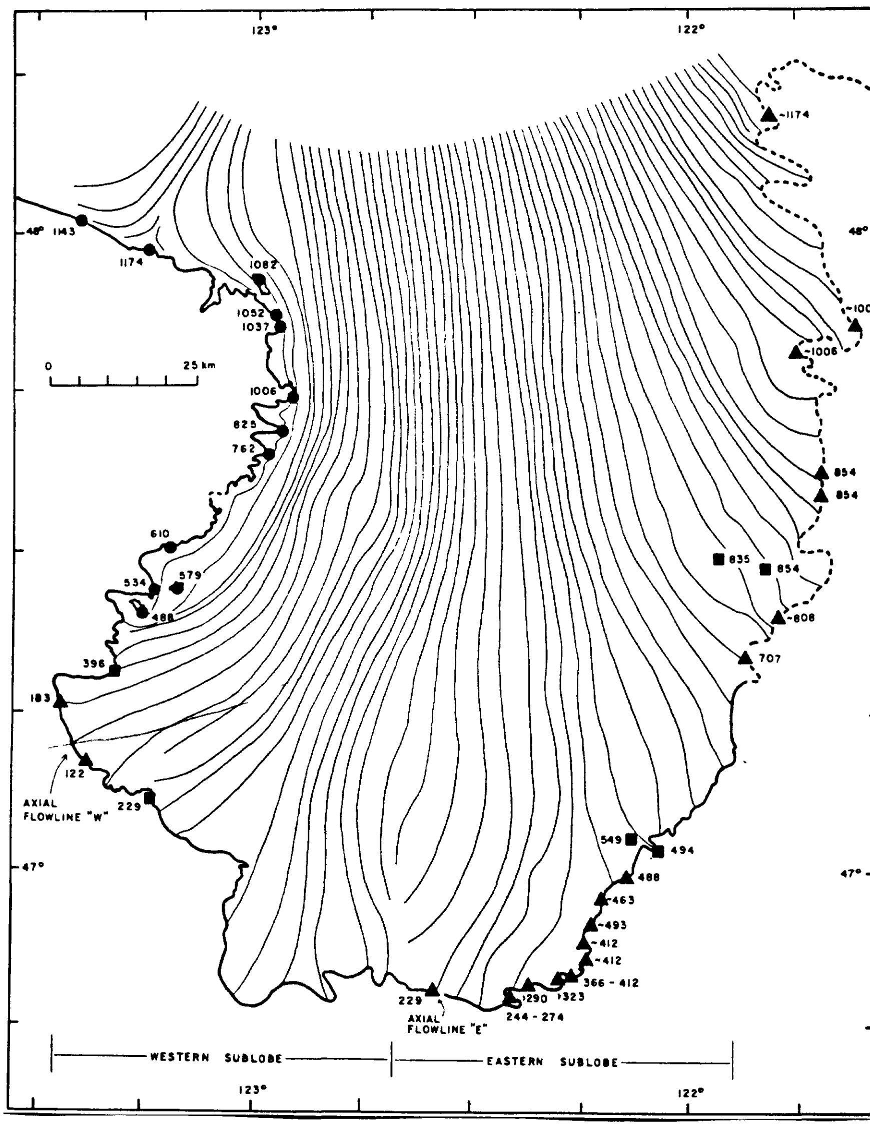

Figure 4.--Map showing the maximum extent of the Puget lobe and the direction of flow at the base of the glacier, as inferred from ice-flow indicators. Numbers along the margin of the lobe indicate the maximum altitude (m) of the ice sheet against the Cascade and Olympic mountains, as recorded by the height of glacial erratics, ice-marginal channels, and glacially abraded bedrock. Squares indicate values cited in the literature (Anderson, 1965; Curran, 1965; Knoll, 1967; Rosengreen, 1965; Crandell, 1963; and Carson, 1970). Circles represent localities where the Vashon ice limit has been determined by William A. Long (written comm., 1978). Triangles represent altitudes of the Vashon limit inferred from geomorphic features. Flowlines "E" and "W" show locations of longitudinal profiles drawn in figure 3.

presence of sublobes prohibits projection of glacier-altitude data from the Cascade Range and Olympic Mountains to the same flowline. Longitudinal profiles for both sublobes were drawn by projecting the glacier-altitude data from the ice sheet margins to selected axial flowlines. Projection must be perpendicular to the inferred regional ice-surface gradient which is parallel to the local flowlines. The resulting inferred longitudinal profiles for the Puget lobe are shown in fig. 5. Differences in altitude between the inferred profiles and the actual data points are due largely to simplifications in the method of reconstruction and to probable local variations in the surface height of the ice sheet. Local variations may have been caused both by bedrock relief at depth (Nye, 1959) and by variations in the rate and type of soft-sediment deformation at the glacier bed (Boulton, 1978). The profile of the ice against the Olympic Mountains shows a major change in slope between Quilcine and Pleasant Harbor. This steepening may have been caused by the interference of Gold Mountain which projects more than 400 m above the general level of the Puget lowland in this vicinity. No similar steepening was observed east of Seattle, perhaps because the large bedrock irregularities there lie well east of the line of projection. In spite of the difficulties and assumptions involved, both Olympic and Cascade longitudinal profiles are closely similar along much of their length.

Malaspina Glacier, at the northeastern border of the Gulf of Alaska, is a reasonable modern analog of the Puget lobe during the Vashon Stade. The Malaspina is a broad piedmont glacier that lies at the base of the St. Elias Mountains in a cool maritime climate. Similar environmental conditions probably prevailed for the Puget lobe during the maximum and recessional phases of the Vashon Stade (Heusser, 1977). A longitudinal profile for the Malaspina Glacier was constructed by drawing a central flowline perpendicular to the glacier contours (fig. 5). In its terminal portion the Malaspina Glacier longitudinal profile agrees very closely with the reconstructed profiles for the Puget lobe. The abrupt increase in gradient of the Malaspina profile 40 km upglacier from its terminus occurs at the mountain front.

Construction of a theoretical longitudinal profile for the Puget lobe depends on several fundamental assumptions: (1) that ice in temperate glaciers can be approximated as deforming plastically, (2) that all glacier movement takes place at the base of the glacier by basal slip and (or) shearing in the lowest layers, (3) that the glacier is moving over a rigid horizontal bed, (4) that the mass balance is equal over the ice surface, and (5) that the ice sheet is in a steady state. General support for the first two assumptions is found in empirical field measurements of modern glaciers and laboratory experiments of ice deformation (Kamb, 1964). The Puget lobe did not advance over a uniform and level bed, but the regional gradient of the drift plain near the center of the lowland is nearly horizontal. The theoretical equilibrium longitudinal profile is similar to, but lies slightly above, both the reconstructed profiles of the Puget lobe and the modern profile of the Malaspina Glacier (fig. 5). This parallelism suggests that the assumptions mentioned above are largely valid, and that the Puget lobe may have been nearly in an equilibrium state at its maximum extent.

The altitude of the Puget lobe over its entire surface at the maximum Vashon extent can be approximated by contouring points of equal elevation of the ice

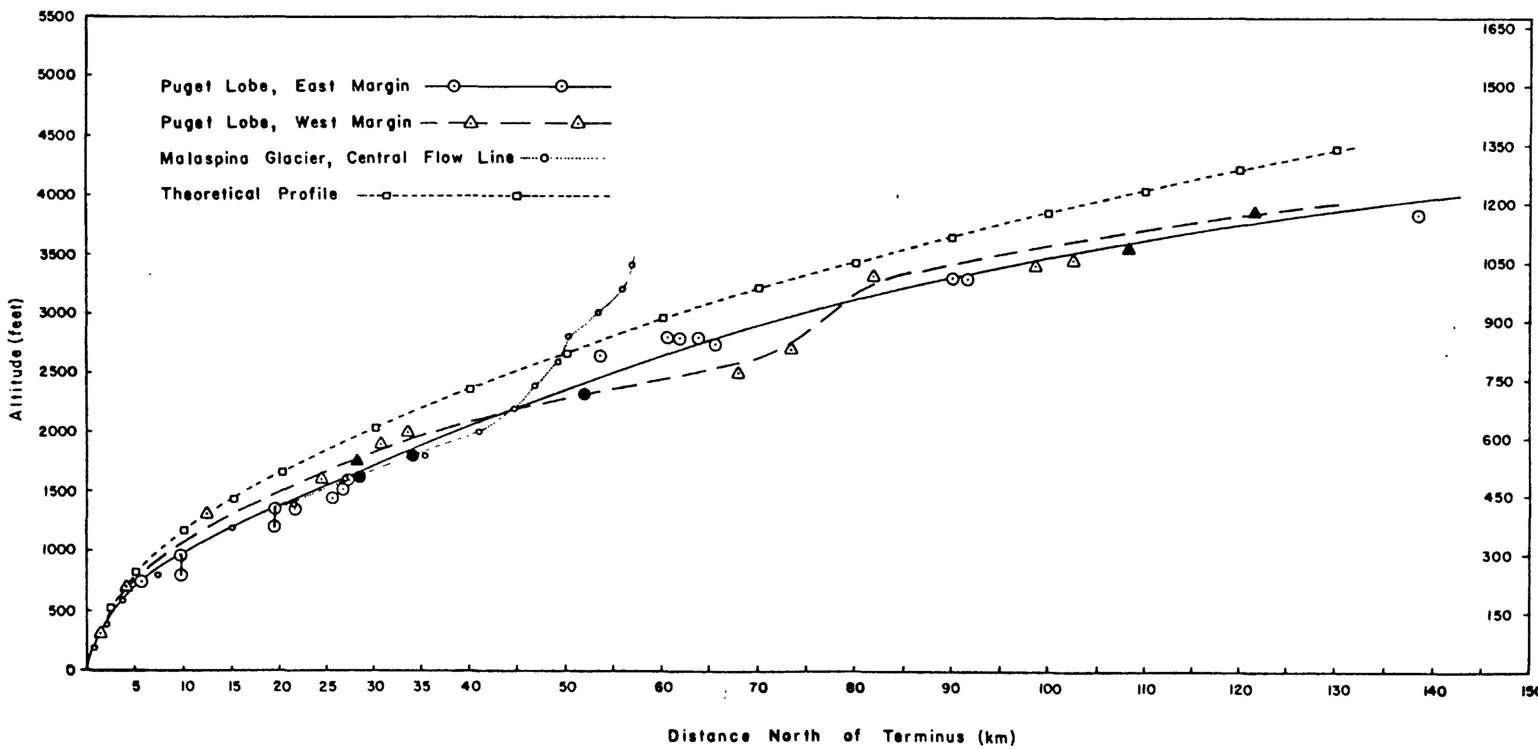

Figure 5.--Longitudinal profiles of the Puget lobe compared to a theoretical equilibrium glacier profile and to the present longitudinal profile of the Malaspina Glacier, Alaska (constructed from U.S.G.S. topographical maps of the Yakutat and Mount St. Elias quadrangles, scale 1:250,000). Solid points are more reliable than open points. Vertical bars connecting points form the east margin of the Cascades indicate a range in altitude. Refer to Figure 4 for location of profiles of the Puget lobe.

(fig. 6). Control points for the contour map are the inferred altitudes of the Puget lobe along its perimeter, based on geological evidence (fig. 4), and the reconstructed altitudes along the axial flowlines (fig. 5). Interpolation between points primarily was done normal to the reconstructed ice-surface gradient. Contouring was difficult in the northwestern portion of the lobe because eastward projections of glacier altitudes from the Olympic Mountains plotted north of equivalent points from the Cascade Range front. This discrepancy may have been caused by peculiar flow conditions in the northwest corner of the lobe where the basal ice was flowing in a slightly different direction from ice at the surface. These conditions may have been initially caused by divergent ice flow at the junction between the west-flowing Juan de Fuca lobe and the south-flowing Puget lobe. Gold Mountain may have caused oversteepening of the glacier surface in this area, and may also have caused variations in ice-flow directions as well. If the contour map of the Puget lobe at its Vashon culmination is accurate, the ice sheet reached altitudes of 1265 m, 1035 m, 770 m, and 465 m above present sea level over Port Townsend, Seattle, Tacoma, and Olympia, respectively.

The reconstructed contour map indicates that the Puget lobe extended as high as 1200 m against the bordering mountain fronts, causing large lakes to be impounded in the major river valleys. Meltwater generated along the margins of the lobe merged with meltwater from the Cascade and Olympic valley glaciers and flowed south along a series of successively lower ice-margin spillways. Near the glacier terminus an extensive system of valley trains and outwash plains were built, and outwash was deposited in the Chehalis Valley along the main meltwater route to the Pacific Ocean.

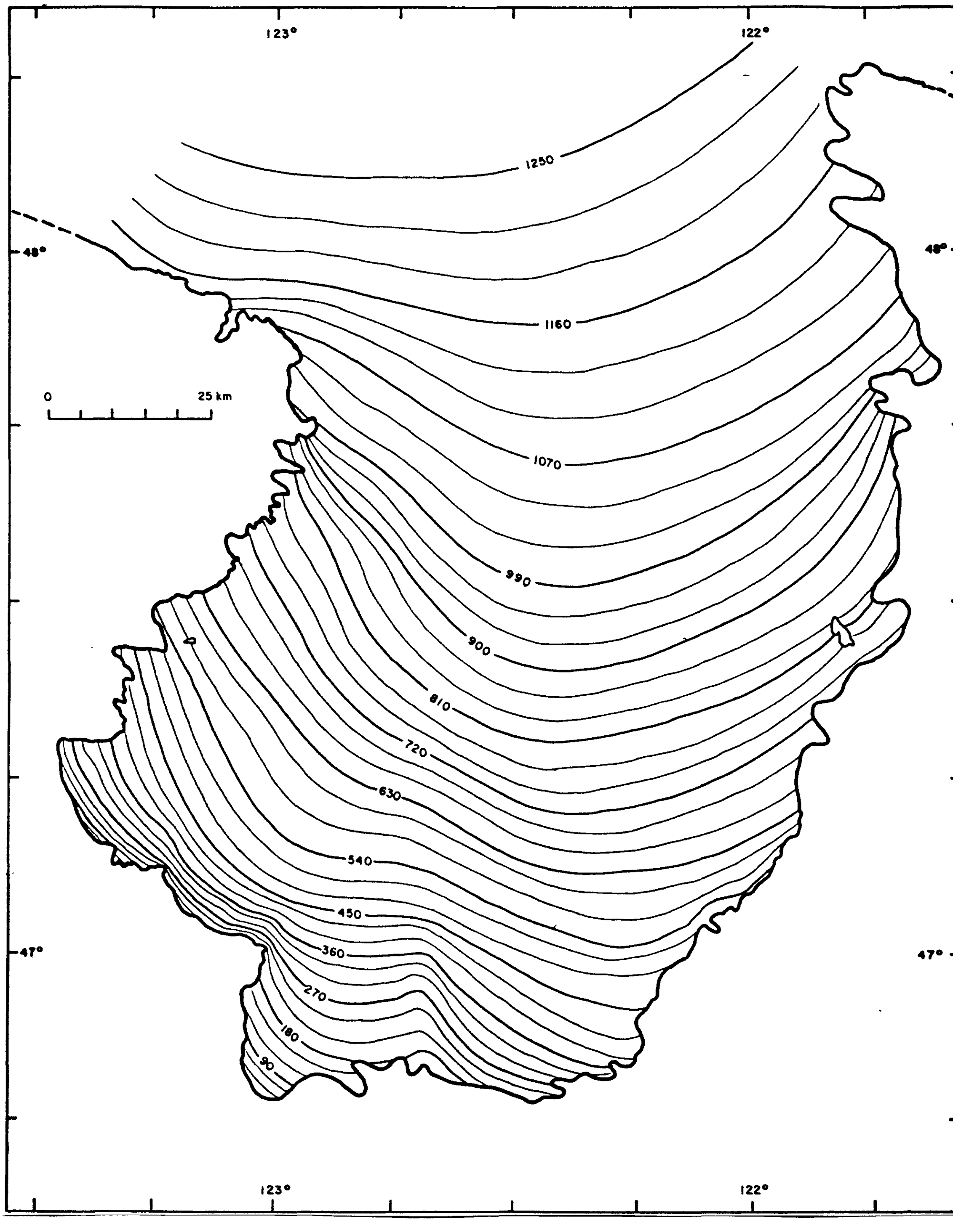

Figure 6.--Contour map of the reconstructed ice surface of the Puget lobe during the Vashon Stade. Altitude in meters; contour spacing, 30 m.

DEGLACIATION OF THE PUGET LOWLAND

The isostatic response of an area that is undergoing deglaciation is critically dependent on the rate and character of glacier retreat. Broecker (1966) has shown that the shape of the isostatic uplift curve in the Great Lakes region was controlled largely by the rate of retreat of the Laurentide Ice Sheet. Variations in the retreat rate of the Puget lobe would also have caused significant changes in the pattern of isostatic recovery.

The successive configurations of the retreating ice margin controlled the evolution of the proglacial lakes in the Puget lowland. Because the determination of isostatic deformation in the Puget lowland is based largely on the features created at the shorelines of these lakes, the configuration of the retreating ice margin must be determined as accurately as possible.

Pattern of Ice Retreat

With the exception of Bretz (1913), previous studies of Vashon glacier recession in the Puget lowland have been concerned largely with retreat in relatively local areas. This investigation, however, required that the recessional configurations of the Puget lobe be mapped on a regional scale. To meet this requirement four different types of data were integrated during preparation of a map showing the regional pattern of glacier retreat: (1) previous studies of retreat in local areas, (2) regional mapping of the recessional sediments and ice-margin meltwater channels, (3) regional construction of theoretical ice-margin configurations based on a simple model of glacier retreat perpendicular to the ice front, and (4) new geologic field studies in the northwest Puget lowland. All data were compiled at the same map scale (1:250,000) so that the different types of data could be compared.

Previous Studies

Terminal zone

During initial retreat of the Puget lobe broad areas of hummocky dead-ice terrain formed near the glacier terminus (Bretz, 1913; Noble and Wallace, 1966; fig. 7). Some of the extensive ice-cored moraine deposits in the region, which included ice-contact stratified drift and kettled outwash terraces, did not melt out completely for thousands of years (Porter and Carson, 1971). Meltwater streams apparently had high discharge at this time, as indicated by huge areas of coarse outwash gravel that extended downvalley from the retreating ice margin and merged southward to join the Chehalis meltwater stream. The systematic uncovering of lower meltwater channels during initial deglaciation resulted in the

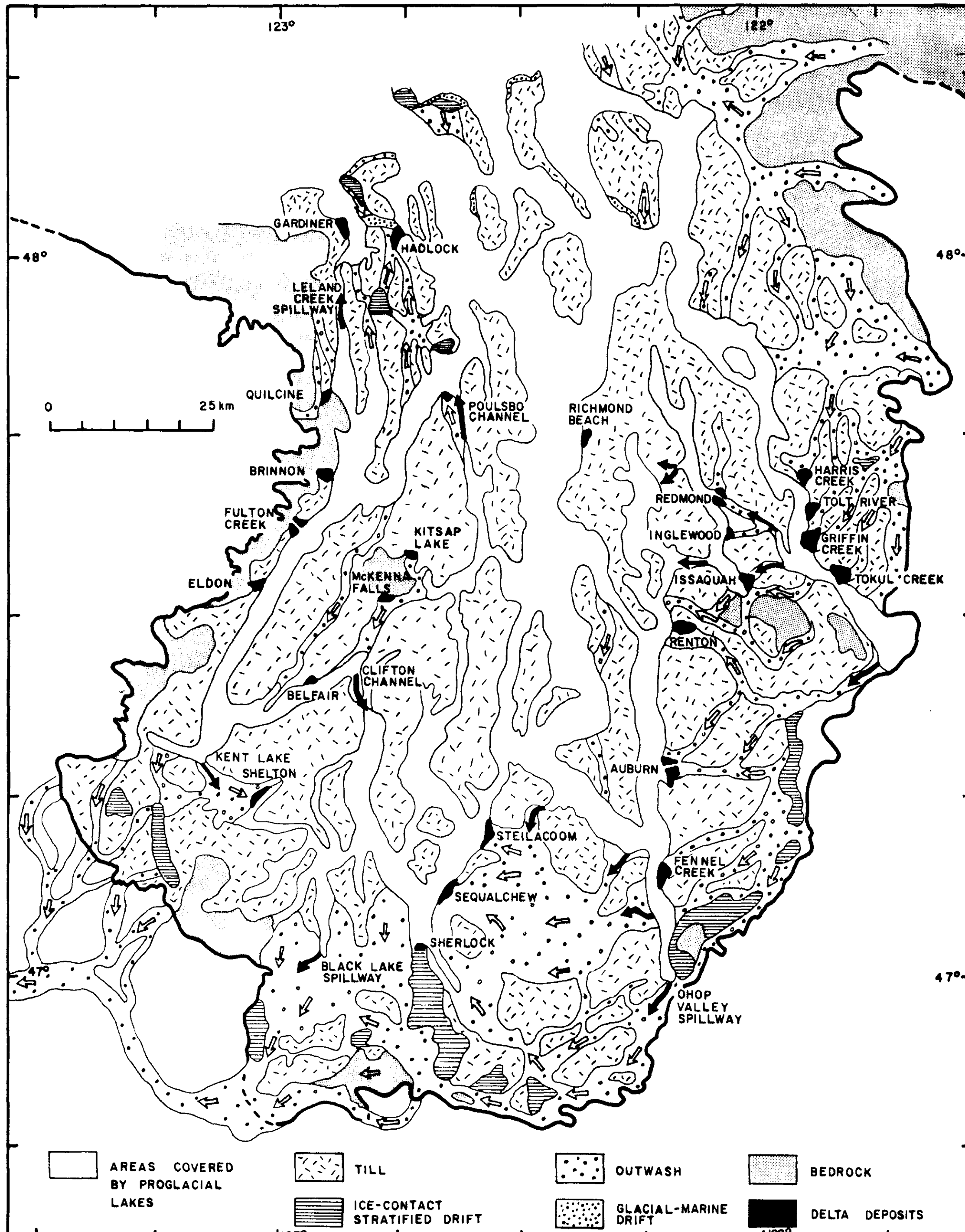

Figure 7.--Generalized geologic map of the Puget lowland showing landforms and sediments related to retreat of the Puget lobe. Principal proglacial lake spillways and deltas are named. Open arrows indicate orientation of inferred meltwater channels. Solid arrows indicate direction of drainage from lake spillways.

formation of extensive gravel terraces in the southern Puget lowland, and indicate that retreat was accomplished largely by systematic northward recession (Noble and Wallace, 1966).

Sublobes formed at the ice terminus during retreat. A pronounced recessional moraine near Nisqually apparently formed in an interlobate position between a north-retreating west sublobe and an east-retreating east sublobe (Bretz, 1913). Other areas of stagnant-ice terrain probably formed where ice tongues or sublobes became separated from the retreating glacier and were buried by aggrading outwash.

Southwest Puget Lowland

Deglaciation of the southwest portion of the Puget lowland was apparently also systematic (Carson, 1970). During glacier recession the ice margin had a nearly north-south orientation, normal to inferred glacier flowlines. Several prominent slightly arcuate ridges of hummocky ice-contact stratified drift lie just beyond large expanses of fluted till plain. These ridges are interpreted as moraines and may represent brief stillstands of the retreating glacier margin (Carson, 1979).

Southeast Puget Lowland

Deglaciation in the southeast Puget lowland is the best documented (Crandell, 1963; Mullineaux, 1970). Large expanses of kettled terraces that occur along the southeast margin of the Puget lowland appear to have formed largely as kame terraces. They apparently formed where gravel-laden meltwater streams were partially confined by the Puget lobe during initial retreat. On the upland drift plain the retreating margin of the Puget lobe had a general northeast orientation during systematic northwest recession. Locally, ice stagnated on the uplands as it thinned, resulting in formation of kames and eskers (Waldron, 1961, 1962). Most troughs of the southeast Puget lowland were occupied by proglacial lakes at this time. Blocks of stagnant ice in the troughs were partially buried by lake sediments during glacier recession (Crandell, 1963; Mullineaux, 1970; Waldron, 1961).

East-central Puget Lowland

Retreat of the Puget lobe in the east-central Puget lowland was studied by Mackin (1941), Anderson (1965), Curran (1965), Rosengreen (1965), and Knoll (1967). The progressive westward decrease in altitude of ice-margin meltwater channels near the Cascade mountain front east of Seattle indicates that the eastern margin of the Puget lobe retreated westward as it thinned near the mountain front. West of the range front on the drift plain, the lobe apparently maintained a northeast-southwest orientation during retreat, perpendicular to the direction of ice flow. Melting of the ice terminus there was accompanied by thinning and progressive northward retreat, but areas of ice-contact stratified drift formed where the ice stagnated locally. During retreat, large outwash deltas were built into proglacial lakes that occupied the present river valleys.

Regional Distribution of Glacial Features

Glacier-recessional sediments

Surficial geologic maps at scales of 1:24,000 and 1:63,360 cover almost all of the Puget lowland (Anderson, 1965; Birdseye, 1976; Carson, 1970, 1976, 1976b; Crandell, 1963; Curran, 1965; Easterbrook, 1968; Frisken, 1965; Gayer, 1977; Hanson, 1976; Hirsch, 1975; Knoll, 1967; Leisch and others, 1963; Luzier, 1969; Molenaar and Garling, 1965; Mullineaux, 1965a, 1965b, 1965c; Newcomb, 1952; Robinson, 1938; Rosengreen, 1965; Sceva, 1957; Todd, 1939; Vine, 1969; Walters and Kimmel, 1968; Waldron, 1961, 1962, 1967; Williams, 1971). Mapped deposits that are related to Vashon ice recession include lodgment till, ablation till, ice-contact stratified drift, outwash, lake-bottom sediments, and deltaic sediments (fig. 7). All of the existing maps of Vashon recessional deposits were compiled at a scale of 1:250,000 with the expectation that regional patterns would emerge which might be used to reconstruct the shape of the retreating ice margin at successive stages. However, deposits characteristic of ice margins, such as broad zones of ice-contact stratified drift or outwash-heads, were not continuous for more than a few kilometers north of the terminal zone; hence, no regional configurations could be inferred. This lack of continuous ice-marginal features suggests that no regionally significant stillstands of the Puget lobe occurred between the southern drift border and Penn Cove, Whidbey Island.

Ice-marginal channels

The upland drift plain in much of the Puget lowland consists of numerous closely spaced elongate hills that give the till plain a fluted appearance. They generally are 5-20 m high and less than 2 km long, but reach heights of about 50 m and lengths of 5 km. These linear hills commonly are completely cut through in places by small meltwater channels 50-500 m wide. Many channels cut across several ridges and are as much as several kilometers long. These relations indicate that many of the channels were probably superimposed on the fluted till plain by ice-marginal meltwater streams. They provide widespread erosional evidence for the shape of the ice margin during glacier recession. Individual channels were mapped first at larger scales, then compiled at a scale of 1:250,000.

Ice-marginal erosional channels are very common north of the Vashon drift border and south of Bremerton. In this area local ice-marginal positions inferred from the glacial deposits closely parallel ice-marginal positions inferred from the pattern of meltwater channels. These two types of data are mutually supporting, and provide good regional evidence for the shape of the glacier margin during retreat (fig. 8).

No evidence for the recessional configuration of the retreating Vashon glacier has yet been found in a large, wedge-shaped area between Quilcine, Seattle, and Fulton Creek (fig. 8). The absence of significant recessional deposits or ice-marginal channels within this broad region suggests that either the glacier receded too quickly for significant marginal features to form, or that the ice stagnated over a relatively large area and melted rapidly.

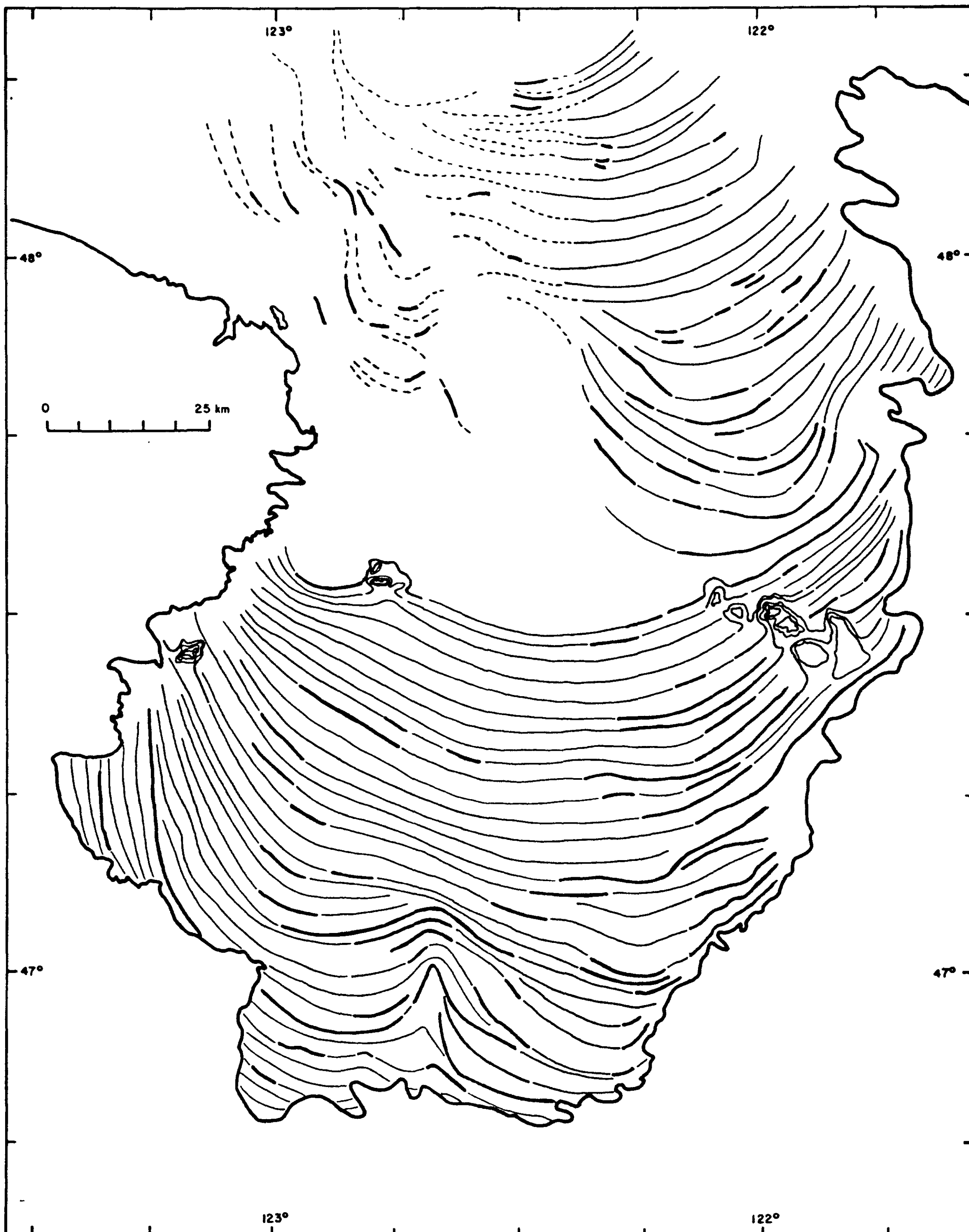

Figure 8.--Map of the Puget lobe showing successive configurations of the ice margin during glacial retreat. Heavy lines were drawn from geologic data. Continuous light lines were drawn from a theoretical reconstruction of ice-marginal positions as inferred from directions of meltwater flow. Concentric areas depict uncovering of nunataks.

Theoretical Ice-Margin Configurations.

Observations of modern glaciers and glaciologic considerations indicate that the fronts of glaciers melt back nearly perpendicular to the flowlines. Under ideal conditions, where ice along each reconstructed flowline melted back at the same rate, the configuration of the retreating ice margin would be everywhere perpendicular to flowlines. If these theoretical conditions are assumed for Vashon recession, a map of the inferred configurations of the Puget lobe during retreat can be drawn. These theoretical reconstructions can then be compared with geologic evidence for the shape of the ice margin at different stages of retreat.

In the southern and northeastern parts of the Puget lowland the ice-margin configurations inferred from geologic field data are strongly supported by those determined theoretically; theoretical ice-margin reconstructions were used in areas where geologic data were lacking (fig. 8). In the northwestern part of the lowland, however, the geologically reconstructed ice margins were much different from those determined theoretically. A north-draining lake spillway and ice-marginal meltwater channels near the head of Discovery Bay (fig. 9) indicate that this area must have been free of ice earlier than areas farther south. If systematic retreat perpendicular to ice-flow directions is assumed, this clearly would be impossible. Some factor other than progressive northward retreat apparently caused rapid glacier recession in the Strait of Juan de Fuca.

Glaciers that calve into tidewater are known to retreat rapidly due to the relatively high temperatures of marine water and to tidal effects. A tidewater reentrant in the main body of an ice sheet into which the ice is rapidly calving is referred to as a calving bay. (Hughes, 1977). Rapid disintegration of the Laurentide and Scandinavian Ice Sheets probably was caused largely by means of calving bays that rapidly destroyed the interior portions of these large ice sheets (Mercer, 1969; Hughes and others, 1978). Probably the tidewater portion of the Cordilleran Ice Sheet in western Washington (Juan de Fuca lobe) experienced rapid retreat as well, which caused deglaciation of the northwestern part of the study area before adjacent parts of the Puget lowland.

Deglaciation of the Northwest Puget Lowland

Most of the northwest Puget lowland west of Hood Canal was mapped by Birdseye (1976), Gayer (1977) and Hansen (1976). Their reports are largely concerned with the distribution and engineering characteristics of surface sediments and do not provide detailed information on the local history of deglaciation. Because deglacial events in this area were extremely critical to recession of the Puget lobe, the northwest Puget lowland was studied in more detail in the field. Mapping of channels and terraces and stratigraphic studies permit a tentative interpretation of the deglacial history of the northwest Puget lowland.

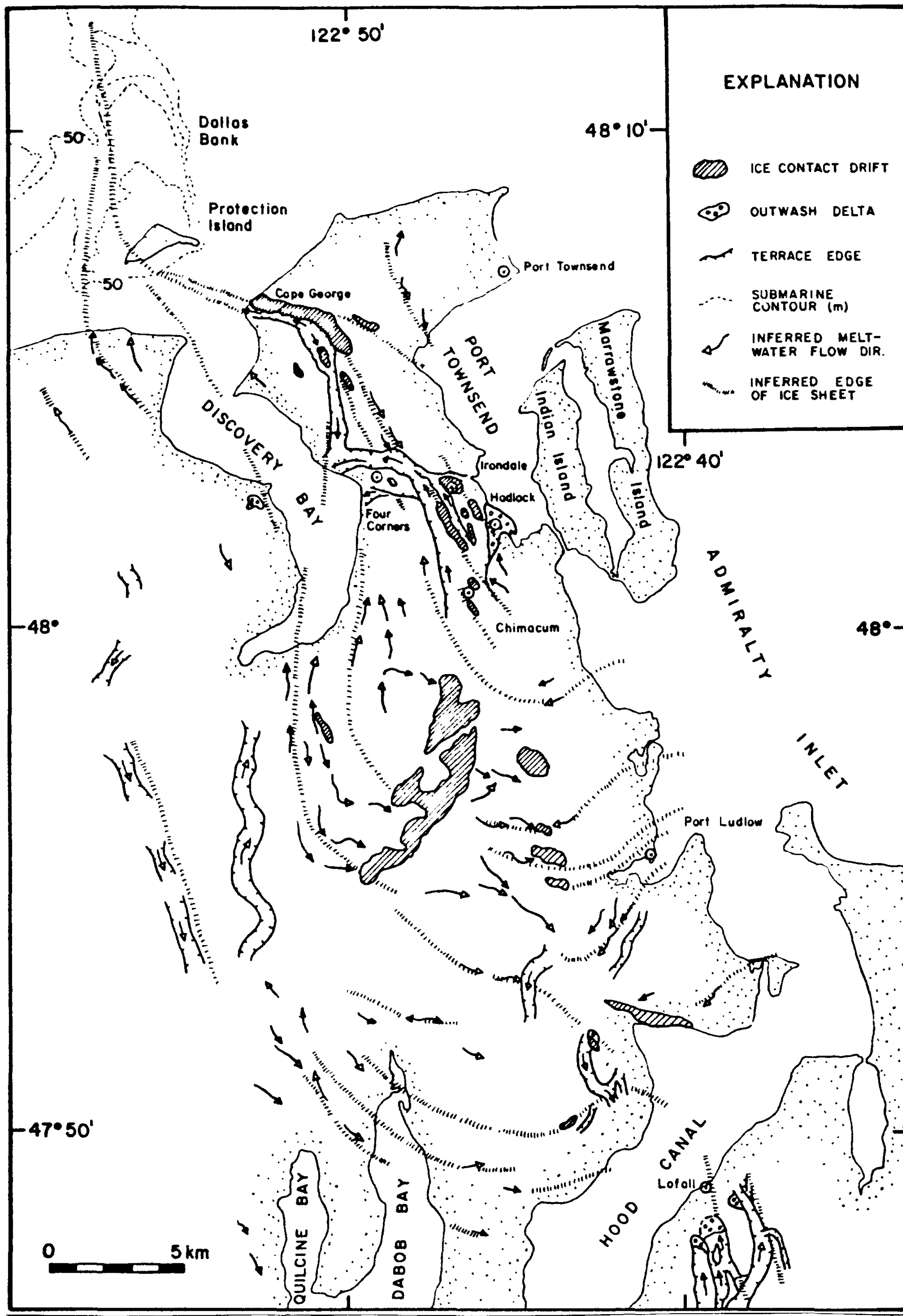

Figure 9.--Map showing deposits and features formed during deglaciation in the northwestern Puget lowland. Inferred ice-marginal positions of the glacier during retreat are also shown.

Lofall Area

Along the east shore of Hood Canal south of Lofall, a series of north-trending broad channels were carved by north-flowing meltwater streams (fig. 9). Coarse outwash, exposed near the base of the channel floors, is imbricate northward at many localities and can be traced north into local ice-contact deltas at various heights. East-facing terrace scarps and cross-cutting relationships exhibited by the channels suggest that they were ice-marginal in origin, and reflect successively lower meltwater routes to Hood Canal. The lowest major channel was described as the Poulsbo channel by Bretz (1913), who inferred that it served as a south-draining spillway from a proglacial lake in Hood Canal (fig. 7). The flat floor and negligible gradient of the Poulsbo channel do suggest that it served as a lake spillway, but the local geologic relations indicate that it drained northward, rather than southward.

The deltaic sediments near the east shore of Hood Canal at the mouth of the meltwater channels are well exposed by gravel pit operations (fig. 9). They consist predominantly of bottomset and foreset beds of sand and fine gravel; topset beds are absent. The foresets can be traced along dip continuously for as much as 10 m and are generally inclined northward. They show a high degree of variability in their local dip directions, suggesting that they were deposited during rapidly changing conditions. The deltaic sequences commonly grade vertically and laterally into ice-contact stratified drift that exhibits both small-scale collapse features and large slumps. Large boulder-sized clasts of till and unconsolidated laminated fine-grained sediment commonly occur in the deltaic sediments and ice-contact stratified drift. The foresets are exposed below several different levels ranging in altitude from about 37 m to 103 m. Exposures were not complete enough to permit detailed comparison of different levels. The deltaic deposits apparently were built into a lake that occupied Hood Canal at a time when the margin of the receding Puget lobe locally trended north-south.

Ice-contact stratified drift is also common in Squamish Harbor directly across Hood Canal from Lofall (fig. 9). The north side of the harbor is mantled with a nearly continuous cover of sand and gravelly ice-contact stratified drift more than 30 m thick in some areas. Ice-contact stratified drift also crops out in several isolated patches near Mud Lake. Two broad meltwater channels, both above about 90 m altitude, trend southward from the ice-contact stratified drift deposits. A gravelly delta at Mud Lake apparently was built northward into a local ice-dammed lake that stood at about 90 m altitude.

Lowland portions of the northeast Olympic Peninsula

Most of the northwest Puget lowland north of Hood Canal consists of an irregular drift plain that ranges in altitude from about 90-100 m. Meltwater drainage channels on the upland exhibit a general pattern that is convex to the south (fig. 9). Most channels trend dominantly northwest near the Olympic Mountains, east across the central drift plain, and northeast toward Admiralty Inlet. The orientation of the channels suggests that a distinct sublobe of the receding Puget lobe developed over this region during retreat.

Hummocky masses of poorly washed ice-contact stratified drift as much as 30 m high and several hundred meters wide occur in Chimacum Valley between Squamish Harbor and Chimacum (Hansen, 1976; figs. 9, 10). Four relatively distinct drift bodies, each oriented approximately northwest-southeast, are separated by flat areas underlain by silty and sandy lake sediment. Each mass of drift apparently lies in the valley bottom near the mouth of an ice-marginal meltwater channel.

West Valley, which lies several kilometers west of Chimacum Valley, is also nearly plugged with ice-contact stratified drift. The drift varies from coarse boulder gravel to nonsorted silty diamicton. Individual drift bodies cannot be distinguished in West Valley; rather, the drift mantles 10 km of the north-south valley. Small areas within the drift are commonly flat-bottomed and presumably underlain by lake sediment.

The relation of the drift in Chimacum and West Valleys to the meltwater channels on the upland suggest that north- and south-draining debris-laden meltwater streams deposited sediment in the larger valleys during systematic northward retreat. The presence of filled basins, the common occurrence of silty, poorly sorted sediment, and the presence of at least one small ice-contact delta as part of a drift complex, suggests that the retreating glacier terminated in a proglacial lake. The spacing between individual drift bodies in Chimacum Creek suggests that glacier retreat was interrupted by brief stillstands that nevertheless were long enough for a significant mass of sediment to accumulate in the valleys. On the Quimper Peninsula near Cape George, a massive body of gravelly ice-contact stratified drift about 7 km long and about 0.5 km wide extends southeast from the Strait of Juan de Fuca toward Hadlock (fig. 9). Interpreted as a moraine by Gayer (1977), it may also reflect a temporary stillstand of the receding Puget lobe. A well-defined outwash train extends both west and south from the ice-contact stratified drift near Cape George.

Well-defined meltwater channels on the upland drift plain occur elsewhere on both the Quimper and Miller Peninsulas (fig. 9). These channels are from 10 to 100 m wide, are floored with coarse outwash, and drain both northeast toward the Strait of Juan de Fuca and southeast toward Discovery Bay and Port Townsend. They indicate that the glacier margin was oriented northwest-southeast during ice recession from the peninsulas.

Four Corners-Hadlock area

The surficial geology of the area between the towns of Hadlock, Four Corners, and Chimacum on the Quimper Peninsula (fig. 9) is extremely complex and poorly understood. Although a complete evaluation of this area is beyond the scope of this report, some discussion is necessary because the area importantly affected deglaciation of the northwest Puget lowland.

Port Townsend and Discovery Bay are separated by a 2-km-broad flat-bottomed trough that also merges with the mouths of West and Chimacum Valleys (fig. 9; plate 1). Four terraces in the trough between the Chimacum Valley and

Figure 10.--Bouldery ice-contact stratified drift near Center, WA. Cattle in background give scale.

Discovery Bay progressively decrease in altitude northward. The highest terrace occurs at about 43 m altitude and exposes about 4 m of very sandy variable planar- and trough-cross-bedded gravel with large scale cross beds up to 3 m high. This sediment clearly originated as coarse outwash that was deposited by west-flowing meltwater streams. A slightly lower terrace at about 41 m altitude has no detectable gradient. It may be largely erosional near Four Corners because Vashon till and pre-Vashon sediments are exposed to within 2 m of the terrace scarp along the west side of Discovery Bay. At the terrace surface near Four Corners about 1.5 m of blocky laminated silt overlies poorly sorted gravel of unknown thickness. The gravel is of unknown origin but the silt apparently was deposited in standing water. Poorly sorted gravel, nearly identical to that at Four Corners, can be traced up the Chimacum Valley to Chimacum, where it commonly contains armoured mudballs and rip-up clasts. Crossbedding and cobble imbrication suggest that meltwater flow was northward, but no gradient was detected on the terrace surface.

The lower terraces also exhibit no apparent gradient. In terrace exposures along the shore of Discovery Bay laminated silt overlies at least 20 m of flat-bedded and low-angle crossbedded sand and fine gravel which commonly contains silt clasts and isolated lenses of gravel. These sandy sediments are interpreted as indicating rapid subaqueous deposition by west-flowing currents near an active glacier terminus.

The floor of the trough east of West Valley forms an irregular surface that exhibits many shallow closed depressions oriented northwest-southeast (fig. 9). The flat-floored channel north of Chimacum grades eastward into sandy ice-contact stratified drift that forms hummocky mounds and irregular surfaces as high as 10 m and as broad as 1 km. This sandy drift rises as much as 10 m above the generally flat surface of the trough, and in places is discontinuously blanketed by silt and silty clay. Similar sandy drift is overlain by a gravel outwash delta at Hadlock, which was built toward Port Townsend by streams that drained north from Chimacum Valley.

The deposits and land forms in the Chimacum-Hadlock area tentatively can be interpreted to reflect the following deglacial sequence. Meltwater from the Puget lobe originally drained west into Discovery Bay via a broad aggrading outwash train south of Four Corners. Continued retreat may have allowed the lakes that were impounded in West and Chimacum Valleys to drain northwest to Discovery Bay, eroding the older outwash deposits. At that time the broad trough between Discovery Bay and Chimacum Creek may have become an embayment of Discovery Bay that extended up the Chimacum Valley. The north-trending zones of inferred ice-contact stratified drift near Hadlock may have formed when the glacier retreated eastward during growth of the embayment. The equivalence in altitude between the eastern and western ends of the trough between Port Townsend and Discovery Bay suggests that both were graded to sea level at that time. Hadlock delta, which occurs as part of the broad surface, supports this interpretation. The crossbedded sand and fine gravel in the lower terraces along Discovery Bay probably was deposited by west-flowing currents of meltwater. Silt overlying these deposits and the ice-contact stratified drift northwest of Hadlock may be glacial-marine mud that accumulated after rapid meltwater flow had ceased in the area.

Meltwater impounded in Puget Sound during deglaciation could possibly have drained to the sea through Chimacum Valley when the Four Corners area was deglaciated. However, the absence of significant meltwater erosion of Everson age in Chimacum Valley suggests that large volumes of impounded meltwater did not drain from Puget Sound to Discovery Bay through this valley. Perhaps erosion was inhibited because the floor of Chimacum Valley probably lay near sea level at that time. Alternatively, stagnant glacier ice that lay in Chimacum Valley may have prevented drainage of the lake.

Radiocarbon Chronology of the Puget Lobe

The earliest probable age for glaciation of the Puget lowland during the Vashon Stade has been determined from radiocarbon dates on wood obtained near Seattle (figs. 2 and 3). Two dates of 15,000 ± 400 (S-1227) and 15,100 ± 600 (W-1305) ^{14}C yr B.P. (same age within one standard deviation) were obtained from wood in colluvium that lies stratigraphically below proglacial-lake sediments (Lawton Clay Member) of Vashon drift that formed during the Vashon advance (Mullineaux and others, 1965). These dates indicate that the Puget lobe had not yet reached the latitude of Seattle by about 15,000 years ago. Other limiting dates of 16,070 ± 600 (W-2125), 18,100 ± 700 (W-1186) and 18,020 ± 600 (W-1892) ^{14}C yr B.P. from samples taken from below Vashon till are consistent with the more closely limiting maximum age of about 15,000 yr B.P.

The minimum age for Vashon glaciation of the Puget lowland has been determined from radiocarbon-dated postglacial sediments just north of the region (fig. 3). These dates, which range from about 12,000 to 13,000 yr B.P., were obtained from marine shells and terrestrial peat formed after retreat of the Vashon glacier. Four dates of 13,100 ± 170 (UW-32), 12,900 ± 330 (UW-398), 12,600 ± 190 (I-1881), and 12,535 ± 300 (I-1079) ^{14}C yr B.P. indicate that the glacier terminus had retreated north of the Puget lowland by about 13,000 years ago.

A date of 13,570 ± 130 (UW-35) ^{14}C yr B.P. from late-glacial sediments in the Snoqualmie Valley (Porter, 1976), and 13,650 ± 550 (L-346a) ^{14}C yr B.P. from basal postglacial sediments in Lake Washington have been interpreted as indicating that the Puget lobe had retreated north of Seattle by about 13,500 yr B.P. Because the first date is based on an indirect association and the latter date has a large standard deviation, the date of 13,500 yr B.P. is only an approximate age for deglaciation of the Seattle area. These dates support the interpretation that the Vashon glacier had retreated north of the Puget lowland by about 13,000 yr B.P.

These dates indicate that the Puget lowland was occupied by ice for only about 2000 years. No precise date can be given for the culmination of the Vashon advance or for the rates of advance or retreat. The minimum mean rate of advance and retreat must have been about 90 m/yr.

PATTERN OF POSTGLACIAL DEFORMATION IN THE PUGET LOWLAND

The surfaces of former bodies of water (water planes) can be studied to determine the pattern of deformation that affected them after their formation. Present differences in altitude between features that clearly formed along the same shoreline must be due to differential uplift or subsidence. In the Puget lowland such altitudinal differences can be used to determine the pattern of postglacial deformation because many marine and lacustrine shoreline features formed during retreat of the Puget lobe.

U.S. Geological Survey 1:24,000- and 1:62,500-scale topographic maps of the Puget lowland were not accurate enough to determine the exact altitudes of water-plane features that were used to reconstruct the pattern of postglacial deformation. The smallest available contour interval generally is 20 ft (6 m), but 50-ft (15-m) contours are the smallest available in the southwest Puget lowland. The altitudes of specific features that were used to reconstruct relative deformation were determined with an automatic level and a transit, accurate to the nearest 0.1 m. Instrumental accuracy was determined by closing several traverses, and was found to be about 0.03 m. Base-level control was taken from the nearest benchmark, where possible, or from the altitudes of road intersections given on the U.S.G.S. maps. The altitudes of marine features exposed in beach cliffs in the northern Puget lowland were determined by measuring their height above high tide with a spirit level; these altitudes are accurate to within about 1.0 m above high tide. The remaining altitudes cited in this report were estimated by interpolation from 20-ft (6-m) contour intervals.

Lake Shoreline (Water-Plane) Data

Lake Sequence

At the culmination of the Vashon advance the rivers and streams of the Cascade Range and Olympic Mountains that drained into the Puget Sound basin were dammed by the Puget lobe, creating a number of lakes at various altitudes that fringed the ice sheet. Along the mountain fronts each lake drained into the next alpine valley to the south; some ice-marginal lakes drained directly over the ice, and were controlled by the altitude of the glacier surface. Marginally dammed lakes near the terminus drained south via specific spillways away from the ice front to the Chehalis River.

During retreat of the Puget lobe, northward drainage to the sea was blocked by glacier ice, and the troughs of the Puget lowland were occupied by large proglacial lakes that drained south to the Pacific Ocean via the Chehalis Valley (plate 1). The evolution of the proglacial lakes was originally outlined by Bretz (1913) and prior to this study has been only slightly modified by subsequent

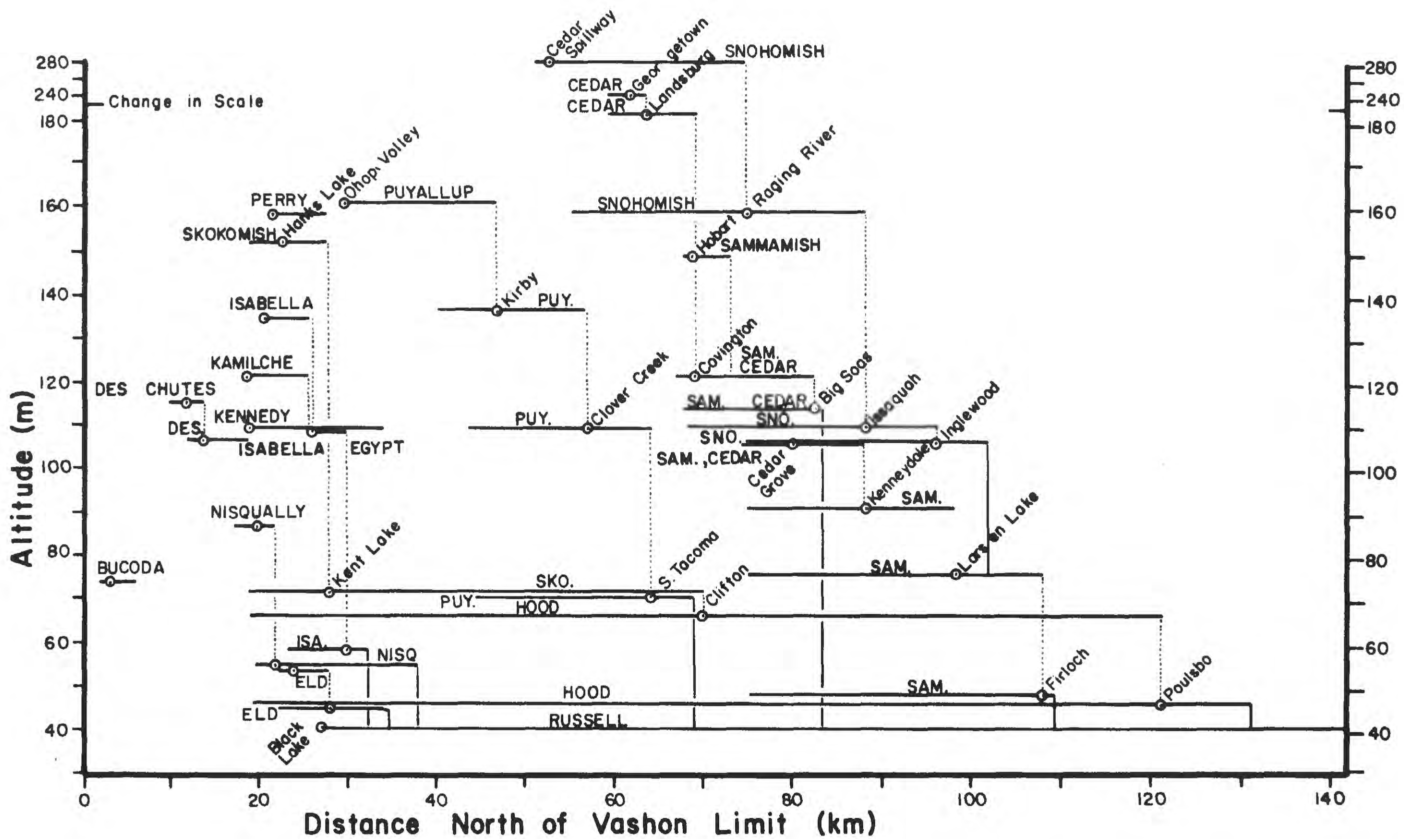

Figure 11.--Inferred sequence of proglacial lakes in the Puget lowland, based on the assumption that no postglacial warping or spillway erosion occurred. Open circles indicate locations and altitudes of major lake spillways. Solid vertical lines indicate a postulated termination of proglacial lakes as they fall to lower lakes. Dotted lines indicate a postulated drop in altitude within a proglacial lake. Horizontal scale is plotted parallel to axial flowline "E" of figure 4.

Figure 12.--South end of Clifton channel near Belfair showing typical channel form. Channel is about 20 m wide at base. Flow direction was away from the viewer. Stumps are about 1 m high.

investigators. The lake sequence was complex because systematic northward retreat caused progressive uncovering of lower lake spillways, resulting in the coalescence of smaller lakes to form larger ones at lower levels (plate 1).

Initial ice recession in the southwest Puget lowland exposed drainage divides at the heads of the Skokomish Valley, and the valleys of Isabella, Camilche, and Kennedy Creeks. These divides separated drainage from the Puget Sound region and the Chehalis Valley, and served as outlets for the water impounded north of the divides. Glacial Lake Nisqually occupied a large reentrant between the east-retreating east sublobe and the north-retreating west sublobe. Glacial Lake Russell occupied the troughs of the southern Puget lowland and expanded northward as the ice sheet receded. Glacial Lakes Skokomish and Hood formed in the Hood Canal basin, but were controlled by different spillways. Glacial Lake Puyallup extended north from the Cascade Range foothills in the Puyallup and Duwamish valleys. Glacial Lakes Cedar, Sammamish, and Snohomish occupied respectively the Cedar River, Sammamish, and Snoqualmie Valleys and each expanded northward during gradual recession of the Puget lobe.

To use the lake shoreline features in the Puget lowland to determine the pattern of postglacial deformation, each feature must be related to a specific lake surface (water plane) independently of its present altitude. The sequence of water planes that formed beyond the retreating Puget lobe can be reconstructed from the altitude of the controlling lake outlets (fig. 11). Each proglacial lake occupied a basin beyond the reconstructed ice front (figs. 8 and 12) that lay below the level of the outlet (fig. 12). The altitudes of observed features associated with any water plane can then be compared to the altitude of the originally horizontal water plane; apparent differences in altitude between the two must be due to postglacial deformation. The causes of apparent deformation, such as regional tilting, folding, or faulting may then be inferred from the pattern. This method is a fairly straightforward and accurate means of determining the amounts and character of postglacial deformation because it does not rely on the variable positions of past sea levels.

Most of the recessional proglacial lakes that formed in the Puget lowland cannot be used to reconstruct the pattern of postglacial deformation. Reconstructed lakes can be used only if the following conditions are met: (1) the amount of spillway erosion associated with each water plane must be measurable and minimal, (2) the shoreline features used must be unequivocally related to a specific water plane in the lake basin under consideration, and (3) water-plane features must be areally separated so that differential warping can be measured.

The ice-marginal lakes that were drained directly over or against glacier ice cannot be used because the height of each lake was continually changing as the glacier thinned. Any particular shoreline feature could have been formed at any time during gradual lowering of an impounded lake. The early lakes that formed south of the Vashon limit were controlled by distinct spillways that drained southwest to the Pacific Ocean, but because these lakes extended only a few kilometers north of the spillways and because no well-defined shoreline features were recognized, they cannot be used to reconstruct deformation (fig. 11). The surface of Lake Nisqually was not controlled by well-defined spillways, hence it also cannot be used.

Lake Puyallup cannot be used to reconstruct differential postglacial deformation because the lake surface was gradually lowered during spillway incision, and because no shoreline features have been identified. The surface of Lake Puyallup was controlled by five major spillways, each eroded to a depth of at least 15 m (fig. 11). Any shoreline features that may have been formed in Lake Puyallup could have been built when the lake stood at almost any height between 73 m and 240 m, hence the amount of differential postglacial deformation cannot be determined.

Lake Skokomish cannot be used to determine subsequent deformation because its surface altitude was controlled by spillways that were continually widened and deepened. A shoreline feature that might have formed in the lake basin could be related to an unspecified stage during spillway erosion.

Glacial Lakes Sammamish, Cedar, and Snohomish had a complicated drainage history during ice recession because of the high local topographic relief and the great number of deeply eroded lake spillways (fig. 11). As many as seven different water planes of Glacial Lake Sammamish, and five water planes of both Glacial Lake Snohomish and Glacial Lake Cedar can be inferred during northward opening of successively lower spillways. The heights of almost all water planes were also constantly changing because of spillway erosion. Although large and well defined deltas and terraces at many levels are widespread in this area, they cannot be assigned to specific water planes, thus Glacial Lakes Cedar, Sammamish, and Snohomish cannot be used to determine postglacial deformation.

Only Glacial Lakes Russell, Hood, and a younger north-draining lake to be described later (Lake Leland) can be used to document the pattern of postglacial deformation in the Puget lowland. These lakes were controlled by spillways that were not deeply eroded, the lakes were large enough for possible warping to be recognized, and numerous shoreline features formed along their margins.

Glacial Lake Russell

Recession of the Puget lobe to a position about 25 km north of its terminus exposed the Black Lake spillway (plate I). The spillway drained west from Budd Inlet via the valley of Percival Creek to Black River and ultimately to the Pacific Ocean via Chehalis Valley. Glacial Lake Russell occupied the troughs of the Puget lowland that lay north of the spillway, and enlarged northward as the ice receded (fig. 11). The Black Lake spillway is about 10 km long, and grades directly into the present floodplain of Black River at its southern end. It is about 0.2 to 1.0 km wide, being narrowest in its middle portion where cut into local sedimentary rocks, and widest south of Black Lake where incised into coarse recessional outwash plains. The gradient of the spillway is almost imperceptible and cannot be exactly determined owing to a later fill of peat and alluvial-fan sediments. The channel slopes southwest at about 0.3 m/km near its middle portion.

The Black Lake spillway is generally shallow (< 10 m) over much of its length and follows a trough-shaped flat-bottomed channel. Bedrock is exposed in the channel flanks south of Tumwater near Black Lake. Near its crest (the highest part of the spillway floor), about 2 km northeast of Tumwater, the spillway is carved into an outwash plain that overlies till and well-indurated Pleistocene sediments. Lenses of very coarse rounded to subrounded boulders on the floor of the spillway lie with pronounced uncomformity on the older, more-compact sediments.

The outwash plain into which the spillway is incised consists of well-defined beds of sandy gravel with crossbeds up to about 1 m thick. West of the spillway crest the outwash plain slopes about 1° southwest. Farther east outwash sediments are coarser and more bouldery, and in places appear similar to ice-contact stratified drift. The head of the outwash plain terminates abruptly at the west wall of Budd Inlet where a 1-km-long terrace edge is now exposed, suggesting that the terrace scarp is an ice-contact face and that the outwash train was deposited when the Puget lobe occupied Budd Inlet. Creation of the spillway and its initial incision, therefore, probably occurred immediately after withdrawal of the Puget lobe from the west side of Budd Inlet.

The shallow depth of the Black Lake spillway near its crest is very significant. The highest point of the spillway floor lies at 41.0 m. The wide braided outwash plain just above the spillway floor lies at 49.7 m indicating that less than 9 m of erosion occurred during the existence of Glacial Lake Russell. A lower, but less-well-defined outwash terrace just southwest of the crest lies at about 45 m, suggesting that incision of the spillway during the existence of Glacial Lake Russell was less than about 4 m.

The Black Lake spillway was the lowest meltwater route to the Pacific Ocean during recession of the Puget lobe. Lake Russell occupied nearly all the troughs of the lowland at least as far north as Seattle (plate 1). The Black Lake spillway was abandoned only after a lower meltwater route to the Strait of Juan de Fuca was exposed.

Glacial Lake Hood

Retreat of the Puget lobe in Hood Canal exposed the Clifton channel (fig. 12; plate 1) which lies between Hood Canal and Case Inlet, about 10 km south of Belfair (Bretz, 1913). The flat-floored channel is about 3 km long and has a relatively constant width of about 200 m, but near Deveraux Lake it widens to about 400 m. The northern end of the spillway along the valley wall of Hood Canal has been deeply incised by the modern stream; no evidence for north-flowing spillway drainage was observed. No southerly gradient was detected along the main channel either, partly because small postglacial alluvial fans have been built onto the valley floor.

The Clifton channel is incised through the fluted Vashon till plain. Two small gravel pits near the spillway crest (66.5 m) expose about 3 m of loose sandy cobble gravel that apparently was deposited during Vashon glacier recession. The amount of incision that occurred in the Clifton channel during overflow of Lake Hood cannot be determined accurately, but possible terracing along the south valley wall suggests less than 18 m.

Lake Hood, which occupied the Hood Canal troughs after the Clifton channel was cleared, drained south to Lake Russell (Bretz, 1913). It extended north in Hood Canal until a lower route was exposed to the Strait of Juan de Fuca. Lake Hood must have been at nearly the same altitude as Lake Russell because there is no apparent gradient along the channel floor, which is cut into unconsolidated Quaternary deposits. If the lakes originally lay at substantially different altitudes, the relatively narrow neck of land between Hood Canal and Puget Sound would have been rapidly eroded.

Glacial Lake Leland

A pronounced, flat-bottomed, low-gradient, north-draining spillway between Quilcine Bay and Discovery Bay follows the valley of Leland Creek, is 10-15 km long, has a fairly constant width of 200-400 m, and reaches an altitude of 68.5 m at its crest (fig. 9; plate 1). Silty lake sediments extend to an altitude of about 60 m in the valley of Quilcine Bay south of the spillway head. The depth of the Leland Creek spillway varies considerably and cannot be determined accurately because the spillway occupies a 300-m-deep valley that formed prior to overflow of the proglacial lake. Near its crest, the spillway is incised into ice-contact stratified drift that extends in altitude to as low as 73 m, suggesting that ice was still present in the area when the spillway was incised to this depth. Bedrock is exposed in the walls of the spillway for much of its length. The channel gradient appears lowest (ca. 2 m/km) at the crest, but cannot be determined accurately because of postglacial alluvial fans.

The existence of the Leland Creek spillway is incompatible with the hypothesis that postglacial deformation has not affected the Puget lowland, for it would have been impossible to maintain northward lake drainage over a spillway at 68.6 m altitude if the Black Lake spillway (41.0 m) provided a lower drainage outlet from the same water body. Because it is unlikely that the Leland Creek spillway drained a local ice-dammed lake, the apparent difference in altitude between the two spillways probably is due to differential postglacial deformation of at least 28 m. The lake that must have been impounded when meltwater drained through the Leland Creek spillway is hereafter referred to as Glacial Lake Leland.

Deposits of Proglacial Lakes

Recessional lake-bottom sediments of laminated silt and clay crop out at altitudes as high as 50 m in the southwestern and west-central part of the Puget lowland (Molenaar and Noble, 1970; fig. 13). They are sparsely exposed over other parts of the Puget lowland because they are largely submerged beneath Puget Sound, Hood Canal, and Lakes Sammamish and Washington, and because they are easily eroded. They cannot be used for determining the specific deformation of water planes because lake-bottom sediments provide only a minimum estimate for the height of a proglacial lake (fig. 14).

Figure 13.--Silty lake clay near Eldon along west shore of Hood Canal. Laminations and fine horizontal bedding characterize these sediments.

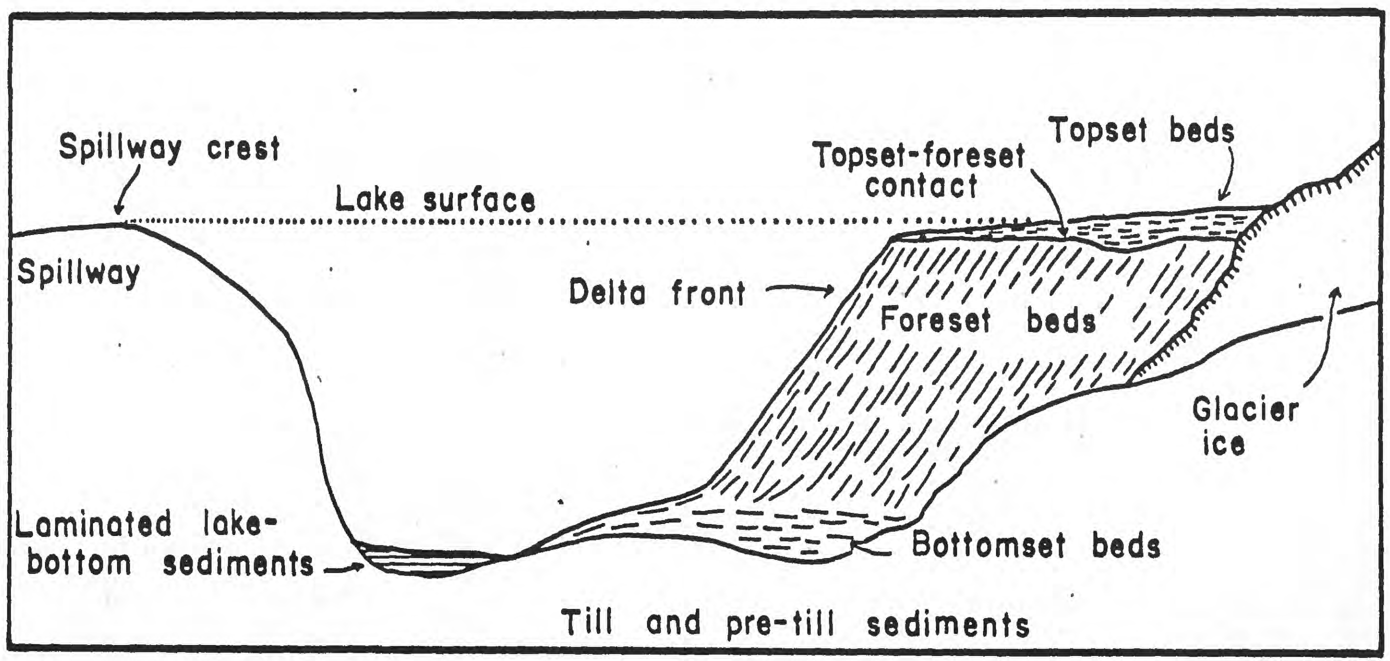

Figure 14.--Generalized diagram showing relation of proglacial lake features to the lake surface. Lake surface lies just above the crest (divide) of the spillway and the contact between delta topset and foreset beds.

Beach features provide a more accurate means of determining the height of proglacial lake shorelines and the degree of deformation that has affected them. Unfortunately, however, no beach deposits or coastal terraces were found anywhere in the Puget lowland that could be definitely related to individual water planes. If present, they were not observed on the relatively steep, thickly vegetated slopes that characterize the region. The time for evolution of proglacial lakes may have been too brief to have formed well-defined beach deposits or terraces.

Besides the lake outlets, the only water-plane features found in the Puget lowland are large deltas that were built into proglacial lakes by meltwater streams from the retreating ice margin (figs. 14 and 15; plate 1; table 3). The deltas occur throughout the lowland and commonly were built against the retreating glacier front, as indicated by abundant ice-contact features in the delta deposits such as bouldery zones and collapse structures.

The contact between topset and foreset beds in these deltas is commonly well exposed, nearly horizontal, and clearly defined (fig. 16). Reworking of the delta sediment by currents and waves often accentuated the topset-foreset contact by producing a pronounced stone line or well-sorted horizon. The topset-foreset contact commonly is nearly horizontal, with less than 50 cm relief. The topset-foreset contact does not exactly correspond to a proglacial lake surface, but lies slightly below it (fig. 14), marking the depth to which fluvial currents can no longer carry the sediment load in traction. For modern gravelly outwash streams in analogous conditions, the contact has been found to range between 0.5 and 4 m below the lake surface, but commonly lies less than 1 m below it (Gustavson and others, 1976). The horizontality, lack of erosional relief, and the presence of wave-washed horizons suggest that topset-foreset contacts of outwash deltas in the Puget lowland formed just below the proglacial lake surfaces.

The errors in reconstruction caused by variations in water depth over spillways and the insensitivity of the topset-foreset contact are in the same direction. The lake surface should lie slightly above both the floor of the spillway and the topset-foreset contact (fig. 14). Significant differences in altitude between the topset-foreset contact of a delta and the spillway that controlled the height of the lake into which it was built, must be due to postglacial deformation because these features originated at nearly the same altitude.

Glacial Lake Russell

Shelton Delta. The Shelton delta extends for several kilometers north of Shelton along the west rim of Oakland Bay in southern Puget Sound (plate 1). The delta surface is generally flat and grades upslope (westward) into McEwen Prairie, an extensive plain of coarse outwash that rises to the west. The delta consists mainly of sandy gravel with rounded to subrounded clasts and includes granitic clasts of northern provenance derived from the Puget lobe. The gravel also contains a relatively high proportion of oxidized clasts derived from coarse pre-Vashon gravel deposits of southern Olympic Mountains provenance. The topset-foreset contact is well exposed at 51.5 m altitude, and is nearly horizontal along most of the exposed top of the delta, though a well-defined 1-m-deep channel is present at one locality. The topset-foreset contact also can be recognized on a textural basis. Gravel of the topsets is generally coarser and more well sorted than in the foresets. Openwork well-sorted sheet-like horizontal beds of gravel occur along the contact in places, suggesting that waves or currents may have partially altered the foresets during deposition.

The Shelton delta was built by east-flowing streams that drained about 10 km of McEwen Prairie. The relatively thick, texturally variable foreset beds suggest that deposition was fairly rapid, and from currents of variable strength and direction. The delta prograded eastward into what is now southern Puget Sound. Lake Russell was the only proglacial lake known to have occupied this area during ice recession.

Sequalchew Delta. The Sequalchew Delta extends for several kilometers along the present margin of Puget Sound northeast of the mouth of Nisqually River, and covers at least 3 km^2 (plate 1). The top of the delta is nearly flat, but is marked by long, broad depressions apparently caused by subsidence at depth, and depressions up to 30 m deep and 100 m wide that were apparently caused by melting of shallow buried ice-blocks. Delta sediments consist mostly of sandy gravel of northern provenance. The delta front has a smooth, slightly convex outline towards Puget Sound. The topset-foreset contact was not exposed during this investigation, but topset beds were traced down to at least 62.1 m altitude. Bretz (1913) observed a definite nearly horizontal topset-foreset contact at 185 feet (56.4 m) altitude.

The Sequalchew delta was built by west-flowing meltwater streams, and prograded westward into Glacial Lake Russell. The present outwash plain, however, has no apparent westward gradient suggesting that the delta sediments may have been derived partially from the Puget lobe. The presence of boulders, shallow kettles, and widespread coarse gravel also suggest that the Puget lobe was very near and that deposition was probably rapid.

Steilacoom Delta. The Steilacoom delta is located along the shore of Puget Sound about 5-10 km south of Tacoma at the western edge of the Steilacoom gravel plains (plate 1). The delta front is uniform in height, even-crested, and is slightly convex toward Puget Sound. The delta extends along the shoreline from Chambers Bay to Sunset Beach and inland at least 5 km. The delta surface is generally very flat, with no detectable westward gradient or thickening of the topset beds. Gravel excavations and boreholes in the region are common and indicate that the delta sediments occur as part of a westward-thickening wedge of outwash gravel that overlies Vashon till and pre-Vashon sediment. Broad depressions up to 10 m deep and 100 m wide occur in the topset and foreset facies suggesting that the deformation was caused either by melting of ice at great depth, or by subsidence during compaction of sediments.

Near Chambers Bay the topset-foreset contact (62.8 m) is extremely well defined, horizontal, and has no significant relief. The coarse zone of cobbles and

Figure 15.--View of Redmond delta. Exposed bank is about 30 m high and shows topset beds overlying foreset beds. Bottomset beds are exposed just beyond the left margin of the photograph.

TABLE 3.-- LACUSTRINE FEATURES USED TO RECONSTRUCT PROGLACIAL LAKE SURFACES ASSOCIATED WITH DEGLACIATION OF THE PUGET LOWLAND.

Proglacial Lake	Feature	Altitude (m) Topset-foreset Contact	Altitude (m) Spillway Crest
	Black Lake spillway	---------	41.0
	Shelton delta	51.5	----
	Sequalchew delta	56.3-62.2	----
Russell	Steilacoom delta	62.8	----
	North Auburn delta	73.2	----
	Renton delta	94.2	----
	Kitsap Lake delta	94.2-99.1	----
	Clifton channel	---------	68.0
	Belfair delta	75.3	----
Hood	Eldon delta	93.2	----
	McKenna Falls delta	95.7	----
	Upper Fulton Creek delta	93.6	----
	Lower Fulton Creek delta	29.3	----
	Brinnon delta	38.4	----
	Quilcine delta	50.0	----
Leland	Leland Creek spillway	---------	68.6
	Poulsbo channel	---------	47.0
	Richmond Beach delta	55.5 $\pm$ 2.0	----
Sammamish	Totem Lake channel	---------	38.4
	Redmond delta	40.9	----
Snohomish	Redmond channel	---------	34.6
	Harris Creek delta	32.0-44.2	----

Figure 16.--Topset-foreset contact in the Steilacoom delta. Horizontal layer of large cobbles represents lag deposit formed at the contact by wave and current action. Foreset beds dip away from viewer. Survey rod is 4 ft (1.3 m) long.

pebbles that mark the contact probably represents winnowing of the finer sediments by wave action (fig. 16). Elsewhere in the delta the contact has been disturbed by large-scale shallow subsidence that involves all the delta facies. Because repeated measurements of the contact altitude would reflect various degrees of subsidence, the maximum contact altitude at Chambers Bay was taken as representative of the whole delta.

Large-scale planar cross beds as tall as 1 m and sloping westward are common in the topset facies. In places individual crossbed sets are nearly parallel, giving the impression of delta foreset bedding. Bretz (1913) described two sets of foreset beds in the Steilacoom delta and interpreted them to represent two separate proglacial lake levels. Possibly Bretz misidentified these parallel sets of large-scale cross beds as a second set of foreset beds.

Rapid facies change, abundance of collapse and fault features, coarse size of the delta gravel, variable strike of foresets, and collapse of the delta surface suggest that the Puget lobe was near the accumulating Steilacoom delta during its deposition. Two different dominant dip directions of the foresets (west and south) suggest two sediment sources. Ice-marginal meltwater streams that drained westward across the Steilacoom plains were the dominant sediment source. The Puget lobe may also have contributed sediment directly to the delta from the north as the delta prograded westward into Lake Russell.

<u>Auburn Delta</u>. The east wall of the Duwamish Valley near Auburn exposes voluminous recessional deltaic sediments that extend north-south for about 5 km (plate 1). These sediments were deposited by west-flowing meltwater that occupied the present White and Green River valleys. Delta gravel throughout this general area consists of a mixture of rounded to subrounded volcanic and granitic clasts characteristic of Puget lobe sediment in this vicinity (Mullineaux, 1970). Three distinct major delta levels, successively lower to the north, were differentiated during this study. Large gravel-pit exposures in each level permit their internal structure to be determined. The widespread presence of ice-contact deformation features and large boulders within and on all three levels suggest that the glacier margin lay nearby during formation of each level.

The highest delta, the south Auburn delta, occurs in a north-draining tributary valley south of the White River. The delta appears generally flat-topped, but contains several distinct surfaces at slightly different altitudes. All delta facies are exposed, but contacts are difficult to determine precisely because different facies have been juxtaposed by slumping, portions of the delta appear to have been eroded away, and because nonstratified masses of ice-contact stratified drift have been inset against finer-grained and more-well-bedded units.

Bottomset beds are well defined, ripple-marked laminated sequences of silt and sand that have been broadly warped in most places (fig. 17). The foreset beds range in texture from nonbedded pockets of well-sorted boulder gravel to silty openwork gravel, and to sandy fine gravel. Individual beds are very thick and dip dominantly west, but vary in azimuth from 180°-340°. Faults and collapse structures are extremely common in the foreset and bottomset beds, particularly near the north end of the exposure. Topsets are 2-4 m thick and vary from well-sorted gravel to pebbly sand. They commonly occur in planar crossbedded units as much as 1 m high. The topset-foreset contact (about 120 m) is very irregular, poorly defined, and absent in some areas. Toward the north it appears to decline slightly in altitude in a stepwise fashion. The south Auburn delta was deposited by meltwater derived from the Puget lobe when it dammed the small north-draining tributary valley. The variable levels of the delta and its ice-contact nature indicate that the impounded lake drained southward over the ice during glacier retreat.

The Auburn delta lies both north and south of the White River southeast of Auburn. The upper surface of the delta is nearly level but is densely kettled near the White River and possibly decreases northward from about 91 m to 85 m. It was not possible to determine whether a single delta surface declines northward or whether a series of levels at slightly differing altitudes are present. The Auburn delta is best exposed in a large gravel pit just south of the White River. The topset-foreset contact there (89.0 m) is very well defined and shows less than about 1 m relief. The Auburn delta was built by west-flowing ice-marginal streams that drained into a large proglacial lake occupying the Duwamish Valley.

The north Auburn delta lies just south of the Green River east of Auburn and extends about 1 km northwest of the slightly higher Auburn delta. Its upper surface is poorly defined due to the presence of a large kettle near White Lake and to extensive gravel excavations that provide excellent exposures. The topset-foreset contact, which lies at an altitude of 73.2 m, is clearly marked and continuous for at least 30 m. No pronounced channelling or significant relief were observed along the contact.

If the sequence of proglacial lakes in the Auburn area is reconstructed independently of the delta deposits, three separate lake levels should have existed. The highest level would have formed within the north-draining tributary valley south of the Green River. Its level would have dropped continuously as the ice retreated northward. When ice retreated north of the tributary valley mouth, the local lake should have merged with Glacial Lake Puyallup, which occupied the Duwamish Valley until the ice retreated north of Tacoma Narrows. During the time the Auburn area was submerged by Lake Puyallup, the South Tacoma Channel controlled the height of the lake surface (plate 1); incision of this channel should have caused a general lowering of the lake during ice retreat. When the Puget lobe receded north of Tacoma Narrows, Lake Puyallup would have dropped to the level of Glacial Lake Russell which continued to expand northward in the Duwamish Valley.

The correspondence between the observed delta levels in the Auburn area, and the deduced lake sequence strongly suggests that each delta built into a separate lake. The ice-contact character of the earliest delta is consistent with the inferred ice-dammed lake. The possible decreasing altitude of the Auburn delta is compatible with gradual incision of the Clover Creek channel. The north Auburn delta, which was built into a lake that stood at or slightly below the level of this channel, is inferred to have been built into Lake Russell. Lake Russell was the only proglacial lake known to have existed in the Duwamish Valley north of the Auburn area.

Renton Delta. The Renton delta is part of a 2 km^2 flat, kettled terrace north of the Cedar River and just east of Renton (plate 1). The terrace, which ranges between 90 and 110 m altitude, consists of recessional gravels, and has been extensively modified by human activity. A large gravel pit near the delta shows that the deposits are usually thin (< 10 m) and overlie early Vashon lacustrine and fluvial sediments that are difficult to distinguish from the delta sediments. The highest occurrences of foreset beds averages about 93.5 m altitude. In places, the entire delta sequence has been eroded by northwest-trending channels.

The Renton delta was derived from the Cedar River when it carried coarse outwash northwest over the Vashon till plain. Both the northwest-trending channels on the delta surface and the occurrence of ice-contact features within the delta suggest that it was built against the retreating glacier. The distribution and direction of meltwater transport indicates that the delta was not built into a local ice-dammed lake, but rather into Lake Russell when it occupied the northern part of the Duwamish Valley.

Kitsap Lake Delta. The Kitsap Lake delta lies just west of Kitsap Lake near Bremerton (plate 1). The steep front of the delta is 60 m high and extends north for about 0.5 km. The delta surface is flat, continuous, and grades westward upvalley into very well defined gravel terraces. No collapse features or meltwater channels are present on the delta or terrace surfaces. The Kitsap Lake delta was described by Sceva (1958) who observed large foreset beds exposed in a deep gravel pit. Foresets could be traced as high as 94.2 m, and topsets could be traced as low as 99.1 m. The topset-foreset contact was not exposed, but lies somewhere within that range. The Kitsap Lake delta was built eastward into Lake Russell by meltwater streams from the west. No local proglacial lakes are known to have occupied this area during deglaciation.

Other deltas. Bretz (1913, p. 135) described a delta above Kenneydale on the eastern bluffs of Lake Washington as "...broken surfaced deposits of sand and gravel, of considerable extent and depth..." He wrote that the summit of the slope, a mile (1.6 km) back from the lake, was a level surface that lay at 290 feet (88.4 m), and that exposures along the flanks of May Creek revealed well-stratified gravel. Nowhere in the "Kenneydale delta" did Bretz describe delta foreset bedding. During this investigation, flat terraces were found above May Creek to about 100 m altitude. These terraces appear to be part of a valley train in the Kenneydale area that extended west toward Lake Russell, which then occupied the valley of Lake Washington. If a recessional delta exists in the area it probably lies below the 100-m level. Because no direct evidence for the delta was found, it is not considered further in this report.

Based largely on an exposure of west-dipping foreset beds near Bellevue airfield, Curran (1965) described a recessional delta extending from near Eastgate to the Duwamish Valley. West-dipping gravelly foresets were found during this investigation at a comparable altitude (107 m) on the south side of the Eastgate channel. Although the foreset strata indicate that a delta is present in the Eastgate area, no evidence was found outside of the Sammamish valley drainage.

Even though the Eastgate delta, as mapped, could have been built into Lake Russell (Luzier, 1969), it may also have been built into Glacial Lake Sammamish. For this reason the Eastgate delta is not included with the deposits related to Glacial Lake Russell.

Inferred deformation. The Shelton (51.5 m), Sequalchew (56.3-62.2 m), Steilacoom (62.8 m), Renton (94.2 m), and Kitsap Lake (94.2-99.1 m) deltas were built into Glacial Lake Russell as it expanded northward, and most were built when the retreating Puget lobe was nearby. The North Auburn delta (93.2 m) possibly was built during the lowest stand of Lake Puyallup, but the parallelism of the reconstructed lake sequence and delta levels strongly suggests that it too was built into Lake Russell. The progressive northward increase in the altitude of the Lake Russell deltas indicates that features marking the former water plane have been uplifted as much as 58 m relative to the Black Lake spillway (fig. 17).

Glacial Lake Hood

Belfair Delta. Sediment of recessional deltas is commonly exposed along the northwestern shore of Hood Canal, but only in a gravel pit exposure 8 km west of Belfair are the stratigraphic relations clear. The Belfair delta lies at the mouth of an unnamed tributary that carried meltwater sediments south to Hood Canal (plate 1). The top of the delta has almost no surface expression other than a poorly defined bench about 100 m broad. Stream incision has largely destroyed the original delta surface. The topset-foreset contact (75.3 m) is exposed for about 20 m across the top of the exposure. It is very distinct, horizontal, and shows less than about 0.5 m relief. Several beds containing a mixture of very well sorted sand and pebbles mark the contact for about 5 m along its length. The distinct bimodal sorting of these beds may reflect differential winnowing of the gravels by wave action, and by subsequent filling of the void spaces by well-sorted sand.

Sediments of the Belfair delta, derived from the south-draining meltwater channel to the north, were accreted southward into the Hood Canal basin. A broad terrace on the channel floor extends upvalley for at least 3 km, and grades south to the delta top. The lack of ice-contact features indicates that the glacier margin probably lay some distance to the north during delta deposition.

Eldon Delta. The Eldon delta lies near the mouth of the Hamma Hamma River along the west side of Hood Canal (plate 1). The flat delta top, which covers about 1-2 km^2, is slightly irregular, possibly owing to channeling during deposition. The extensively slumped northwest half presumably was built against ice. Glacially grooved and abraded bedrock are exposed for about 1 km along Highway 101 below the Eldon delta. The topset-foreset contact is poorly defined because there is no significant textural change at the contact, and because only about 5 m of it is exposed. It can be defined only by a pronounced change in dip that occurs at 93.2 m altitude.

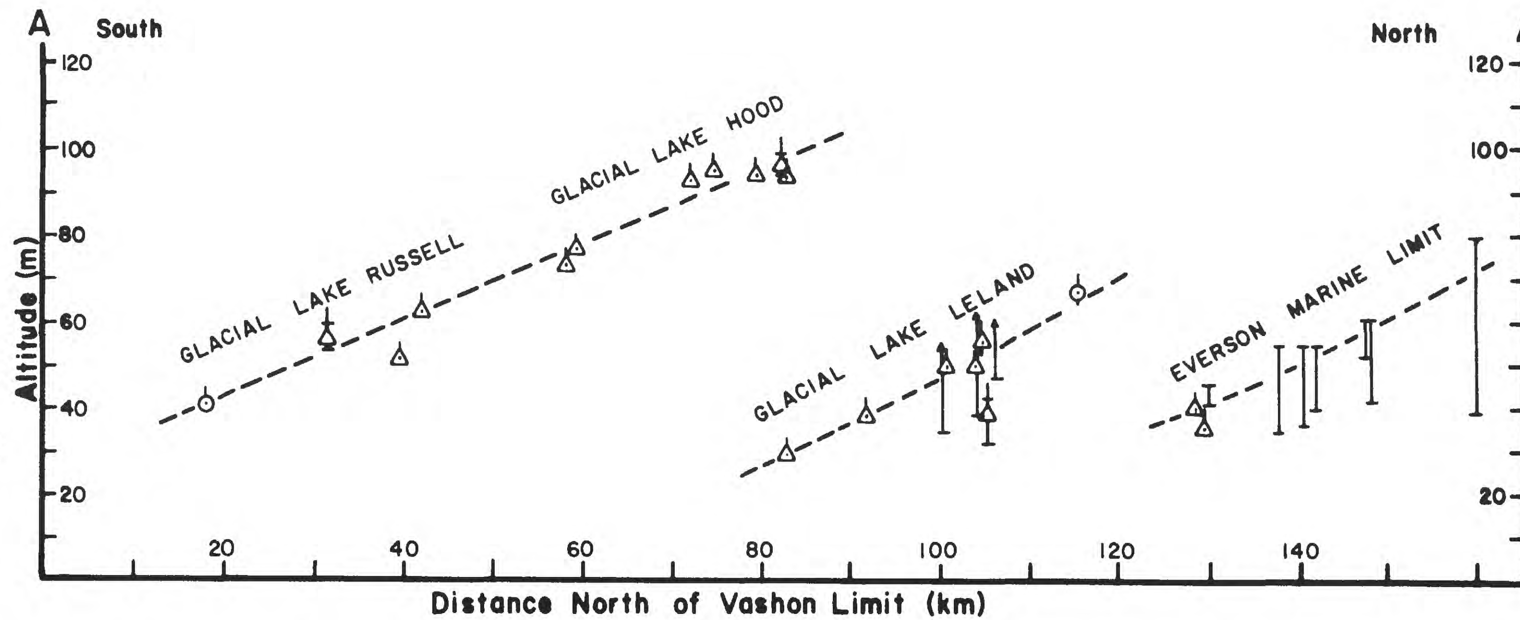

Figure 17.--Profile of water-planes associated with deglaciation of the Puget lowland plotted along line A-A' of figure 23. Circles represent lake spillways. Triangles represent deltas. Vertical bars and arrows indicate a range in altitude. Vertical lines indicate probable margin of error. Points for Glacial Lake Russell and Glacial Lake Leland that depart farthest from the mean slope represent errors introduced by projection.

The Eldon delta was built by streams flowing south and southeast to Hood Canal. The apparent upward decrease in bedding variability and thickness, the coarse texture, and the slumped nature of the deposits suggest continuous deposition during ice retreat or during a time of decreasing sedimentation rate. The highly collapsed nature of the northwest part of the delta implies that the deposit was at least initially built against ice.

McKenna Falls Delta. The McKenna Falls delta is part of a densely kettled terrace and valley fill that extends along the base of Green Mountain about 10 km southwest of Bremerton (plate 1). Although deeply dissected by tributary streams and greatly modified by gravel-pit excavations, the delta is generally flat-topped except for several large kettles near its front. The gravel is largely of northern provenance but common subangular clasts of basalt and greenstone indicate that much of it was derived locally from Gold Mountain to the north.

The topset beds occur as part of a large terrace that extends east of the main exposure of the foreset beds. Horizontal topset beds are missing in much of the exposure. In these areas distorted beds of muddy and bouldery gravel, which show folds up to 2 m amplitude, lie directly on the foreset beds. The topset-foreset contact is well defined where present and occurs at 95.7 m. Below the muddy beds the foresets extend upward to approximately the same level as the topset-foreset contact, which shows as much as 2 m relief. The nonstructured beds which overlie the foresets are not well understood; they may represent flow till, ice-contact deposition and deformation in a shallow part of a proglacial lake.

The great range in dips, wide range in sediment size, the abundance of locally derived subangular bouders, and the widespread collapse features indicate that the McKenna Falls delta is ice-contact in origin (fig. 18). The foreset dips indicate that much of the sediment was derived from meltwater streams that occupied the valley of Union River. The delta was built southward into Hood Canal and probably not into a locally ice-dammed lake. The McKenna Falls delta grades laterally in all directions to ice-contact stratified drift. Deformed ice-contact lacustrine sediments at higher altitudes that flank Green Mountain may represent local deposition in a kame terrace.

Upper Fulton Creek Delta. A broad flat-topped delta extends west from Hood Canal near the mouth of Fulton Creek (plate 1); (Carson, 1976a). The delta surface is about 1 km broad and slopes gently southward, with no obvious surface channelling or irregularities. Till and glacially abraded bedrock underlie the delta sediments which crop out in shallow cuts and house excavations along a road leading northwest from Hood Canal. They contain abundant basalt and greenstone clasts of local provenance but also include gravel of northern provenance. The topset-foreset contact (93.6 m) is very sharp because the topsets are finer in texture and lie with sharp angularity over the foresets. The contact is exposed over a distance of about 20 m and no significant erosional relief on the contact was observed.

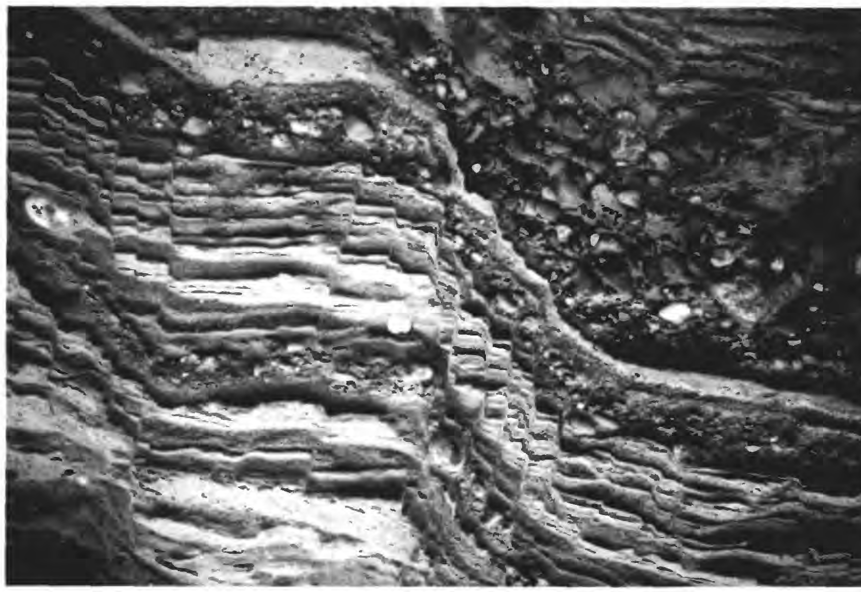

Figure 18.--Deformation in ice-contact strata within the McKenna Falls delta. Note high angle en-echelon reverse faults and wide range in sediment textures. Coin (dime) in center of photograph gives scale.

Clasts of northern provenance, ice-contact deformation in the bottomset beds, and textural similarity to the other recessional outwash deltas in the Puget lowland suggest that the Fulton Creek delta was built by meltwater streams derived largely from the Puget lobe. The south-dipping foresets indicate that the delta prograded southward into a large proglacial lake that occupied Hood Canal.

Inferred deformation. The Eldon (93.2 m), McKenna Falls (95.7 ± 1.0 m), and Upper Fulton Creek (93.6 m) deltas were built into Lake Hood. The altitude of Lake Hood was controlled by the Clifton channel (66.6 m) which has been eroded to an unknown depth (fig. 17). The Belfair delta (77.3 m) lies very near the Clifton channel and cannot be associated unequivocally with either Lake Hood or with the earlier Lake Skokomish, which was controlled by the Kent Lake spillway (73.4 m).

If the Belfair delta was built into Lake Skokomish it has been uplifted only about 2 m relative to the Kent Lake spillway which is more than 20 km to the south. This deformation is only about 10% of that inferred from the deformation of shoreline features of Glacial Lake Russell, which show a gradually increasing amount of deformation northward (fig. 19). This reconstruction also requires that the Shelton delta of Lake Russell and the Kent Lake spillway have been differentially uplifted by as much as 20 m, even though they are separated by less than 10 km. These apparent inconsistencies suggest that some alternate explanation for the Belfair delta exists.

The geologic setting of the Belfair delta supports, but does not prove, the alternative that it was built instead into Lake Hood. Meltwater channels and terraces north of the delta indicate that the glacier may have retreated as much as 3 km north of the Belfair delta prior to its deposition. If these inferences are correct, the Clifton channel would probably have been exposed and the proglacial lake would have already dropped to the level of Lake Hood.

The absence of a recognizable gradient between the north and south ends of the Clifton channel indicates that Lakes Hood and Russell were nearly equal in altitude during erosion of the outlet. A large difference in altitude between the lakes would have caused a steep gradient to develop in the channel as it crossed the narrow neck of land between the two lakes.

The pattern of deformation inferred from Lake Russell strongly supports the argument that Lakes Russell and Hood lay at equal altitudes. The altitudes of the north Auburn and Renton deltas (Lake Russell) and the inferred amount of deformation between them is identical, within the margin of error, to the altitudes of the Belfair and Eldon deltas (Lake Hood) and the slope between them (fig. 17). Therefore, during its northward extension, Lake Hood drained via the Clifton channel and lay at the level of Lake Russell. This equality of altitude suggests that the Clifton channel was either gradually eroding during its existence, or that it was originally submerged but eroded during its later stages. The apparent lack of a large volume of sediment at the mouth of the Clifton channel also suggests that it was initially submerged during deglaciation. Because Lakes Hood and Russell probably lay at nearly the same altitude the pattern of deformation inferred from Lake Hood and Lake Russell are comparable; both were controlled by the Black Lake spillway.

Lake Leland

The Brinnon delta was described by Bretz (1913) as a very large well-formed outwash delta that was built into Lake Hood when its surface altitude was 120 ft (36.5 m). Another delta on the west side of Hood Canal was discovered near the mouth of Fulton Creek about 5 km south of the Brinnon delta. Its topset-foreset contact lies at 29.3 m altitude. Both deltas apparently were built into open water, the reconstructed surface of which now lies below the present altitude of the Black Lake spillway (41.0 m).

It is unlikely that Brinnon and Fulton Creek deltas were built into a lake controlled by the Black Lake spillway and subsequently have been warped down relative to it. The pattern of deformation defined by shoreline features of Glacial Lakes Russell and Hood indicate that substantial postglacial uplift increases northward from the former glacier terminus, making downwarping to the north unlikely.

Because the Black Lake spillway was the lowest meltwater route to the ocean south of the Strait of Juan de Fuca, the presence of these deltas requires a body of water that drained north. As will be documented later, postglacial sea level never reached the altitude of these deltas; therefore, a north-draining proglacial lake must have existed. The Leland Creek spillway (68.6 m), which drained north between Quilcine Bay and Discovery Bay from Hood Canal to the Strait of Juan de Fuca, is the only known north-draining spillway that could have controlled the height of this lake (plate 1).

The Quilcine delta, which lies midway between the Brinnon delta and the Leland Creek spillway, was also probably built into Lake Leland. It is therefore included with the discussion of deposits of that lake.

Lower Fulton Creek Delta. The Lower Fulton Creek delta extends about 100 m north and south of Fulton Creek near its mouth on the west side of Hood Canal. On the southwest side of the creek a poorly defined bench extends several tens of meters inland from the delta front. On the north side of the creek, however, a prominent bench is absent, and the delta has no apparent surface expression. The sediments that make up the delta are exposed in roadcuts along Highway 101.

Sediment of the Fulton Creek delta is unlike any delta described above. Gravel consists almost entirely of locally derived subangular to subrounded lithic clasts. Only a few granitic clasts, perhaps reworked from older drift, were found in the entire exposure. Near the delta, bedrock is commonly exposed below the recessional sediments. A well-compacted deposit of silty gray till, that ranges in thickness from 0.5-10 m, overlies the bedrock at several nearby localities along the west side of Hood Canal. The topset-foreset contact occurs at the same height (29.3 m) on both sides of Fulton Creek. Owing to channeling of 50-100 cm relief, the contact was difficult to define exactly in many places.

Unlike the previously described deltas, the Lower Fulton Creek delta apparently was built largely by local tributary streams rather than by an ice-marginal stream peripheral to the Puget lobe. The coarse gravel which makes up the bulk of the deposit, suggests that the delta was built by a vigorously flowing stream that was aggrading its channel with local, presumably glacial, sediments.

Brinnon Delta. The Brinnon delta is one of the best-exposed and most accessible deltas in the entire Puget Sound region. It lies at the mouth of the Dosewallips River along the west shore of Hood Canal and is well exposed in a gravel pit along Highway 101. The delta forms a large flat bench about 1 km^2 on the north side of the river and clearly exhibits an arcuate front convex toward Hood Canal (plate 1). The delta must have originally extended across the entire valley floor but was subsequently eroded near the river mouth. From the delta surface prominent alpine outwash terraces related to alpine glaciation rise westward toward the Olympic Mountains (Frisken, 1959). The composition of the gravel, and the complete absence of granitic clasts suggests that the Brinnon delta was built completely from gravel derived from the Dosewallips drainage. The unchanneled topset-foreset contact (38.4 m) is well defined and horizontal for approximately 100 m of exposure.

The Brinnon delta was derived locally and was not built in contact with the Puget lobe. It represents extensive aggradation of the Dosewallips River valley during deglaciation of the Puget lowland. Alpine glaciers probably were the source of the large amount of locally derived sediment.

Quilcine Delta. The Quilcine delta, 3 km west of Quilcine along northern Hood Canal (plate 1), forms a prominent flat bench between the Quilcine Fish Hatchery and Washington State Highway 101 where the Big Quilcine River enters Quilcine Bay. The delta sediments can be observed where remnants of the delta fill are exposed by roadcuts on both sides of Highway 101. The topset-foreset contact (50.0 m) is very sharp, horizontal, and unchannelled.

The local provenance of the sediments, northeastward dip of foreset bedding, and imbrication direction indicate that the Quilcine delta was derived from the valley of the Big Quilcine River and built into Lake Leland. The coarse texture and the apparent steep northeast gradient of the delta top suggest that deposition was from a large alpine river probably fed by alpine glaciers.

Inferred deformation. The Lower Fulton Creek, Brinnon, and Quilcine deltas were built directly into Lake Leland which drained northward to Discovery Bay. All were built by local streams that drained Olympic valleys. The topset-foreset contacts of the Fulton Creek, Brinnon, and Quilcine deltas lie at 37.7 m, 28.6 m, and 17.0 m below the floor of the Leland Creek spillway, respectively, yet probably drained over it. This indicates that the amount of postglacial deformation gradually increases northward along Hood Canal in a nearly linear fashion (fig. 17). Because little apparent erosion of the long, bedrock-floored Leland Creek channel occurred, it can also be used as a point on the north-south profile of deformation.

The slope of the deformation profile defined by water-plane features associated with Lake Leland is nearly identical to that defined by the combined data from Lakes Russell and Hood (fig. 17). The similarity of the curves suggests that postglacial deformation follows a regional pattern.

Other Delta Deposits. Retreat of the Puget lobe as far north as Seattle would probably have exposed the Poulsbo channel of 47.0 m altitude (Bretz, 1913; plate 1). This 8-km long channel lies between Hood Canal and Liberty Bay (Puget Sound) between 2 and 10 km north of Poulsbo. It has a relatively constant width of 300-500 m, and is deeply incised into the Vashon till plain in this area. Its gradient is undetermined because of postglacial erosion and later filling, but must have been very low. The Poulsbo channel provided a low connection between proglacial lakes in the troughs of Hood Canal and Puget Sound during deglaciation of the Seattle area. Ponded proglacial meltwater in the central Puget lowland probably fell to the level of Lake Leland when the Poulsbo channel was exposed; therefore, the channel represents a minimum height for Lake Leland.

Redmond delta in the Sammamish Valley (Curran, 1965) and the Harris Creek delta in the Snoqualmie Valley (Anderson, 1965) were built into proglacial lakes in the Puget lowland. Like the Brinnon and Lower Fulton Creek deltas of Lake Leland, they lie at or below the altitude of the Black Lake spillway. It is unlikely that they were warped down or not uplifted relative to the Black Lake spillway because the pattern of deformation inferred from Lakes Leland, Russell, and Hood indicate that the Puget lowland near these latitudes was raised as much as 70 m relative to the Black Lake spillway. Therefore, probably the Redmond and Harris Creek deltas were built into Glacial Lakes Sammamish and Snohomish when these lakes drained either into Lake Leland or directly to the sea.

The lowest spillway of Glacial Lake Sammamish prior to entry of the sea was the Totem Lake channel, a poorly defined, broad connection between the Sammamish Valley and Lake Washington at 34.8 m altitude. If Glacial Lake Sammamish drained directly to the sea, deep erosion of the channel, especially near its western end, would be expected; yet there is only an extremely slight westward gradient. The channel floor is mantled with laminated apparently lacustrine silt and clay, also suggesting that Lake Leland extended into the Sammamish Valley. The Redmond delta, therefore, was probably built at the level of Lake Leland.

Redmond Delta. The Redmond delta, which lies just east of Redmond at the northern end of Sammamish Lake, is everywhere composed of loose sandy gray recessional outwash gravel derived from the Puget lobe. The delta was described by Bretz (1913) and later studied by Anderson (1965). Bretz indicated that the Redmond delta consisted of four different levels at 160 ft (49 m), 140 ft (43 m), 130 ft (40 m), and 120 ft (37 m), and inferred that the higher levels were being dissected while the lower ones were being created. The highest level is still well preserved and can be easily studied. The lower levels are densely kettled, and are largely destroyed by gravel-pit operations; hence their geologic relations are poorly known.

The western edge of the Redmond delta stands about 38 m above the floor of the Sammamish Valley and is slightly convex to the west. The gravel pit here exposes a complete sequence of delta facies that overlie the dense Vashon till. The topset-foreset contact here is also poorly defined, but appears to be approximately horizontal; it is well exposed at 40.9 m altitude at the north end of the delta, where it is level, unchannelled, and marked by a pronounced line of well-sorted cobbles and boulders. Toward the east, topsets become more lenticular and cross-bedded, and the topset-foreset contact becomes channelled, with about 1-1.5 m relief, and appears to average about 1.5 m lower in altitude than the highest delta level to the west. In this area, the original delta was apparently slightly incised and reworked by meltwater streams during glacial retreat.

The extensive upper portion of the Redmond delta was probably built directly into Glacial Lake Sammamish when it lay at the level of Lake Leland by northwest and west-flowing meltwater streams which were marginal to the Puget lobe. All lower levels apparently represent downcutting during retreat, as suggested originally by Bretz (1913).

Systematic downcutting and terracing of the Redmond delta requires a sustained westward flow of meltwater near the delta after retreat of the Puget lobe from the area. Such flow could only have been derived from the Redmond channel which drained from Glacial Lake Snohomish to the east. The Redmond channel (114.0 m) is a long, 300- to 1000-m-wide, relatively straight spillway which lies between the Snoqualmie and Sammamish valleys, and follows the valleys of Patterson and Evans Creeks. It controlled the height of Glacial Lake Snohomish until the Puget lobe withdrew north of Everett. If the gradient of the Redmond Channel was very low during terracing of the Redmond delta, as it apparently was during the final stages of lake drainage, Glacial Lakes Snohomish and Sammamish would have existed at nearly the same altitude. If these conditions prevailed during deposition of the Harris Creek delta in the Snoqualmie Valley, then its topset-foreset contact defines the level of Lake Leland as projected into the Snoqualmie Valley. The Harris Creek delta topset-foreset contact would, however, provide a maximum estimate for the height of such a projection.

Harris Creek Delta. The Harris Creek delta lies on the east side of the Snoqualmie Valley 4 km north of Carnation (plate 1). It is exposed in a shallow gravel excavation along the Snoqualmie Valley Road on the north side of Harris Creek. The delta appears to have been built by streams that drained Harris Creek, and at one time plugged the entire valley mouth. The Puget lobe must have been north of the delta at that time, for no ice-contact features are present within the exposed delta sediments.

It cannot be determined whether a contact between foreset beds and overlying flat-bedded gravel at 32.0 m is the true topset-foreset contact, or whether the upper gravels were deposited during downcutting of the main surface. The observed contact, therefore, represents only a minimum height for the true topset-foreset contact. The delta surface (42.4 m) represents the maximum possible height.

Knoll (1967) reported a higher level of the Harris Creek delta and stated that the main altitude of the topset beds was about 300 ft (92 m). No evidence for the 92-m level was found during the present investigation. The following sequence was observed topographically above, but stratigraphically below, the Harris Creek delta; overlying a 50 -m-thick gray till, loose horizontally bedded gravels at about 46 m altitude grade upward and form a prominent bouldery recessional outwash terrace at about 92 m. Although in some localities the gravel is mainly massive sand, foresets were nowhere exposed. This sequence is interpreted as recessional outwash that overlies Vashon till. The lack of ice-contact stratified sediment in the Harris Creek delta indicates that it could have formed long after formation of the higher terraces.

Richmond Beach Delta. The Richmond Beach delta lies on the east side of Admiralty Inlet about 5 km south of Edmonds (plate I). It forms part of a clearly defined terrace at 59 m altitude that contains a few large kettles east of the present shoreline. Because the delta now lies entirely within a county park, exposures are poor and sparse. Topsets can be traced down to an altitude of about 58 m, and foresets can be traced up to about 54 m. The actual contact is not exposed, but probably lies at about 55.0 ± 2.0 m.

If the Puget lobe lay at or north of Richmond Beach the Poulsbo channel would have been exposed; hence, Lake Leland would have been able to expand eastward into Admiralty Inlet. Therefore, the Richmond Beach delta probably was built into Lake Leland.

The sediment source for the Richmond Beach delta was the ice-marginal meltwater streams that flowed southwest across the drift plain north of Seattle. The presence of kettles suggests that blocks of ice must have been buried during delta deposition. No other evidence for the nature of the ice margin during delta construction was found.

Inferred deformation. The Richmond Beach (55.0 ± 2 m), Redmond (40.9 m), and Harris Creek (32.0 - 42.4 m) deltas show a general west-to-east decrease in the inferred height of Lake Leland. The Totem Lake (>38.4 m) and Redmond channels (>34.6 m), provide minimum estimates for the height of Lake Leland and support the estimates derived from the delta sediments. The topset-foreset contact of the Richmond Beach delta also lies above the level of the Poulsbo channel (47.0 m) and above the topset-foreset contact of the Quilcine Delta (50.0 m). These data, combined with data from Glacial Lakes Sammamish and Snohomish, indicate that the inferred surface of Lake Leland reaches a maximum in the center of the Puget lowland, and decreases both west and east towards its margins.

Marine-Limit Data

Relative Sea Level

Relative sea level is here defined as the altitude of sea level at any specific location and time relative to modern sea level. The marine limit, here defined as the highest altitude at which post-glacial-marine features or sediments occur, is coincident with the highest position of relative sea level. The marine limit can be approximately reconstructed for the northern Puget lowland by mapping the distribution and altitude of glacial marine drift, meltwater channels, and marine deltas (fig. 19; plate 1).

Deglaciation of the northern Puget lowland about 13,000 years ago apparently was extremely rapid because incursion of marine waters caused rapid melting and collapse of the Puget lobe. Although global (eustatic) sea level was rising at this time, perhaps on the order of about 1 cm/yr (Curray, 1965), deglaciation may have been so rapid that nearly the same relative sea level existed across the northern lowland. The maximum height of relative sea level (marine limit) in the northern Puget lowland can therefore be used as a nearly isochronous water-plane.

Glacial-marine Drift

A minimum estimate for the marine limit during deglaciation may be determined from the altitude of marine sediments which now lie above sea level (fig. 20; table 4). At four localities within the northern Puget lowland marine pelycypod shells have been found *in situ* within muddy stony sediments (Easterbrook, 1968; Gayer, 1977). The fossiliferous sediments generally gradationally overlie the Vashion till and exhibit many characteristics similar to it, including a silty matrix, wide range of grain sizes, and striated stones. The drift is therefore interpreted as glacial debris which was deposited below sea level.

Glacial-marine sediment typically is relatively massive, but locally contains bedded clayey-silts and silty-clays that commonly contain isolated large clasts or gravel lenses that are dispersed in the fine-grained sediments. The drift may contain unreworked marine fossils in its unoxidized portions; where oxidized it typically is brownish-gray, and has well-developed columnar jointing and blocky fracture.

Similar deposits which overlie Vashon drift occur widely across the northern Puget lowland between the Miller Peninsula and the Stillaguamish River. Although marine fossils are not present at these localities, the sediments are also interpreted as glacial-marine in origin. Preliminary data on exchangeable sodium content (Pevear and Thorson, 1978) support a marine origin for the nonfossiliferous sediments. To avoid confusing inferred glacial marine-drift with glacial-lake deposits, only those areas where no proglacial lakes are known to have existed were studied.

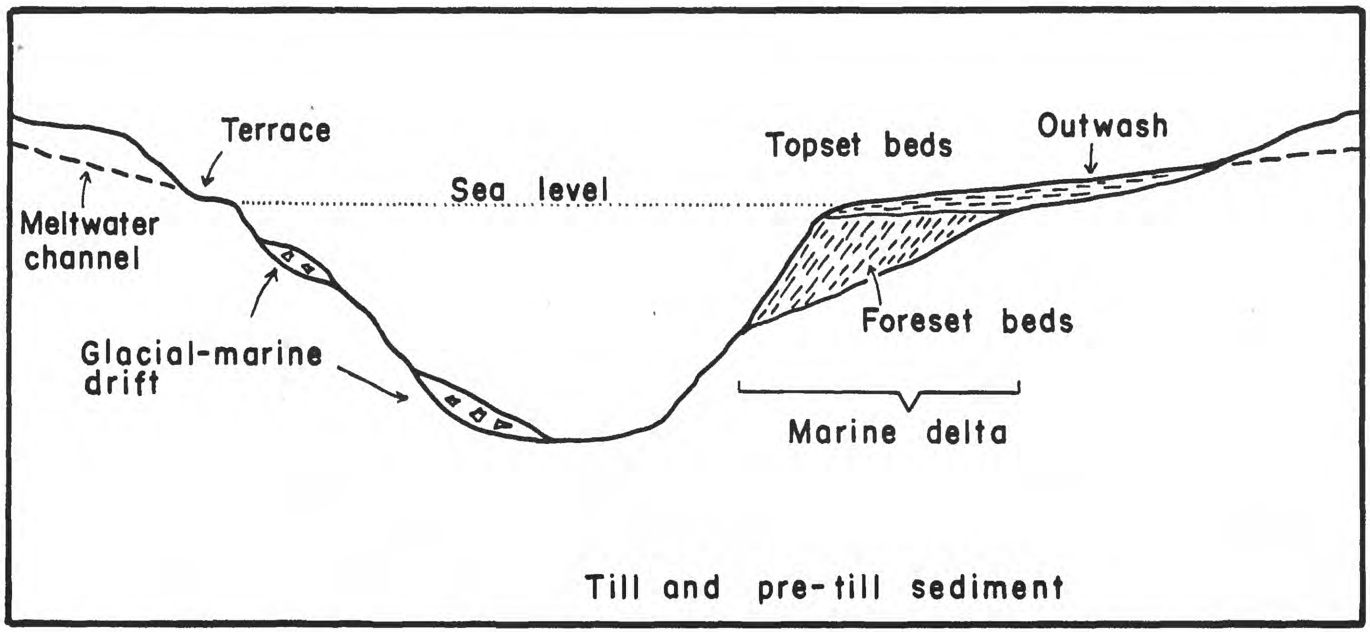

Figure 19.--Geologic features used to reconstruct position of relative sea level in the northern Puget lowland. Relative sea level lies below features formed by fluvial erosion (channels) and deposition (outwash), yet above the position of glacial-marine sediments. Topset-foreset contact of marine delta lies slightly below actual position of relative sea level.

Figure 20.--Typical appearance of pebbly glacial marine silt in northern part of study area. This exposure is on the north shore of the Quimper Peninsula west of Point Wilson. Columnar jointing is characteristic (compare with Figure 13). Digging tool in the center of photograph is about 40 cm long.

TABLE 4.-- EVERSON-AGE DEPOSITS IN THE NORTHERN PUGET LOWLAND INFERRED TO BE GLACIAL-MARINE IN ORIGIN.

Location	Marine Pelycypod Shells	Maximum Altitude (m)
Protection Island	Yes (Gayer, 1976)	10
Quimper Peninsula		
Four Corners	No	41
Cape George	No	35
McUrdy Point	No	15
West Beach	No	18
Fort Worden	No	8
Kala Point	No	31
Marrowstone Island		
Nordland	No	5
Whidbey Island		
Penn Cove	Yes (Easterbrook, 1968)	6
Rodena Beach	No	31
Fort Casey	No	52
Greenbank	No	37
Camano Island		
Brown Point	No	39
Juniper Beach	Yes	32
Triangle Cove	Fragments	5
Indian Beach	No	42
Mabana	No	40
Mainland		
Tulalip Shores	No	20
Mission Beach	No	21

Meltwater Channels

Numerous flat-bottomed meltwater channels cross the narrow upland portions of the northern part of the Puget lowland between Sequim Bay and Stanwood (plate 1). These channels are generally <200 m wide, have nearly constant slope, and commonly appear graded to terraces above the modern beach cliffs. The mouths of the channels occur no lower than about 30 m altitude. Because erosion by meltwater does not extend far below local base level, and because all recessional lakes are inferred to have existed above the maximum relative sea level, these channels provide an upper limit for the altitude of the marine limit (plate 1).

An individual channel could be accounted for in several ways, such as subglacial drainage or submarine density flows, and might not represent the position of sea level at a specific point. If a population of channels shows a consistent pattern of minimum altitudes, then they can be used to estimate the maximum altitude of the marine limit. All channels were mapped at a scale of 1:24,000 and then transferred to a 1:250,000-scale map so that they could be compared on a regional basis (plate 1).

The channels are trough-shaped in cross section but exhibit great variability in their long profiles. They can be grouped into two representative types (fig. 21). Type-A channels have a smoothly graded long profile that steepens abruptly near the present shoreline. All such channel mouths have been incised by postglacial streams and (or) destroyed by beach cliff recession. The lowest contour line that crosses the channel floors was used as an estimate of maximum relative sea level. The Type-A channel may have formerly extended beyond and below its present lower limit. Type-B channels show a smoothly graded long profile that abruptly terminates at a broad terrace or bench parallel to the present shoreline. This parallelism, and the absence of meltwater erosion below them, suggests that the channel mouths and terrace surfaces may mark the actual relative sea level during deglaciation. Type-B meltwater channels therefore represent a maximum altitude of the marine limit; probably a close maximum.

Meltwater channels used to estimate the maximum height of the marine limit are best developed on both shores of Whidbey Island south of Greenbank, and on the west shore of Camano Island (plate 1). Type-B channels on the east side of Whidbey Island terminate on local terraces at a rather consistent altitude of 42.7-48.8 m (140-160 ft). On the west shore of Whidbey Island the altitude of Type-A channel mouths are more variable, but occur at about 54.8-61.0 m (180-200 ft). Along the west shore of Camano Island north of Indian Beach, the Type-B channel mouths generally occur no lower than 54.8-61.0 m (180-200 ft) and apparently increase in altitude northward.

Several well-defined channels on the Miller Peninsula are 10-50 m in width and consistently extend down to about 54.8 m (180 ft). On the Quimper Peninsula near Cape George, opposite Discovery Bay, several well-defined meltwater channels lie at similar altitudes. Both channel types are also present on the mainland near Tulalip where they range down to altitudes of about 54.8-67.1 m (180-220 ft), but can occur as high as 91.5 m (300 ft). Several clearly defined Type-A channels

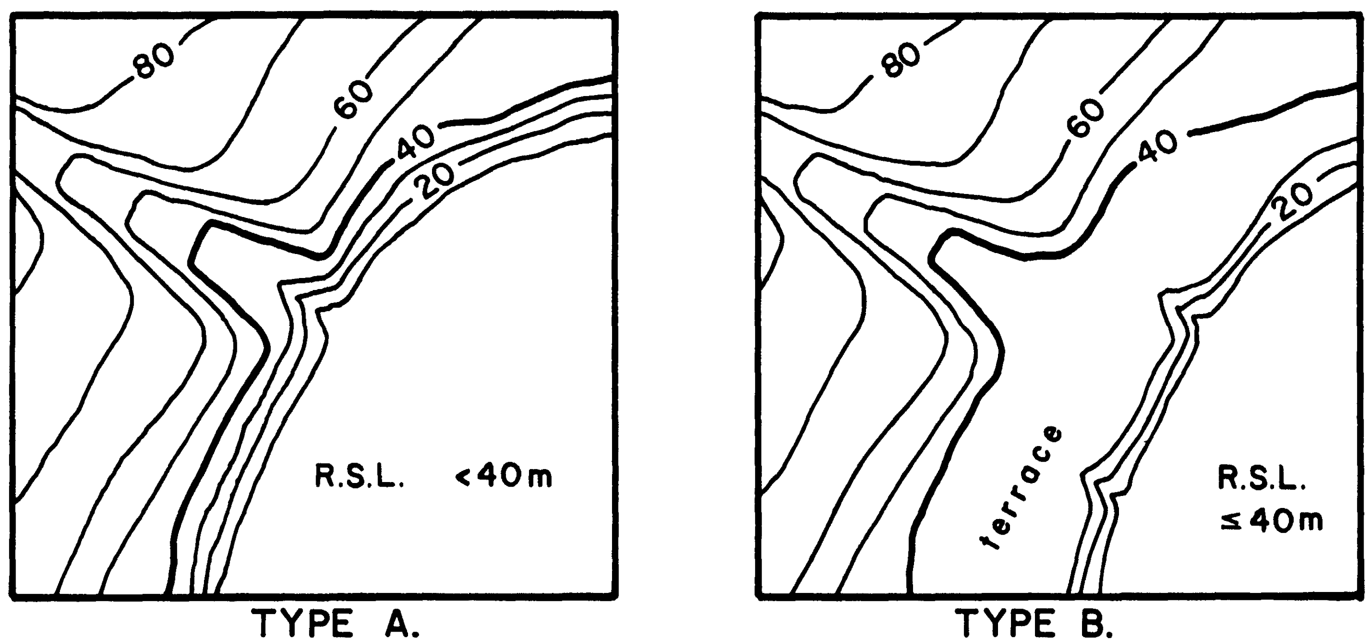

Figure 21.--Contour maps showing types of meltwater channels common in the northern Puget lowland. Maximum height of relative sea level (R.S.L.) can be inferred from the lowest contour line that shows channel morphology (heavy line).

on the Hansville Peninsula near Foulweather Bluff extend as low as 36.6 m (120 ft) but show considerable scatter in their lowest altitudes. No well-defined channels at or below this altitude were observed south of Port Gamble. Higher channels, however, become common, especially toward the southern Puget lowland. The lowest channel altitudes in this area were probably controlled by the proglacial lakes described previously. Although they represent maximum estimates for the height of relative sea level, they were not mapped to avoid confusion with channels thought to be controlled by relative sea level.

Marine Deltas

Gardiner Delta. A well-defined outwash delta occurs near the west shore of Discovery Bay about 3 km east of Gardiner (plate 1; fig. 9). The delta surface is about 200 m wide and flat topped near the bay, but becomes more poorly defined inland. The slightly arcuate delta front is convex toward the north and northeast. The sediment of the Gardiner delta, well exposed by gravel-pit operations, consists entirely of light-colored, fairly well-sorted loose sandy pebble gravel that contains a high percentage of granitic clasts of northern provenance. The topset-foreset contact, which has a maximum altitude of 40.2 m and is exposed for only about 12 m, is well-defined and relatively unchannelled, but has about 1.5 m relief that cannot be attributed entirely to slumping or channelling. Relief on this contact is expressed as variations of water level, which may have been partially caused by fluctuations in the former tides.

The dominance of clasts of northern provenance, the northward inclination of the foreset beds, the presence of silt- and till-rip-up clasts and sediment deformation, and the absence of a valley wall to the east suggest that the Gardiner delta was built by a north-flowing stream along the margin of the retreating Puget lobe.

The Gardiner delta was built into water occupying Discovery Bay. A proglacial lake could be impounded in Discovery Bay during glacier retreat if the bay mouth was dammed at Diamond Point; the lake would have been forced to spill out through Eagle Creek valley to Sequim Bay. The orientation of meltwater channels on the Miller Peninsula, however, suggests that Diamond Point probably was cleared of ice prior to deglaciation of the Gardiner delta area, and there is no apparent lake spillway cut into the Eagle Creek drainage. It is therefore unlikely that the Gardiner delta was built into a proglacial lake. Rather, it probably built directly into marine water that extended into Discovery Bay during glacier retreat. The topset-foreset contact in the Gardiner delta therefore probably marks the actual height of local sea level during deglaciation. Relief on the topset-foreset contact may be due either to differences in the strength of currents or to tidal variations during deposition. The absence of similar relief where the topset-foreset contact is relatively unchannelled in the other lacustrine deltas supports this hypothesis.

Hadlock Delta. The Hadlock delta covers more than 1 km^2 near Hadlock (plate 1; fig. 9). The delta has a flat top that slopes gradually northward except where locally interrupted by kettled terrain, and a steep front convex to the north.

The west shore of Port Townsend and a large gravel pit west of Hadlock provide excellent exposures of the delta structure and sediment. The delta consists completely of light-colored gravel deposited by Puget lobe meltwater. The poorly exposed topset-foreset contact appears to be nearly horizontal and extensively channelled, with at least 1 m relief. It lies at a maximum altitude of 35.5 m.

The Hadlock delta was built north and east into Port Townsend by north-draining meltwater of the Puget lobe. Because of the inferred northwest orientation of the retreating glacier margin, a proglacial lake probably did not exist in Port Townsend north of Hadlock. The topset-foreset contact of the Hadlock delta therefore probably represents the height of local sea level during deglaciation.

Stanwood Area. An extensive terrace about 3 km long and 2 km wide at the mouth of the Stillaguamish River near East Stanwood (plate 1) is remarkably flat, with a constant altitude of about 36.6 m (120 ft). Thinly bedded and laminated fine sand, silt, and clayey silt mantle the surface; two well logs indicate that fine sand, probably the same deposit, extends to a depth of about 7.6 m (Newcomb, 1951). These deposits are similar to modern sediments of the Stillaguamish River near its mouth. Because of their fine grain size, their laminated appearance, and a lack of any type of deformation, these sediments are inferred to be estuarine and (or) alluvial deposits that accumulated on the terrace after deglaciation.

A large isolated outwash terrace that extends no lower than 61 m altitude and formed during ice recession lies east and north of the Stillaguamish River deposits described above. Relative sea level during deglaciation of the Stanwood area must have been between the altitude of the estuarine (?) surface and of the lowest outwash sediments.

Inferred Marine Limit

The altitude of the marine limit in the northern Puget lowland can be bracketed between the minimum altitudes of meltwater channels and the maximum altitude of glacial-marine drift. Where the meltwater channels terminate in coastal terraces they may give a close maximum limiting altitude for the marine limit. The two marine deltas provide altitudes for the marine limit that fall within the estimated range at their localities. The marine limit can be defined approximately with the available data (plate 1; fig. 22).

Localities having both glacial-marine drift and meltwater channels can be used to bracket the marine limit as specific points (fig. 17). The upper limit is based on an average of 3 to 5 altitudes of the mouths of meltwater channels. The lower limit is based on localities where glacial-marine drift occurs well above sea level. The inferred marine limit is drawn near the upper limit of the range because the channels probably represent a more closely limiting estimate. The marine limit is inferred to rise gradually northward in altitude from about 42.5 m (140 ft) on south Whidbey Island to 73.2 m (240 ft) on northwest Camano Island.

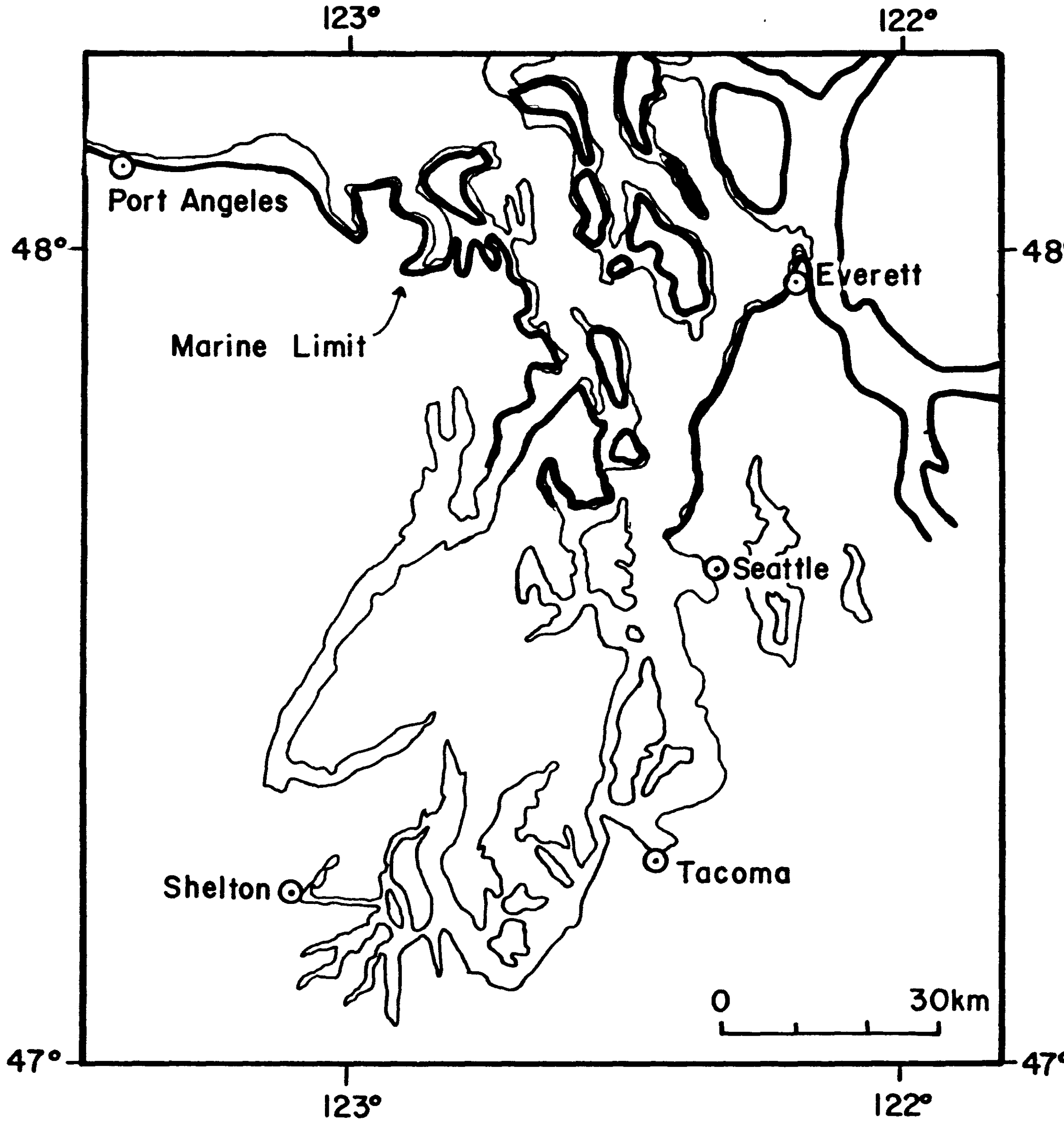

Figure 22.--Map showing inferred location of Everson-age marine limit in the northern Puget lowland. The Everson-age marine limit south of Seattle lies below the present shoreline.

Global (Eustatic) Sea Level

The amount and pattern of postglacial deformation in the northern Puget lowland can be reconstructed from the inferred marine limit only if the position of eustatic sea level at the time of deglaciation is known. Eustatic sea level sensu stricto is almost impossible to determine accurately because the earth's surface is constantly undergoing differential elevation change in response to variations in ice and water loads (Bloom, 1970; Clark and others, 1978). More data on the position of eustatic sea level following the last glaciation exists for the continental margin of the eastern United States than for most other areas. Macintyre and others (1977) have interpreted these data to support the eustatic sea-level curve of Curray (1965). Their data indicate that along the east coast of the United States eustatic sea level was about 62 m below present sea level about 13,000 years ago.

This estimate for eustatic sea level at 13,000 yr B.P. can be used for the northern Puget lowland only if a wide margin of error is applied. The postglacial isostatic response of the eastern U.S. to changing ice and water loads has affected Curray's estimate for the position of eustatic sea level to some degree. Because the Puget lowland experienced different paleogeographic conditions and isostatic response, eustatic sea level derived for the eastern U.S. at that time is not strictly applicable to the northern Puget lowland. The amount of gravitational attraction of the Pacific Ocean by the Cordilleran Ice Sheet during deglaciation was probably small and cannot be defined precisely, but probably it was at least slightly different from such effects in the eastern U.S. In addition, the age of deglaciation of the northern Puget lowland is estimated at about 13,000 years ago; however, an error of several hundred years would cause considerable variation in the estimate of eustatic sea level. Owing to the problems outlined above, the position of eustatic sea level during deglaciation of the northern Puget lowland probably cannot be specified to within less than about 10 m. This error margin must be applied to the reconstruction of postglacial deformation in the lowland based on marine-limit data.

Pattern of Postglacial Deformation

Integration of Lacustrine and Marine Data

A major objective of this study is to document the pattern of postglacial deformation in the Puget lowland. Proglacial lacustrine features and raised marine features provide independently derived and mutually supporting data that accurately define the pattern of postglacial deformation (figs. 23, 24, and 25). In order to compare these two types of data, however, the lacustrine data must first be integrated with the marine-limit data.

All of the deltas south of Fulton Creek were built into Lakes Russell and Hood, both of which were controlled by the Black Lake spillway (41.0 m) during their entire history. The difference in altitude between each delta topset-foreset

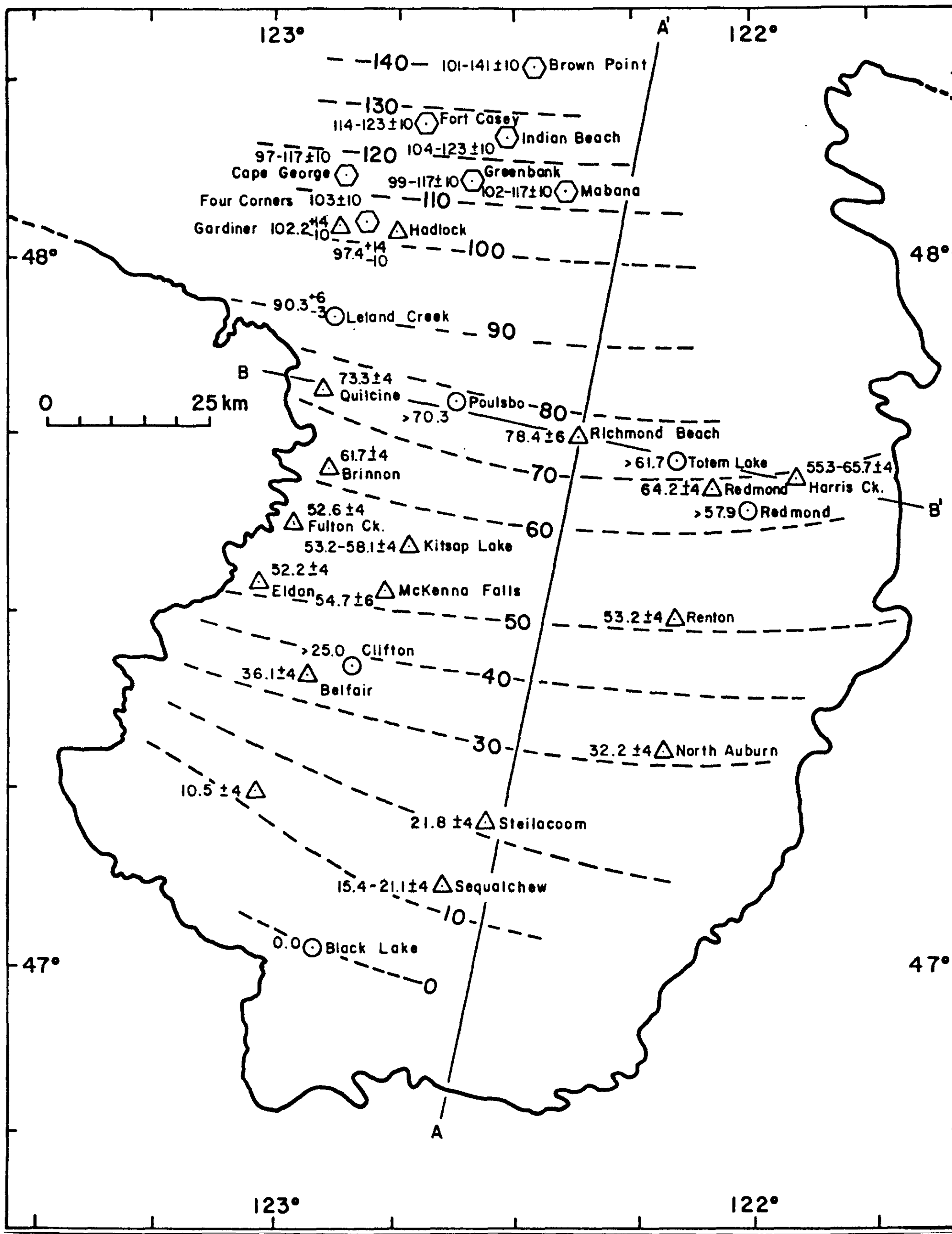

Figure 23.--Contour map showing inferred postglacial deformation in the Puget lowland. Contour interval is 10 m. Circles, triangles, and hexagons represent lake spillways, deltas, and localities where marine features allow the marine limit altitude to be estimated, respectively. Numbers indicate the amount of inferred deformation (m). Lines A-A' and B-B' represent lines of projection parallel to and normal to the axis of the lowland, respectively.

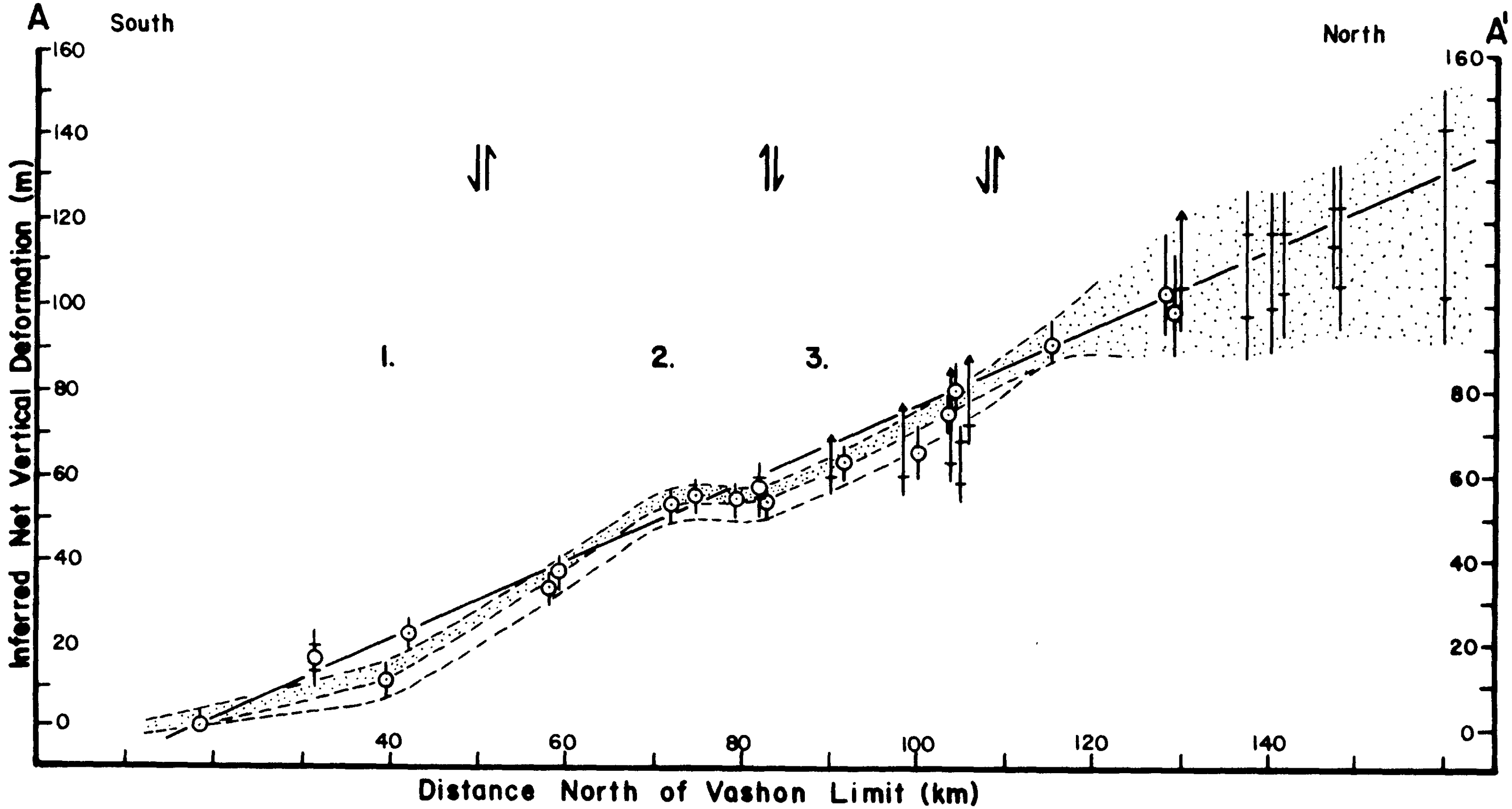

Figure 24.--North-south profile of inferred net postglacial deformation parallel to the axis of the Puget lowland along line A-A' (fig. 23). Circles and vertical bars indicate the amount of and range of deformation, respectively. Vertical lines indicate margin of error. South-sloping solid line indicates mean slope of deformation (deformation line) defined by all of the data points. Total width and stippled width between the dashed lines indicate possible and probable margins of error, respectively, for points from the western side of the Puget lowland. Numbers 1., 2., and 3. indicate the location of inflections where the probable range of deformation values depart significantly from the deformation line. Paired vertical arrows near the top of the diagram indicate the location and inferred structural movement along boundaries of major crustal blocks in the Puget lowland (fig. 29).

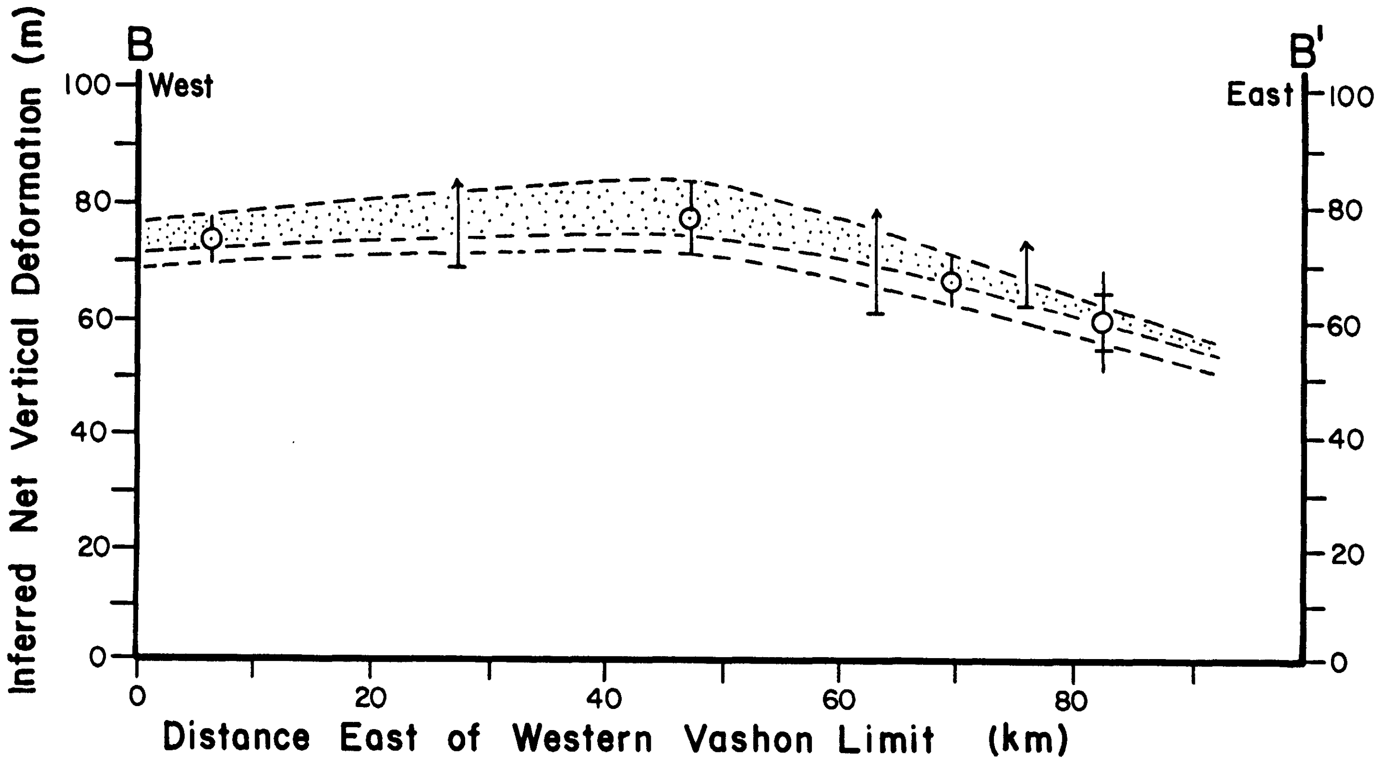

Figure 25.--East-west profile of inferred postglacial deformation plotted normal to the axis of the Puget lowland along line B-B' (fig. 23). Circles and vertical bars indicate the amount of and range of deformation, respectively. Vertical lines indicate the margin of error. Total width and stippled width of dashed lines indicate possible and probable margin of error, respectively.

contact and the Black Lake spillway is the amount of net vertical deformation that has occurred at each point relative to the Black Lake spillway. The regional pattern of deformation defined by these points is one of generally northward-increasing uplift.

The warping of features associated with or controlled by the level of Lake Leland indicates that this regional pattern of deformation continued north of the Lake Russell deltas. At Fulton Creek two deltas exist at greatly different levels (plate I; fig. 17). The higher earlier delta was built into the south-draining Lake Russell, and the lower delta was built into Lake Leland when it occupied the troughs of Hood Canal and drained northward. The difference in altitude (64.3 m) between levels does not reflect a differential uplift because no differential uplift can occur at a single geographic point; rather, the difference indicates that Lake Russell lay 64.3 m above Lake Leland near Fulton Creek. If this difference in altitude is added to all of the water-plane features of Lake Leland, they can be used to extend the pattern of deformation north as far as the Leland Creek spillway (figs. 23 and 24). The deltas and spillways of Lakes Russell, Hood, and Leland are thus integrated to a common datum; they define the relative amount of postglacial deformation from Black Lake to near Discovery Bay.

Although the precision is lower, marine features of the northern Puget lowland can be used to define the actual amount of postglacial deformation that has occurred. The difference between the present altitude of the marine limit (plate I; fig. 17) and the altitude of the inferred contemporaneous eustatic sea level (a negative value) represents the actual amount of deformation that has occurred (figs. 23 and 24). The two independently derived types of data, marine and lacustrine, coincide within the limits of error, thereby supporting the interpretations based on each. The coincidence also indicates that the relative amounts of deformation inferred from lacustrine data approximates their actual or absolute value. When integrated, it is possible to present all uplift data from the Black Lake spillway to northern Camano Island as absolute values (figs. 23 and 24).

Potential Error

The potential error of these methods of reconstruction must be known in order to interpret properly the deformation pattern. The altitude of a proglacial lake that drains through a certain spillway cannot usually be accurately known because neither the amount of spillway erosion nor the depth of the water in the outlet channel are known. The geologic relations near the Black Lake spillway, however, indicate that erosion and water depth together account for less than 9 m, and probably less than 4 m, error. The apparently shallow water depth inferred for the Black Lake spillway (<4 m) seems reasonable because the channel was everywhere at least 200 m wide. The maximum postglacial channel erosion (<4 m) also appears reasonable because the spillway has an extremely low gradient and was cut partially into bedrock. The surface of Lake Russell at the crest of the spillway, therefore, probably remained above 41 m, but below 45 m, during its entire existence. The relative deformation between each delta and the Black Lake spillway was, therefore, reduced by this margin of error (4 m; figs. 23 and 24).

Because the Leland Creek spillway occupies a broad preglacial valley, the amount of spillway erosion and water depth cannot be exactly determined. Just south of the spillway crest, recessional ice-contact stratified drift extends down to an altitude of about 76 m. Because the surface of Lake Leland must have been below this level, the sum of spillway erosion and water depth probably accounts for less than 8 m of error above the measured altitude of the valley floor. The presence of alluvial fans and swamp deposits on the valley floor indicates that the true base of the spillway may lie as much as several meters below the measured altitude. Because the Leland Creek spillway is cut largely into bedrock and was occupied for a much shorter length of time than the Black Lake spillway, the more closely limiting error (4 m) determined for erosion and water depth of the Black Lake spillway was used.

The error associated with using the topset-foreset contact of a coarse outwash delta as an estimate of the water-plane altitude is also <4 m. This error occurs in the same direction as that introduced by spillway erosion; hence it does not further reduce the relative deformation between deltas and spillways. It does, however, introduce a relative error of <4 m between any two deltas.

The error associated with the determination of the marine limit is much greater than for the lacustrine features. In most cases only gross minimum and maximum estimates are known. Marine deltas carry the same error as the lacustrine deltas, but at least an additional 3- to 5-m error is introduced by unknown variations in former tidal ranges. When the inferred position of eustatic sea level is used to determine the amount of deformation that has occurred, an additional uncertainty of about 10 m is introduced.

The difference between the altitudes of water-plane features as measured with surveying instruments and their actual altitudes (measurement error) is probably less than 30 cm for all deltas and spillways. These measurement errors were introduced largely by the imprecision in relocating the local bench marks. The non-deltaic marine-limit features carry a much higher measurement error of about ± 2 m.

Inferred Pattern

The dominant pattern of postglacial deformation in the Puget lowland is one of nearly steady increase in uplift northward (figs. 23 and 25), with a mean slope of 0.92 m/km (4.9 ft/mi). A high correlation (r = 0.97) exists between the inferred amount of deformation and the distance north of the Vashon limit. Possibly the regional slope of the deformation profile (fig. 24) is slightly concave upward near the northern part of the study area. The margin of error that must be applied there (± 14 m), however, is too great to allow this observation to be verified.

At the latitude of Seattle there are enough control points to infer the pattern of deformation across the Puget lowland perpendicular to the regional north-south slope (fig. 25). Although the geologic relationships on which this plot is based are less well known and the margin of error is greater, a consistent east-west regional pattern across the lowland also can be observed. The greatest uplift appears to have been in the center of the Puget lowland, with decreasing amounts eastward and westward from the center.

The combination of imprecise but actual values of deformation (marine data) with the more precise but relative amounts of deformation (lacustrine data) suggests that no significant deformation appears to have occurred south of the Black Lake spillway. If the maximum error involved in the reconstruction of the marine limit is used, the southern limit of deformation cannot be extended south of the drift limit more than about 20 km.

Uplift apparently was greatest in the center of northern Puget lowland. Significant uplift, however, also occurred very near the Cascade Range and Olympic Mountain fronts (fig. 23). The absence of widely recognized postglacial movement along faults and folds parallel to the mountain fronts in the northern Puget lowland indicates that the regional pattern of deformation extended well beyond the margin of the lowland.

The relatively narrow margin of error for the lacustrine control points permits the identification of possible inflections (deviations) from the north-south regional slope of inferred deformation (deformation line). To minimize the error in projecting all the control points to a single north-south profile, only those points from the western Puget lowland were used to identify possible inflections. Three significant inflections from the deformation line were observed in the southern Puget lowland at about 40 km, 70 km, and 90 km north of the glacier terminus (fig. 24). In these areas the probable range in values for the reconstructed lake surface deviated from the mean values by as much as 5 m.

CAUSES OF POSTGLACIAL DEFORMATION

Dominant Pattern

It was assumed throughout this investigation that the major cause of observable postglacial deformation in the Puget lowland was a return to isostatic equilibrium following deglaciation. Glaciation of the region by the Cordilleran Ice Sheet should have caused a large positive mass imbalance. Isostatic compensation for this imbalance should have occurred during the advance and maximum stand of the Vashon-age Puget lobe. After rapid deglaciation of the area a negative isostatic anomaly should have existed. To investigate and verify these assumptions it is necessary to determine the distribution of the mass imbalance caused by glaciation of the Puget lowland. The pattern of mass imbalance can then be compared to the pattern of postglacial deformation.

The changes in mass caused by Vashon-age glaciation are impossible to determine precisely because many mass changes other than the ice load were involved. Sediments were brought into and were eroded and transported from the Puget lowland, and a certain amount of sediment was deposited as glacial drift. The net change in mass caused by redistribution of the surface sediments is unknown, but probably small. Large proglacial lakes occupied the troughs of the Puget lowland during the advance and retreat of the Puget lobe; the water volumes involved and the duration of the impounded lakes would also have affected the pattern of loading. However, the total maximum estimated volume of the proglacial lakes probably never equaled more than several percent of the volume of the Puget lobe. Because proglacial lakes existed during both advance and retreat phases, and because the net erosion or deposition of sediment was relatively small compared to the mass of the Puget lobe, these effects are considered negligible.

The volume of the Puget lobe can be derived from a carefully constructed isopach map (fig. 26). First, the present topography of the Puget lowland was generalized to within 30 m (100 ft). The contour map of the ice-sheet surface (fig. 6) was then superimposed on the generalized topographic map. Ice thicknesses were determined where the contour lines from both maps intersected. These thicknesses were contoured, resulting in an isopach map that was then smoothed to eliminate local irregularities (fig. 26).

The mass of ice within each isopach interval can be calculated as the product of ice volume times the density of glacier ice ($0.917 g/cm^3$ at 0^oC). The total mass of the Puget lobe was calculated to be 8.9×10^{18} gm. The mass increased steadily northward from the glacier terminus, but little ice lay south of the Black Lake spillway. A regression of ice-sheet mass vs. distance north of the terminus has a very high correlation coefficient ($r = 0.99$; fig. 27).

The generalized mass distribution of the Puget lobe can be compared to the pattern of postglacial deformation in the Puget lowland. The very strong correlation ($r = 0.97$) between the mass distribution south of a given point and the

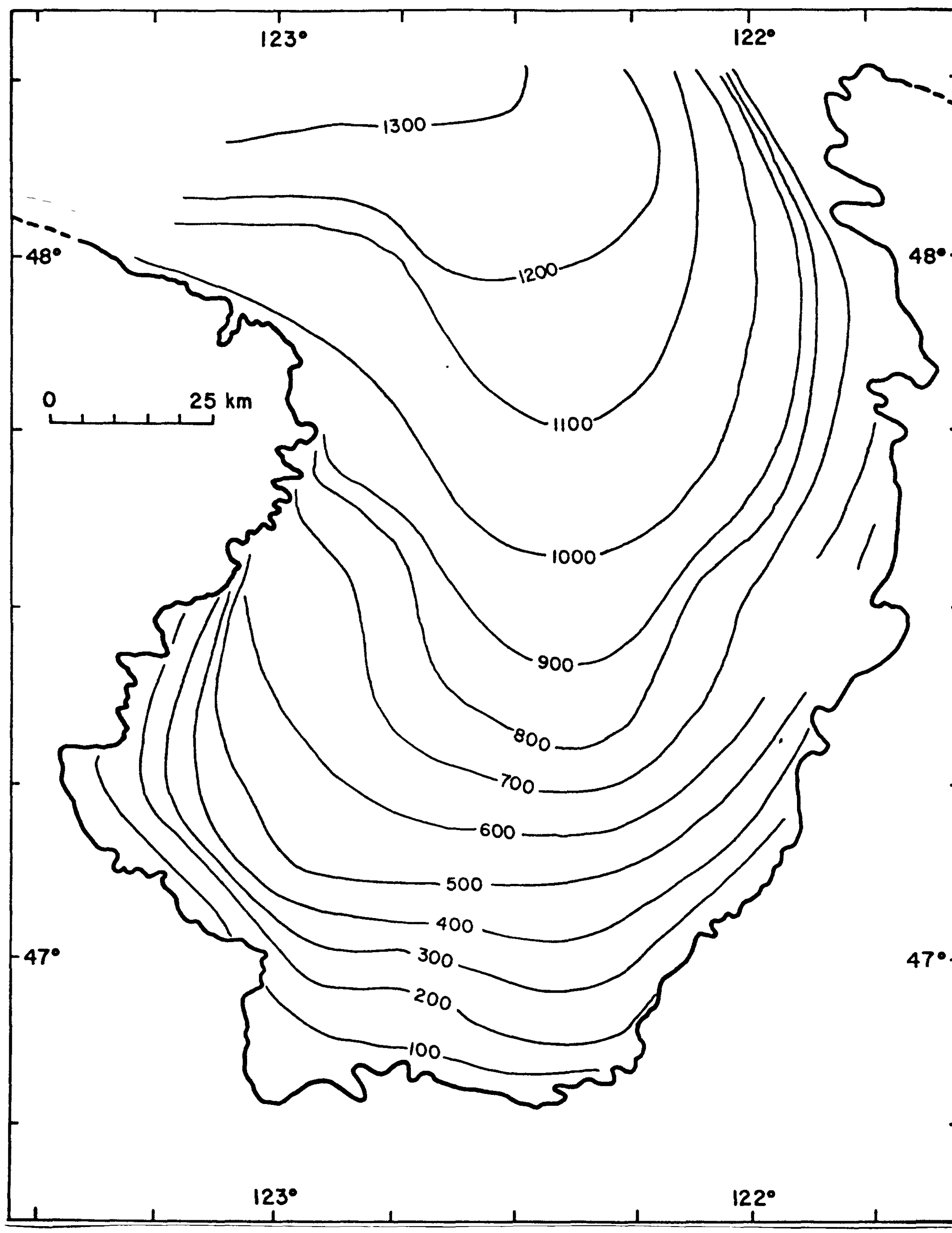

Figure 26.--Smoothed isopach map of the Puget lobe. Contour interval is 100 m.

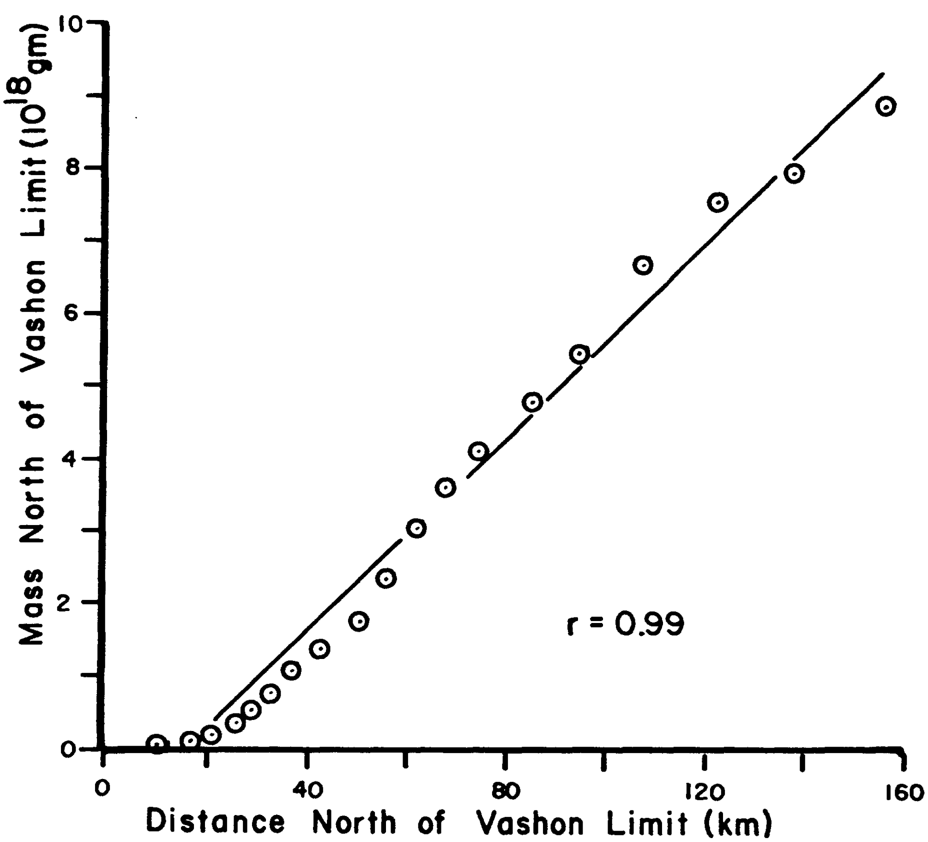

Figure 27.--Relation between the inferred mass of the Puget lobe and the distance north of the Vashon limit.

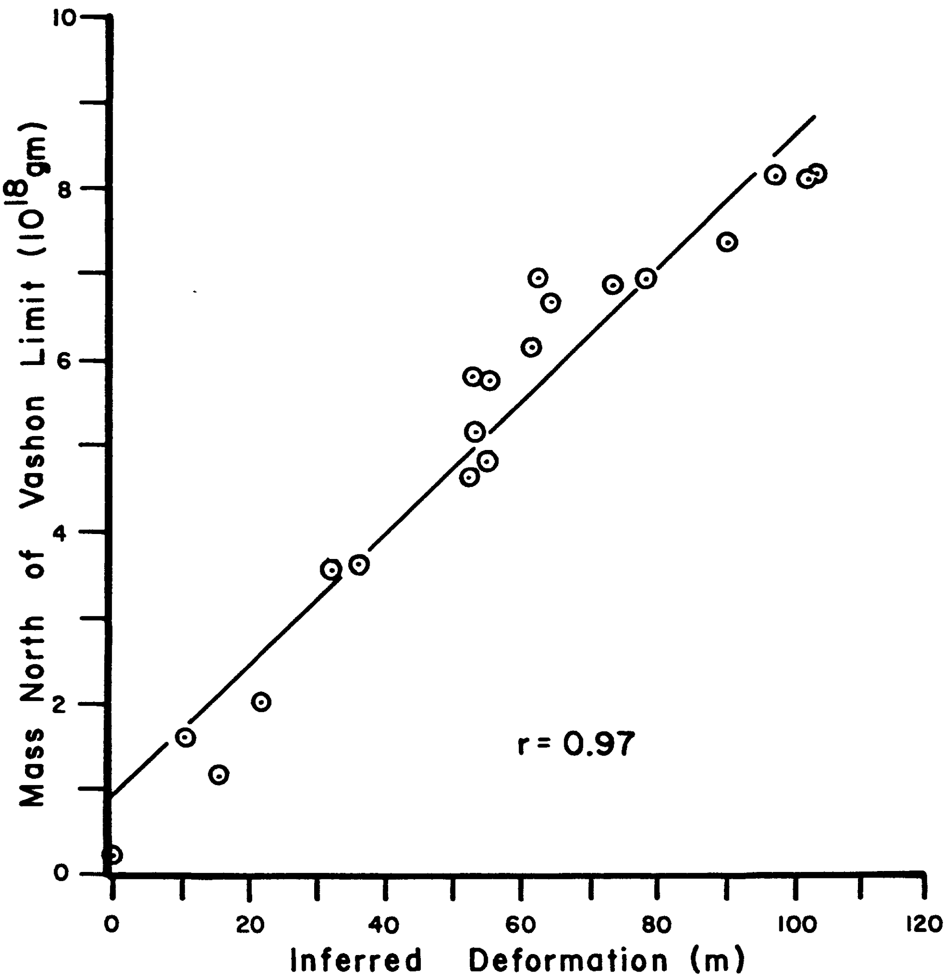

Figure 28.--Relationship between inferred postglacial deformation and inferred mass of the Puget lobe south of control point.

postglacial deformation inferred for that locality, indicates that these parameters are closely related (fig. 28). The absence of factors other than the effective rigidity of the crust that may link these variables, indicates that the mass of the Puget lobe over a point and the inferred deformation are related causally. Isostaticresponse to Vashon-age glaciation of the Puget lowland, therefore, was the primary cause of postglacial deformation. The effective rigidity of the crust caused elimination of local variations in mass differences, thereby strengthening the relationship between the ice-sheet mass and postglacial deformation.

Inflections

The inflections or second-order perturbations from the deformation line are more difficult to explain than the major pattern. Although the inflections constitute less than 10% of the total deformation, they are very significant because they may have been caused by regional tectonism, or by variations in the rate of retreat of the Vashon glacier.

Rate of Retreat

The rate of glacier retreat can have a profound effect on the shape of the isostatic rebound curve (Broecker, 1966). If the rate of retreat increases, progressively less recovery occurs prior to deglaciation, causing the rebound curve to develop a concave-upward shape. Conversely, a decelerating rate of deglaciation would cause a flattening of the recovery curve. Significant stillstands or readvances of the glacier front would also cause variation in the convexity or concavity of the recovery curve.

Both the radiocarbon chronology and the regional geology indicate that no significant stillstands or readvances of the Puget lobe occurred within the study area during deglaciation. Probably the rate of glacier retreat increased in the northern Puget lowland because of a calving bay near the Strait of Juan de Fuca. The slight concave-upward inflections at 40 km and 90 km north of the glacial limit may have been caused by slight increases in the retreat rate not revealed by the geology and radiocarbon chronology. But they may be due to other causes as well.

Crustal Structure

The deep geologic structure of the Puget lowland is complex and poorly understood, mainly because the area is thickly mantled with Quaternary sediment. Large anomalies in the gravity and aeromagnetic fields in the Puget lowland can be modeled as boundaries between discrete blocks of the crust (Rodgers, 1970; Gower, 1978; fig. 29). Gravity gradients are low within the blocks but steepen sharply at the boundaries. The sharpest and most significant anomaly boundary occurs along a line between Seattle and Bremerton. Tertiary sedimentary rocks crop out immediately south of the boundary, but unconsolidated and poorly consolidated sediments

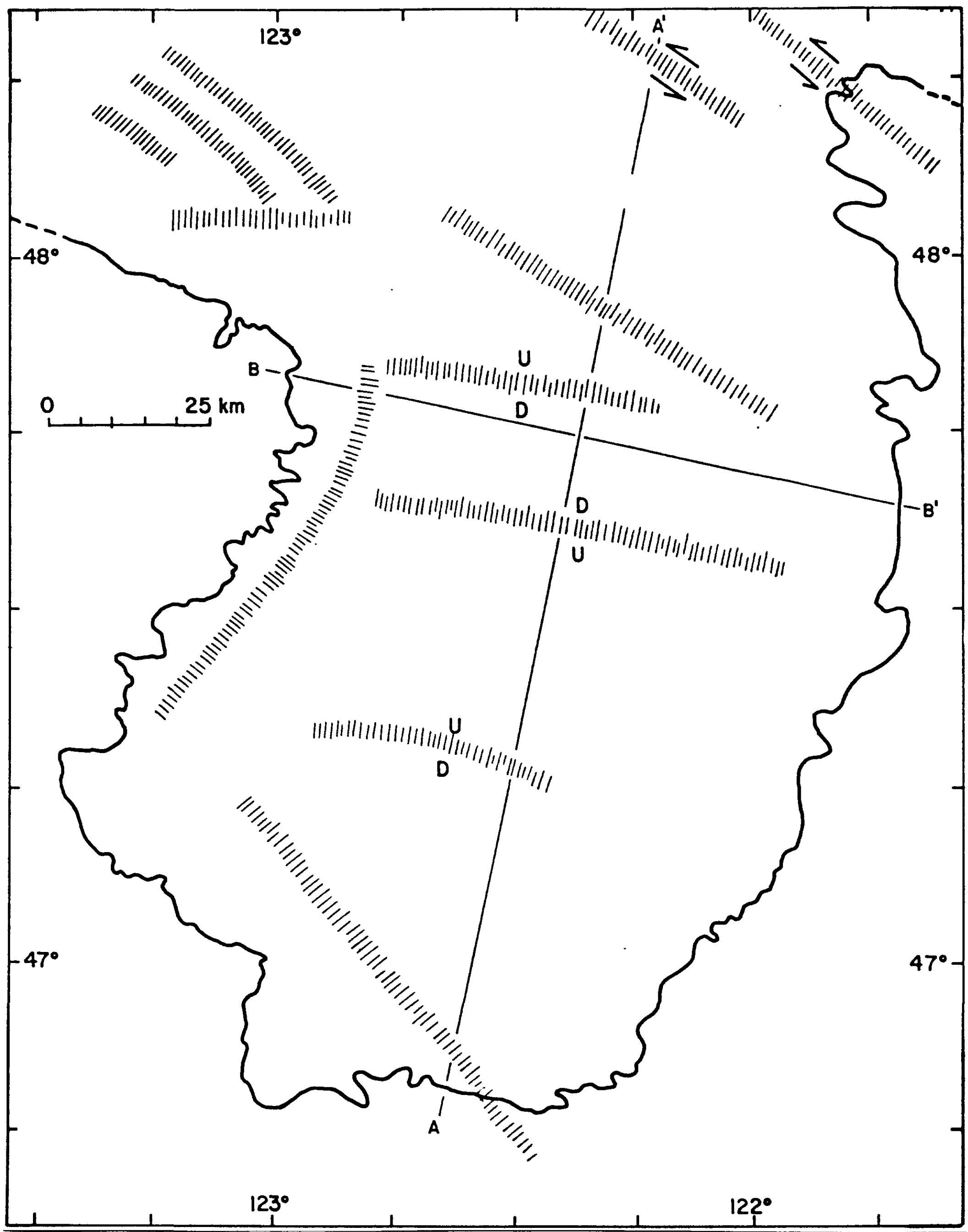

Figure 29.--Map showing boundaries of crustal blocks in the Puget lowland inferred from geophysical data (from Gower, 1978).

extend deeper than 1200 m to the north. The southern block is inferred to be structurally higher because it is composed of denser, older rocks. The remaining anomalies are less accurately located, and their inferred structural relations are more poorly known.

The number and relative spacing of geophysical boundaries as projected to the deformation line are nearly identical to the number and relative spacing of significant inflections that occur on it (figs. 24 and 29). However, the locations of and the inferred sense of movement along the geophysical boundaries do not coincide exactly with the inflections. The apparent association of the inflections with the geophysical boundaries may be coincidental. Alternatively, the structural boundariesmay have partially controlled isostatic compensation and (or) non-isostatic postglacial deformation.

Eldon-Fulton Creek Inflection

The best-documented inflection on the deformation line occurs west of Bremerton about 70 km north of the Vashon limit. Four control points in this area (figs. 23, 24) indicate that the deformation line abruptly decreases in slope for about 10 km between Eldon and Fulton Creek, where it is apparently nearly horizontal (about 0.04 m/km, 4% of the regional slope). If the regional slope is subtracted, there appears to have been about 9 m of differential deformation between Eldon and Fulton Creek, with the north side warped down relative to the south. The apparent absence of evidence for a significant stillstand of the glacier in this sector strongly suggests that the inflection is related to structural, rather than glacial-isostatic, causes.

The Eldon-Fulton Creek (E-FC) inflection, and possibly the other inflections, are probably related to the underlying crustal structure. Because of the uncertainty involved with the inflections at 40 km and 90 km north of the Vashon limit, only the E-FC inflection is considered. Three hypotheses unrelated to glacial isostasy can be invoked to explain the E-FC inflection: (1) it was caused by compaction of near-surface sediments, (2) it represents differential response of the crust to glacier loading, or (3) it was caused by postglacial deformation unrelated to loading or unloading.

All of the control points on the E-FC inflection lie in areas where Tertiary rocks occur very near the surface. The uniform local slope of the deformation line appears too regular to be accounted for by local differences in sediment compaction. Most compaction should have occurred when the Puget lobe lay over the area, yet the control points were derived from recessional features. These arguments indicate that the sediment compaction hypothesis probably is incorrect.

The second hypothesis requires that different rocks within the upper crust responded differently during isostatic deformation. The greater isostatic response should have occurred in the less-dense rocks north of the Seattle-Bremerton boundary because they should be more compliant than the denser rocks to the south. However, the observed inflection across the Seattle-Bremerton geophysical boundary (down to the north) indicates that relative deformation occurred in the opposite direction. This suggests that the E-FC inflection was not caused by differential response to ice-sheet loading.

A northeast-oriented geophysical boundary parallels Hood Canal. Although the control points east of Hood Canal are slightly higher and are less well defined than control points on the west side, no great discrepancies in inferred postglacial deformation occur. If the rocks on opposite sides of this major structural boundary were greatly different in their isostatic response, significant differences in postglacial deformation should be observable. The apparent absence of large differences in deformation on opposite sides of Hood Canal implies (1) that rocks in the upper crust responded uniformly across this structural boundary in response to ice-sheet loading, and (2) that no significant uplift of the western crustal block, relative to the eastern block, has occurred since deglaciation.

The partial refutation of the glacial-stillstand, sediment-compaction, and differential-response hypotheses, and the lack of support for them suggest that the E-FC inflection was caused by postglacial warping unrelated to glacial-isostatic recovery. The Fulton Creek area apparently has been warped down about 4 to 9 m relative to the Eldon area. The apparent absence of down-to-the-north warping beyond the E-FC inflection suggests that it is of local origin. The other inflections at 40 and 90 km possibly also resulted from similar localized north-south postglacial warping.

RATES OF ISOSTATIC UPLIFT

Northern Puget Lowland

The rates of isostatic uplift in many areas of the world have been determined by radiocarbon dating of organic remains that are related to past sea levels (Andrews, 1975). Such data must also be used in the Puget lowland because no dates are currently available from the postglacial lake deposits. Minimum uplift rates for the lowland can be determined by knowing the age and magnitude of the postglacial isostatic anomaly, and the time at which it was largely compensated. There are not enough radiocarbon dates to define a local uplift curve.

The lowest probable altitude, and the oldest probable age for the marine limit, when used in conjunction with the youngest possible age for the approximate completion of isostatic recovery, will provide a minimum estimate for the rate of recovery. The age of the marine limit in the Puget lowland can be no older than about 13,000 years. At least five radiocarbon dates between about 13,100 and 12,535 ^{14}C yr B.P. (table 2) indicate that the region well north of the lowland was deglaciated by that time. The oldest date of 13,100 ± 170 ^{14}C yr B.P. (UW-32) provides a close maximum age for the marine limit in the northern Puget lowland.

The magnitude of the postglacial isostatic anomaly in the northern Puget lowland rises northward, reaching a maximum of about 140 m near northern Camano and central Whidbey Islands. If this figure is adjusted for the maximum error expected in estimating the height of eustatic sea level (20 m), and adjusted for an additional 10 m of postglacial deformation unrelated to glacial isostasy, then the minimum magnitude of the original postglacial isostatic anomaly was about 110 m.

The youngest probable age for the approximate completion of isostatic recovery can be inferred from terrestrial peats on Whidbey Island that have been submerged by Holocene rise of sea level. At Cranberry Lake, Heusser (1960) reported 40 ft (12.5 m) of terrestrial peat which overlies brackish-water peat and a volcanic ash. D. D. Biederman (unpublished report, Univ. Washington, 1967) interpreted these as indicating a gradual rise in sea level since deposition of the ash. Biederman reported a similar gradual marine submergence at Hancock Lake of at least 20 ft (6 m) since ash deposition. The ash exposed at Cranberry Lake is almost certainly the Mazama ash because no other major Holocene tephra is known to occur in the northern Puget lowland. Its estimated age of 6600 yr B.P. (Powers and Wilcox, 1964) provides a minimum age for the beginning of the Holocene marine submergence. Because the gradual submergence could only have occurred after isostatic uplift was largely complete, the age of the tephra also provides a minimum age for the approximate completion of isostatic recovery. From the radiocarbon dates and associations outlined above, the minimum mean rate of postglacial isostatic rebound is computed to be 1.7 cm/yr.

Western Washington and Southwestern British Columbia

In an attempt to improve and support the estimate for the rate of isostatic recovery in the Puget lowland, all of the radiocarbon dates related to past changes in sea level in northwest Washington and southwestern British Columbia were reviewed. These data have been previously outlined, reviewed, and interpreted by Mathews and others (1970), Clauge (1975), and Fulton (1971). When the available 60 dates are plotted, they show several geographic clusters. These clusters, which lie near Whidbey Island, the San Juan Islands, and Whatcom County, Washington, and the Victoria, Fraser lowland, Parksville, and Courtenay areas in southwestern British Columbia, were further restricted to include only those dates within a radius of 30 km from the center of each cluster. Dates from widely different localities were not used because it is not reasonable to assume that all areas had the same deglacial and isostatic history.

In northern Washington the dates from Whidbey Island and the San Juan Islands were few in number and low in altitude; hence, they could not be used to establish an uplift curve. Easterbrook (1963) interpreted dates from Whatcom County to indicate that large, rapid, and oscillating changes in sea level occurred following deglaciation. These inferred changes in sea level in Whatcom County are based mainly on stratigraphic interpretations, and are controversial. Easterbroook invoked tectonic and local glacial isostatic movements, combined with eustatic sea-level changes, to explain the inferred local fluctuations in sea level. He suggested that the readvance of Cordilleran ice in the Fraser lowland during the Sumas Stade was partially responsible for a post-Vashon re-submergence. The results of this investigation suggest that a readvance involving such a small quantity of ice could not cause a resubmergence of hundreds of feet (about 100 m) at a distance of tens of kilometers beyond the limit of the Sumas ice. Because eustatic sea level was rising too slowly (about 1 cm/yr) at that time to account for the rapid resubmergence, the deformation, if it occurred, was probably due mainly to tectonic causes.

Dates from the Parksville and Courtenay areas along Vancouver Island fall between 11,500 $\pm$ 200 (L-4416) and 12,500 $\pm$ 450 (GSC-9) ^{14}C yr B.P., and the standard errors of most of the radiocarbon dates overlap. There are not enough dates available to draw a well-defined uplift curve. Dates of glacial-marine sediments from near Victoria provide good control for the minimum ages and minimum heights of relative sea level. The dates, however, are not sufficiently separated in time to allow the definition of a well-controlled uplift curve during the initial phase of isostatic recovery (Mathews and others, 1970); the oldest six dates have overlapping standard errors, yet were derived from samples that span more than 60 m in altitude (about 80% of the total postglacial uplift). Although rate of uplift in the Victoria area was very high, it cannot be accurately defined. Isostatic compensation in this area apparently was largely complete within a few thousand years following deglaciation (Mathews and others, 1970).

In the Fraser lowland near Vancouver, 20 radiocarbon dates older than 5000 yr B.P. are related to former sea-level positions (table 5). These sample localities and dates are sufficiently separated in altitude and time, respectively, to define an approximate uplift curve (fig. 30). Mathews and others (1970) interpreted a

TABLE 5.-- RADIOCARBON DATES FROM THE FRASER LOWLAND, B.C. ASSOCIATED WITH POSTGLACIAL EMERGENCE.

(Dates taken mainly from a list compiled by W. H. Mathews, 1970)

Lab Number	Age ^{14}C yr B.P.	Location N. Lat.	Location W. Long.	Relative Sea Level (m)	Material
GSC-6	12,625 ± 450	$49^{\circ}01'$	$122^{\circ}50'$	> 53.4	shells
GSC-64	12,460 ± 170	$49^{\circ}08'$	$122^{\circ}55'$	> 38.1	shells
GSC-74	12,230 ± 200	$49^{\circ}15'$	$122^{\circ}56'$	> 11.9	shells
GSC-168	11,930 ± 190	$49^{\circ}10'$	$122^{\circ}35'$	> 12.2	shells
GSC-186	11,680 ± 180	$49^{\circ}06'$	$122^{\circ}30'$	> 94.5	shells
GSC-227	11,300 ± 190	$49^{\circ}35'$	$123^{\circ}13'$	> 41.2	shells
L-221e	11,000 ± 900	$49^{\circ}06'$	$122^{\circ}23'$	> 76.2 ± 3.0	wood with shells
GSC-185	10,690 ± 180	$49^{\circ}35'$	$123^{\circ}13'$	≃ 41.2	shells with wood
S-19	10,430 ± 150	$49^{\circ}05'$	$122^{\circ}47'$	> 32.0	shells
GSC-228	9,420 ± 180	$49^{\circ}15'$	$122^{\circ}56'$	< 11.9	peat
S-113	9,000 ± 150	$49^{\circ}33'$	$122^{\circ}24'$	< 77.7	charcoal
GSC-225	8,360 ± 170	$49^{\circ}02'$	$122^{\circ}16'$	<-10.7	wood
GSC-229	8,290 ± 140	$49^{\circ}13'$	$122^{\circ}42'$	<-10.4	peaty silt
S-47	8,150 ± 300	$49^{\circ}33'$	$121^{\circ}24'$	< 68.6	charcoal
GSC-3	7,600 ± 150	$49^{\circ}13'$	$122^{\circ}48'$	<-10.1	peaty silt
S-61	7,350 ± 150	$49^{\circ}33'$	$121^{\circ}24'$	≃-10.1	charcoal
GSC-321	7,340 ± 160	$49^{\circ}15'$	$122^{\circ}58'$	≃- 8.2	peat
S-99	7,300 ± 120	$49^{\circ}13'$	$122^{\circ}48'$	<-10.1	detrital peat
GSC-395	6,790 ± 150	$49^{\circ}03'$	$123^{\circ}04'$	>-13.7	shells
M-1547	5,490 + 500	$49^{\circ}33'$	$122^{\circ}20'$	≃ 0.0	charcoal

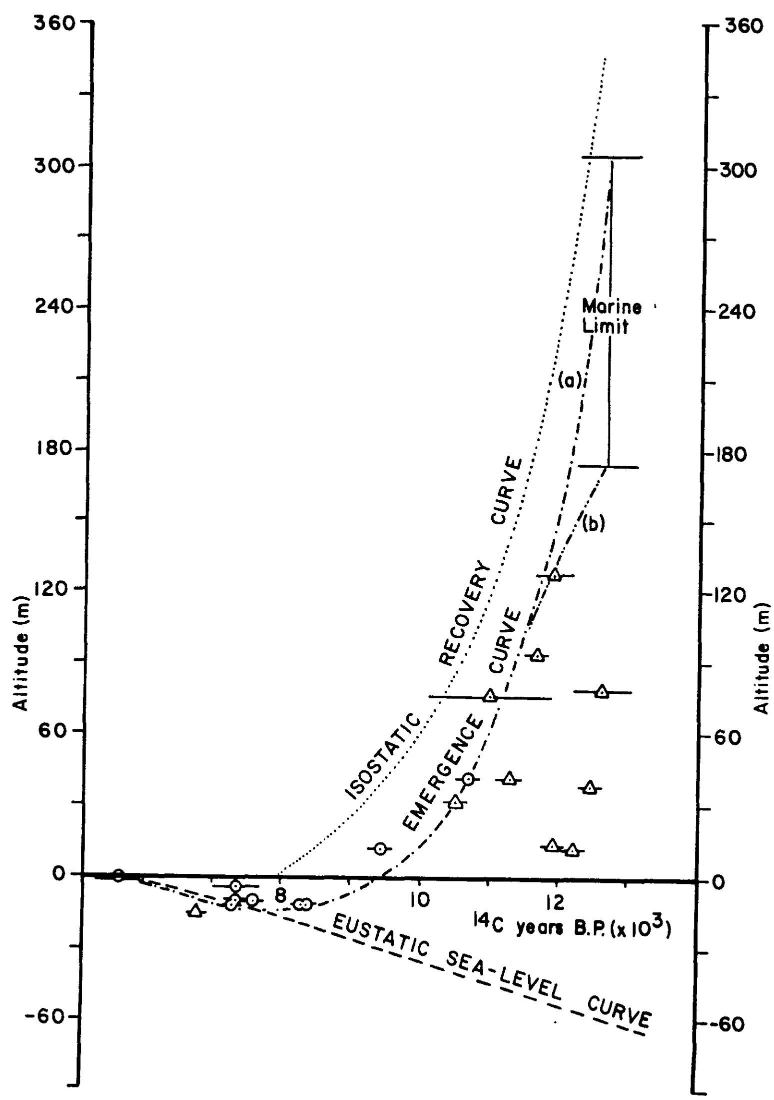

Figure 30.--Radiocarbon-dated sea-level relations in the Fraser Lowland, B.C. Triangles and circles represent locations and ages of radiocarbon samples that lay below, or at (or) above sea level, respectively. Emergence curves a and b are based on maximum and minimum estimates for the height of the marine limit, respectively. Eustatic sea-level curve is redrawn from Curray (1965).

postglacial recovery curve for the Fraser lowland that includes a large post-Vashon resubmergence similar to that postulated by Easterbrook (1963). This resubmergence was also based on stratigraphic interpretations and its age was inferred mainly from dates on samples obtained outside the Fraser lowland. In this study no evidence was found that discounts a Sumas resubmergence in the Fraser lowland, but the existing radiocarbon dates apparently do not require it. The uplift curve for the Fraser lowland used in this study is based only on the published radiocarbon dates that were tabulated by Mathews and others (1970, footnote 2, p. 690).

The age and altitude of the marine limit in the Fraser lowland is crucial to the postglacial recovery curve for this region. Armstrong (1957) indicated that the marine limit near Vancouver extended to at least 575 ft altitude (175 m), the highest known occurrence of marine shells. In this area "...stony clays and deltaic deposits possibly marine, are common up to about 1000 ft (300 m) above sea level" (Mathews and others, 1970, p. 695). Although the higher figure of 1000 ft (300 m) is probably more accurate than the estimate of 175 m, both estimates are used to determine the uplift rates in the Fraser lowland. The age of the marine limit in the Fraser lowland is not known exactly but it probably is close to 12,625 $\pm$ 450 ^{14}C yr B.P. (GSC-6), the oldest dated marine sediments. Two other radiocarbon dates from the Fraser lowland offer critical control for the shape of the uplift curve. A date of 10,690 $\pm$ 180 (GSC-185) was obtained from a sample that indicates a relative sea-level of about 135 ft (41 m) above present. A date of 8360 $\pm$ 170 ^{14}C yr B.P. (GSC-225) from terrestrial deposits at about 35 ft (11 m) below present sea level indicates that relative sea level lay below -11 m at that time.

A smooth curve for the changes in relative sea level can be drawn to include these three critical dates (fig. 30). The remaining dates, both above and below the inferred relative sea level, support this provisional curve. A curve of isostatic recovery can be constructed by subtracting the eustatic sea level curve from it (fig. 30).

The uplift curve is remarkably similar in form to that determined from other parts of the world where the rate of recovery is known more accurately (Hafsten, 1960; Washburn and Stuiver, 1962; Andrews, 1968). These curves can be expressed by the following equation:

$$U_{pt} + U_p - U_p e^{-kt}$$

where U_{pt} is the amount of uplift that has occurred at time t, U_p is the total uplift measured, and k is a decay constant (Andrews, 1975). The parameters commonly used to express the rate and shape of the uplift curve are the half-response time ($t_{1/2}$), the decay constant (k), and the relaxation time (T_r). These parameters cannot be derived directly from the recovery curve for the Fraser lowland because it is poorly controlled. Nonetheless, they can be derived from the more general equation for isostatic decay discussed above.

The values for isostatic parameters obtained from minimum rates of uplift for the Fraser lowland and the northern Puget lowland are shown in table 6. In the Fraser lowland half-response times range from 800-1600 yr and initial rebound rates for uplift range from 23 to 7 cm/yr. Calculated for the northern Puget lowland, the half-response time is about 1650 years and initial rebound rate is 3.3

cm/yr. Under conditions of rapid emergence, resubmergence, and re-emergence postulated by Easterbrook (1963) and Mathews and others (1970), the minimum rates of uplift would have been several times faster, and the relaxation time would have been on the order of 500 years (Mathews and others, 1970).

TABLE 6.-- MINIMUM POSTGLACIAL ISOSTATIC UPLIFT RATES FOR THE NORTHERN PUGET LOWLAND, WASHINGTON, AND THE FRASER LOWLAND, B.C.

Half-response time ($t_{1/2}$) (years)	Decay constant (k) (x 10^{-4})	Relaxation time (T_r) (years)	Uplift Rates (cm/yr)	
			$t_{1/2}=1$	$t_{1/2}=2$
Puget Lowland				
1657	4.20	2378	3.3	1.7
Fraser Lowland				
1036 1/, 2/	6.68	1497	17.5	8.7
772 2/, 3/	8.98	1113	23.5	11.8
1572 1/, 4/	4.40	2272	7.4	3.67
872 3/, 4/	7.95	1258	13.4	6.7

1/ Calculation based on emergence between 12,625 and 10,690 yr B.P.

2/ Calculation based on maximum height of marine limit.

3/ Calculation based on emergence between 12,625 and 8,320 yr B.P.

4/ Calculation based on minimum height of marine limit.

COMPARISON OF OBSERVED AND PREDICTED ISOSTATIC DEFORMATION

Comparison by Mass Compensation

The earth's crust acts as an elastic membrane that operates to distribute a local load over a large area. Mass imbalances that are larger than the strength of the crust can cause disequilibrium. Isostatic response to disequilibrium occurs largely by viscous Maxwellian flow in the lower crust and (or) upper mantle (McConnell, 1968). Crittenden (1963) has shown that for a load comparable in size to the Puget lobe, elastic effects can be considered negligible. If the volume of the isostatically depressed area and the density of the materials that underwent viscous deformation are known approximately, then the approximate mass of the original isostatic anomaly can be calculated.

Measured Anomaly

The volume of the isostatically depressed area within the Puget lowland can be determined directly from the contour map of measured deformation (fig. 23), but the deformation beyond the margin of the lowland is more difficult to determine. In order to infer the deformation beyond the margin of the Puget lowland, the Puget lobe must be modeled as a slab in cross section. Crittenden (1963) has described field evidence from Lake Bonneville to support the theoretically determined profile of isostatic deformation beyond the edge of such a slab. Although the theoretical profile of deformation beyond the margin of a slab is not a linear function, a linear decrease in isostatic deflection beyond the loaded area can be used to approximate the cross sectional area of the deflection. If the slope of the deformation line derived from within the Puget lowland (0.92 m/km) is used to determine the profile of deformation beyond the borders of the lowland, then the approximate volume of deformation beyond the lowland margin can be approximated (table 7; fig. 31).

The product of the volume of isostatic depression and the density of the substratum in the viscous zone is the <u>mass</u> of the measured isostatic anomaly. Using a value of 3.3 gm/cm^3 for the density of subcrustal rocks, an average of the values used by workers in glacial isostasy (Bloom, 1970; Crittenden, 1963; Mathews and others, 1970; McConnell, 1968), the mass of the depressed area within and beyond the Puget lowland is calculated to be about 4.3×10^{18} gm.

Predicted Anomaly

The mass of the isostatic anomaly caused by the maximum glacial load under conditions of equilibrium can also be calculated if the volume of the ice is known and the yield strength of the crust can be estimated. A surface load will not necessarily cause an isostatic compensation exactly equivalent to its mass because

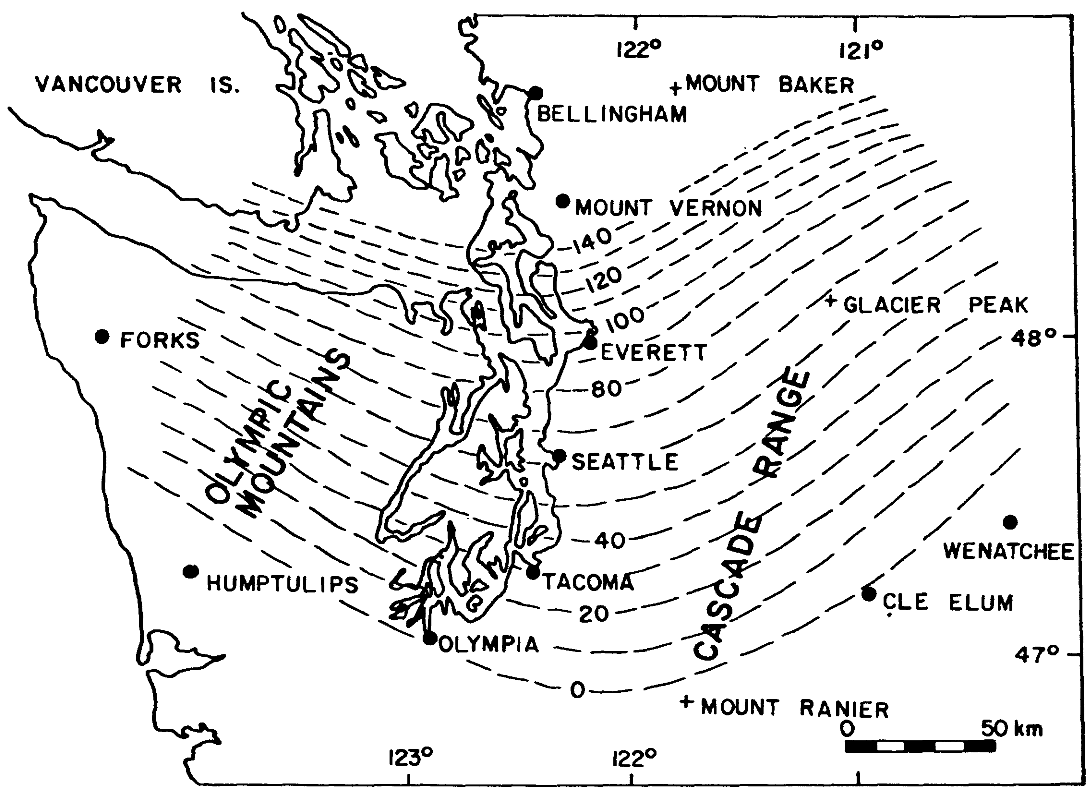

Figure 31.--Hypothetical pattern of isostatic deformation beyond the margin of the Puget lowland. Deformation within the lowland (fig. 23) was extrapolated beyond the lowland margin by assuming that deformation occurred with similar north-south (fig. 24) and east-west regional gradients. The isostatic effects of valley glaciers in the Cascade Range and Olympic Mountains are not considered in this analysis, but deformation contour lines east of Glacier Peak were adjusted in the direction suggested by possible isostatic effects of the Cordilleran Ice Sheet east of the Cascade Range. Data from the Puget lowland suggest that isostatic effects of the Juan de Fuca lobe were relatively minor near the northeastern Olympic Peninsula.

TABLE 7.-- COMPARISON BETWEEN THE MASSES OF THE MEASURED POSTGLACIAL ISOSTATIC ANOMALY AND THE EXPECTED EQUILIBRIUM ANOMALY AT THE VASHON MAXIMUM.

Component	Area (km^2)	Volume (km^3)	Density (gm/cm^3)	Mass ($\times 10^{18}$ gm)	Percent (%)
			Calculated Anomaly		
Glacial load					
Puget lobe	14,231	9,705	0.917	8.9	----
Area to north	6,300	9,859	0.917	3.9	----
Subtotal	20,531	19,564	----	12.8	----
Yield strength 1/	20,531	1,129	0.917	1.1	----
Total	-----	----	----	11.7	100
			Measured Anomaly		
Within and beyond Puget lowland	34,913	1,300	3.3	4.29	37

1/ Amount of glacial load compensated for by yield strength of crust.

the crust has a certain inherent rigidity, i.e., deformation will not occur until the yield strength of the crust is exceeded. The yield strength of the crust is not well known, but appears to be quite low (Crittenden, 1967). A yield strength of 5 bars, a value used in these calculations, is equivalent to a vertical column of ice about 55.5 m thick. If the mass of a layer of ice this thick over the surface of the glaciated area is subtracted from the mass of the glacier itself, the mass of ice that should have caused an isostatic imbalance can be determined. Only about 10% of the glacial load would not have been compensated for due to the strength of the crust (table 7).

At its maximum extent the Puget lobe, as strictly defined, had a volume of about 9700 km^3 and a mass of about 8.9×10^{18} gm. The quantity of ice that occupied the isostatically depressed area was less than 10% of the ice-sheet mass given above, or about the same amount as that caused by the yield strength of the crust.

The Cordilleran Ice Sheet north of the Puget lowland can be approximated as a body of ice that was rectangular in plan (70-km wide and infinite length) and had a gradient of 3.9 m/km, increasing northward. The influence of this portion of the ice sheet on the isostatic anomaly in the Puget lowland should decrease northward with approximately the same slope as the deformation line. If the ice-sheet in this area is subdivided into a series of slabs, each 10-km wide in north-south extent, the isostatic depression in the Puget lowland that was caused by each slab can be estimated. According to these calculations, the ice load north of the Puget lowland would have caused an additional 3.9×10^{18} gm. of isostatic anomaly within the Puget lowland (table 7). The measured isostatic anomaly apparently was only about 37% of the anomaly determined for full isostatic equilibrium of the maximum ice load. A wide but undetermined margin of error must be applied to this and earlier estimates.

Although not at their maximum extents, valley glaciers of the central and southern Cascades and Olympic Mountains were still present during the Vashon Stade. Regardless of their mass relative to the Puget lobe they do not enter significantly into the comparison between the predicted and observed isostatic anomalies in the Puget lowland because the isostatic depression beyond the edge of the Puget lobe was determined independently of the valley glaciers. The major mass of valley glaciers were also relatively distant from the Puget lobe.

Comparison by Depth of Compensation

Another means of calculating the amount of isostatic compensation that occurred is to compare the thickness of the ice sheet to the depth of the isostatic depression. The Puget lobe in the northern Puget lowland can be modeled as a slab with low north-south and east-west gradients. Under equilibrium conditions deformation should have reached its maximum value in the center of the lowland because, on the basis of the Lake Bonneville data (Crittenden, 1967, p. 266), the Puget lobe was sufficiently wide (100 km) that edge effects can be disregarded. The maximum isostatic anomaly at equilibrium should have been the product of the maximum ice thickness and the ratio of densities for the ice and the substratum that yielded isostatically. Where the measured isostatic deformation is 100 m, the equilibrium isostatic depression should have been about 340 m. The measured isostatic anomaly is, therefore, approximately 34% of that calculated for equilibrium conditions.

The proportion of observed isostatic compensation relative to its predicted equilibrium value is nearly the same for both calculations, suggesting that the estimates of the magnitude of the isostatic anomaly at full compensation are approximately correct. The discrepancy between observed and predicted values is due to (1) the errors involved in the assumptions and approximations, (2) the amount of isostatic disequilibrium that remained at the time of maximum isostatic compensation, and (3) the amount of isostatic rebound that occurred during glacial retreat but prior to deglaciation of the Puget lowland.

Restrained Rebound

Progressive thinning of the Puget lobe accompanied northward retreat of the glacier margin. Large amounts of mass were therefore removed above the control-point localities prior to their complete deglaciation; these losses were partially adjusted for isostatically during glacier retreat. The amount of isostatic rebound that occurred during ice wastage is referred to as restrained rebound. Andrews (1970) has shown that restrained rebound can be an extremely significant part of total isostatic recovery, perhaps as much as 75%. The relative amount of restrained rebound is critically dependent on the rate of glacier retreat; slowly retreating glaciers allow a much higher proportion of isostatic recovery to occur at a given point prior to deglaciation. Rapidly retreating glaciers should experience relatively little restrained rebound.

Although the amount of restrained rebound for the Puget lowland cannot be determined, it can be estimated by knowing the amount of time it took for the lowland to be deglaciated. A maximum estimate can be made by assuming that (1) restrained rebound occurred everywhere in the lowland during the total time required for deglaciation, and (2) the rate of restrained rebound during this time was equal to the relatively greater rate of rebound when the area was completely ice free. Because the maximum time possible for deglaciation is about 1500 years, approximately equal to the maximum half-response time, restrained rebound must have accounted for no more than about 50% of the total isostatic recovery. The actual value was probably much lower because many areas were deglaciated prior to 1500 years after retreat began, and because the rate of restrained rebound was almost certainly lower than the rate of postglacial recovery. If deglaciation of the lowland was instantaneous, no restrained rebound would have occurred.

Relative Isostatic Compensation

The proportion of isostatic equilibrium that was attained in the Puget lowland during the Vashon Stade can be estimated by using the maximum (50%) and minimum (0%) values for restrained rebound. If the above assumptions and calculations are valid, and if the errors involved are not too great, then isostatic adjustments of the Puget lowland to glacial loading reached about 35% to 70% of the equilibrium value. Because the Puget lobe apparently existed for no more than about 2000 years, and is inferred to have been at its maximum extent for no more than a few hundred years, this degree of compensation appears surprisingly high, and indicates that the Puget lowland responded very rapidly to the mass imbalance caused by ice-sheet glaciation.

IMPLICATIONS

The observation that the Puget Sound region can be easily deformed in a brief time by relatively small crustal loads suggests that the subcrustal viscosity in this region is abnormally low. Alternatively, but probably less likely, the rapid depression and recovery of this region may be attributed to the presence of a strong elastic crust.

The foci of hundreds of recent small earthquakes within the Puget lowland have been accurately located by the University of Washington seismograph network (Crosson, 1972, 1977). The most easily observed feature of the earthquake-epicenter pattern is that they are randomly distributed within, and commonly restricted to, the Puget lowland; relatively few earthquake epicenters lie within the Cascade Range and Olympic Mountains or within the foothills south of the lowland.

Earthquake focal depths, however, fall into two general groups. The foci in the deeper group occur between about 40 km and 60 km depth, and possibly decrease in depth eastward. The foci of the shallow group extend no deeper than about 25-30 km and apparently decrease in both frequency and depth away from the axis of the Puget lowland.

The distribution in depth and area of the shallow group of small earthquakes in western Washington is remarkably similar to the pattern of isostatic recovery. This correspondence suggests that incomplete isostatic recovery, or adjustments to completed rebound, may be related in some way to the pattern of recent seismicity. However, the similarity in the pattern of seismicity and isostatic deformation may be coincidental.

A low effective viscosity of the substratum below the Puget lowland would carry significant implications for regional tectonism in western Washington. The mechanisms of subduction and volcanism in this area are critically related to the strength, viscosity, and distribution of rocks below the surface. A low-viscosity substratum suggests that active tectonic forces are responsible for maintenance of the anomalies. Data provided by glacial isostasy in western Washington should provide new constraints for models of ongoing tectonism.

The relatively rapid loading and unloading of the Puget lowland by as much as 10,000 km^3 of ice caused large stresses to be built up within the upper crust. Lateral load differences may locally have set up stresses large enough to cause rupture and faulting of near-surface rocks. Thus, surface faults that show postglacial movement may have been reactivated or even caused by isostatic recovery following deglaciation, but mechanisms completely independent from glacial isostasy cannot be ruled out at this time.

CONCLUSIONS

(1) At the time of its maximum extent about 14,000 years ago, the Puget lobe of the Cordilleran Ice Sheet extended across the Puget lowland between the Cascade Range and Olympic Mountains. At that time it apparently was nearly in an equilibrium state.

(2) Recession of the Puget lobe largely involved systematic northward retreat of the active glacier terminus.

(3) During initial stages of retreat, a complex system of south-draining proglacial lakes formed in the troughs of Puget Sound beyond the ice margin.

(4) Retreat of the Juan de Fuca lobe was complete prior to deglaciation of the Puget lowland. This resulted in a change in the direction of drainage of the proglacial lakes from south to north.

(5) The different altitudes of numerous outwash deltas built into the proglacial lakes define water planes that have been progressively warped up towards the north.

(6) The amount of postglacial warping in the northern Puget lowland, as determined from emerged marine sediments, agrees closely with the amount determined independently from the lacustrine data. These two data sets indicate that the amount of postglacial uplift at the northern margin of the Puget lowland was about 140 m, and that it decreased progressively southward with a nearly uniform gradient of 0.92 m/km.

(7) Postglacial warping of the Puget lowland was caused largely by isostatic rebound associated with recession of the Puget lobe.

(8) Significant inflections that deviate from the regional slope of the deformation line probably were caused by differences in the rate of glacier retreat and by non-isostatic postglacial warping.

(9) Isostatic deformation caused by glaciation of the Puget lowland did not extend far beyond the southern terminus of the Puget lobe, but areas beneath the central Cascade Range and Olympic Mountains were affected considerably.

(10) Comparison between the predicted and observed isostatic anomalies indicate that between about 35% and 70% of isostatic equilibrium was attained at the glacial maximum. Less than about 50% of the total isostatic anomaly was compensated during restrained rebound.

(11) The rates of isostatic uplift in the Puget lowland following deglaciation were at least 3 cm/yr. In the Fraser lowland, where as much as 350 m of postglacial rebound may have occurred, minimum uplift rates were at least 7 cm/yr.

(12) The study of isostatic rebound in the Puget lowland has important implications for investigations of regional crustal structure, Quaternary faulting, and recent seismicity.

REFERENCES CITED

Anderson, C. A., 1965, Surficial geology of the Fall City area, Washington: Unpub. M.S. thesis, Univ. Wash., 70 p.

Andrews, J.T., 1968, Postglacial rebound in Arctic Canada; similarities and prediction of uplift curves: Can. Jour. Earth Science, v. 5, p. 39-47.

______1978, A geomorphological study of post-glacial uplift with particular reference to Arctic Canada: Inst. British Geog. Spec. Publ. No. 2, 156 p.

______1975, Glacial systems; an approach to glaciers and their environments: North Scituate, Mass., Duxbury Press, 191 p.

Armstrong, J.E., 1957, Surficial geology of the New Westminster map-area, British Columbia: Canada Geol. Survey Paper 57-5, 25 p.

Armstrong, J.E., Crandell, D.R., Easterbrook, D.J., and Noble, J.R., 1965, Late Pleistocene stratigraphy and chronology in southwest British Columbia and northwestern Washington: Geol. Soc. of America Bull., v. 76, p. 321-330.

Birdseye, R.V., 1976, Geologic map of east-central Jefferson County, Washington: Wash. State Div. Geol. and Earth Resources Open-File Report 76-26, scale 1:24,000.

Bloom, A.L., 1970, Pleistocene shorelines; a new test of isostasy: Geol. Soc. of America Bull., v. 78, p. 1477-1494.

Boulton, G.S., 1978, Glacier erosion, p. 1 in Volume of Abstracts, Symposium of glacier beds, the ice-rock interface: Ottawa, Canada, National Research Council's subcommittee on glaciers, 32 p.

Bretz, J H., 1913, Glaciation of the Puget Sound region: Washington Geol. Survey Bull., v. 8, 244 p.

Broecker, W.S., 1966, Glacial rebound and the deformation of the shorelines of proglacial lakes: Jour. Geophys. Research, v. 71, p. 4777-4794.

Carson, R.J., III, 1970, Quaternary geology of the south-central Olympic Peninsula, Washington: Unpub. Ph.D. dissertation, Univ. Wash., 67 p.

______1976a, Preliminary geologic map of the Brinnon area, Jefferson County, Washington: Wash. State Div. of Geology and Earth Resources Open-File Report 76-3, scale 1:24,000.

______1976b, Geologic map of north-central Mason County, Washington: Wash. State Div. of Geology and Earth Resources Open-File Report 76-2, scale 1:62,500.

Clark, J.A., Farrell, W.E., and Peltier, W.R., 1978, Global changes in postglacial sea level; a numerical calculation: Quaternary Research, v. 9, p. 265-287.

Clague, J.J., 1975, Late Quaternary sea level fluctuations, Pacific Coast of Canada and adjacent areas: Canada Geol. Survey Paper 75-1c, p. 17-21.

Crandell, D.R., 1963, Surficial geology and geomorphology of the Lake Tapps Quadrangle, Washington: U.S. Geol. Survey Prof. Paper 338-A, 84 p.

Crandell, D.R., Mullineaux, D.R., and Waldron, H.H., 1958, Pleistocene sequence in the southeastern part of the Puget Sound Lowland, Washington: American Jour. of Science, v. 256, p. 384-397.

Crittenden, M.D., Jr., 1963, Effective viscosity of the earth derived from isostatic loading of the Pleistocene Lake Bonneville: Jour. Geophys. Research, v. 68, p. 5517-5530.

______1967, Viscosity and finite strength of the mantle as determined from water and ice loads: Geophys. Jour. Roy. Astr. Soc., v. 14, p. 261-279.

Crosson, R.S., 1972, Small earthquakes, structure, and tectonics of the Puget Sound region: Bull. of the Seis. Soc. of America, v. 62, p. 1133-1171.

______1977, Seismic studies related to regional tectonics in the Pacific Northwest: Geol. Soc. of America Abstracts with Programs, v. 9, p. 940.

Curran, T.A., 1965, Surficial geology of the Issaquah area, Washington: Unpub. M.S. thesis, Univ. Wash., 57 p.

Curray, J.R., 1965, Late Quaternary history, continental shelves of the United States, p. 723-735 in Wright, H.E., and Frey, D.C., eds., The Quaternary of the United States: Princeton, N.J., Princeton Univ. Press, 922 p.

Easterbrook, D.J., 1963, Late Pleistocene glacial events and relative sea level changes in the northern Puget lowland, Washington: Geol. Soc. of America Bull., v. 74, p. 1465-1484.

______1968, Pleistocene stratigraphy of Island County, Washington: Washington Dept. of Water Resources Water-Supply Bull. 25, Part 1, 34 p.

Frisken, J.G., 1965, Pleistocene glaciation of the Brinnon area, east-central Olympic Peninsula, Washington: Unpub. M.S. thesis, Univ. Wash., 75 p.

Fulton, R.J., 1971, Radiocarbon geochronology of southern British Columbia: Geol. Survey of Canada Paper 71-37, 28 p.

Gayer, M.J., 1977, Geologic map of northeastern Jefferson County, Washington: Wash. Div. of Geology and Earth Resources Open-File Report 76-21, scale 1:24,000.

Gower, H.D., 1978, Tectonic map of the Puget Sound region, Washington, showing locations of faults, principal folds, and large-scale Quaternary deformation: U.S. Geol. Survey Open-File Report 78-426, scale 1:250,000, 17 p. text.

Gustavson, T.C., Ashley, G.M., and Boothroyd, J.C., 1975, Depositional sequences in glaciolacustrine sedimentation, p. 264-280 in Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and glaciolacustrine sedimentation: Tulsa, Oklahoma, Society of Econ. Paleon. and Mineralogists Special Publication No. 23, 320 p.

Hafsten, Ulf, 1956, Pollen-analytic investigations on the late Quaternary development in the inner Oslofjord area: Bergen, Norway, A.S. John Griegs Biktrykkeri, Universitetet I Bergen, Naturvitenskapelig rekke, Nr. 8, 161 p.

Hansen, K.L., 1976, Geologic map of the Uncas-Port Ludlow area, Jefferson County, Washington: Wash. Div. of Geology and Earth Resources, Open-File Report 76-20, scale 1:24,000.

Heiskanen, W.A., and Vening Meinesz, F.A., 1958, The earth and its gravity field: New York, McGraw-Hill, Inc., 470 p.

Heusser, C.J., 1960, Late Pleistocene environments of North Pacific North America: Am. Geogr. Soc. Spec. Pub. 35, 308 p.

______1977, Quaternary palynology of the Pacific slope of Washington: Quaternary Research, v. 8, p. 282-306.

Hirsch, R.M., 1975, Glacial geology and geomorphology of the upper Cedar River watershed, Cascade Range, Washington: Unpub. M.S. thesis, Univ. Wash., 48 p.

Hughes, T., 1977, West Antarctica ice streams: Reviews of Geophys. and Space Physics, v. 15, p. 1-46.

Hughes, T., Denton, G.H., and Grosswald, M.G., 1977, Was there a late Wurm ice sheet?: Nature, v. 206, p. 596-602.

Kamb, Barclay, 1964, Glacier geophysics: Science, v. 146, p. 353-365.

Knoll, K.M., 1967, Surficial geology of the Tolt River area, Washington: Unpub. M.S. thesis, Univ. Wash., 91 p.

Liesch, B.A., Price, C.E., and Walters, K.L., 1963, Geology and ground-water resources of northwestern King County, Washington: Wash. Div. of Water Resources Water-Supply Bull. 20, 241 p.

Luzier, J.E., 1969, Geology and ground-water resources of southwestern King County, Washington: Wash. Dept. of Water Resources Water-Supply Bull. 28, 260 p.

Mackin, J.H., 1941, Glacial geology of the Snoqualmie-Cedar area, Washington: Journal of Geology, v. 49, p. 449-481.

Macintyre, I.G., Pilkey, O.H., and Stuckenrath, Robert, 1978, Relict oysters on the United States Atlantic continental shelf; a reconsideration of their usefulness in understanding late Quaternary sea-level history: Geol. Soc. of America Bull., v. 89, p. 277-282.

Mathews, W.H., Fyles, J.G., and Nasmith, H.W., 1970, Postglacial crustal movements in southwestern British Columbia and adjacent Washington State: Can. Jour. Earth Science, v. 7, p. 690-702.

McConnell, R.J., Jr., 1968, Viscosity of the mantle from relaxation time spectra of isostatic adjustment: Jour. Geophys. Research, v. 73, p. 7089-7105.

Mercer, J.H., 1969, The Allerod oscillation; a European climatic anomaly?: Arctic and Alpine Research, v. 1, p. 227-234.

Moen, A.D., ed., 1971, Symposium on tectonism of the Pacific Northwest: Transactions American Geophys. Union, v. 53, p. 628-645.

Molenaar, Dee, and Garling, M.E., 1965, Water resources and geology of the Kitsap Peninsula and certain adjacent islands: Wash. Div. of Water Resources Water-Supply Bull. 18, p. 24-41.

Molenaar, Dee, and Noble, J.B., 1970, Geology and related ground water occurrences, southeastern Mason County, Washington: Wash. Div. of Water Resources Water-Supply Bull. 24, p. 21.

Mullineaux, D.R., 1965a, Geologic map of the Auburn quadrangle, King and Pierce Counties, Washington: U.S. Geol. Survey Map GQ-406, scale 1:24,000.

______ 1965b, Geologic map of the Black Diamond quadrangle, King County, Washington: U.S. Geol. Survey Map GQ-407, scale 1:24,000.

______ 1965c, Geologic map of the Renton quadrangle, King County, Washington: U.S. Geol. Survey Map GQ-405, scale 1:24,000.

Mullineaux, D.R., Waldron, H.H., and Rubin, Meyer, 1965, Stratigraphy and chronology of late interglacial and early Vashon time in the Seattle area, Washington: U.S. Geol. Survey Prof. Paper 672, 92 p.

Newcomb, R.C., 1952, Ground-water resources of Snohomish County, Washington: U.S. Geol. Survey Water-Supply Paper 1135, 133 p.

Noble, J.R., and Wallace, F.F., 1966, Geology and ground-water resources of Thurston County, Washington, v. 2: Wash. Div. of Water Resources Water-Supply Bull. 10, p. 1-61.

Nye, J.F., 1959, The motion of ice sheets and glaciers: Jour. of Glaciology, v. 2, p. 493-507.

Pevear, D.R., and Thorson, R.M., 1978, Distinction between till and glaciomarine drift using exchangeable sodium: Geol. Soc. of America Abstracts with Programs, v. 10, p. 471.

Porter, S.C., 1970, Glacier recession in the southern and central Puget lowland, Washington, between 14,000 and 13,000 B.P.: Abstracts, American Quaternary Association, Bozeman, Montana, p. 107.

______1976, Pleistocene glaciation in the southern part of the North Cascade Range, Washington: Geol. Soc. of America Bull., v. 87, p. 61-75.

Porter, S.C., and Carson, R.J., III, 1971, Problems of interpreting radiocarbon dates from dead-ice terrain, with an example from the Puget Lowland of Washington: Quaternary Research, v. 1, p. 410-414.

Powers, H.A., and Wilcox, R.E., 1964, Volcanic ash from Mount Mazama (Crater Lake) and from Glacier Peak: Science, v. 144, p. 1334-1336.

Rigg, G.B., and Gould, H.R., 1957, Age of Glacier Peak eruption and chronology of postglacial peat deposits in Washington and surrounding areas: American Jour. of Science, v. 225, p. 341-363.

Robinson, J.W., 1938, General geology of the Lake Cushman and Hamma Hamma River area of the Olympic Mountains: Unpub. B.S. thesis, Univ. Wash., 32 p.

Rodgers, W.P., 1970, A geological and geophysical study of the central Puget Lowland: Unpub. Ph.D. dissertation, Univ. Wash., 123 p.

Rosengreen, T.E., 1965, Surficial geology of the Maple Valley and Hobart quadrangles, Washington: Unpub. M.S. thesis, Univ. Wash., 71 p.

Sceva, J.E., 1957, Geology and ground water resources of Kitsap County, Washington: U.S. Geol. Survey Water-Supply Paper 1413, 178 p.

Todd, M.R., 1939, Glacial geology of the Hamma Hamma Valley and its relation to the glacial history of the Puget Sound Basin: Unpub. M.S. thesis, Univ. Wash., 48 p.,

Vine, J.D., 1969, Geology and coal resources of the Cumberland, Hobart, and Maple Valley quadrangles, King County, Washington: U.S. Geol. Survey Prof. Paper 624, 67 p.

Waldron, H.H., 1961, Geology of the Poverty Bay quadrangle, Washington: U.S. Geol. Survey Map GQ-158, scale 1:24,000.

______1962, Geology of the Des Moines quadrangle, Washington: U.S. Geol. Survey Map GQ-159, scale 1:24,000.

______1967, Geologic map of the Duwamish Head quadrangle, King and Kitsap Counties, Washington: U.S. Geol. Survey Map GQ-706, scale 1:24,000.

Waldron, H.H., Mullineaux, D.R., and Crandell, D.R., 1957, Age of the Vashon glaciation in the southern and central parts of the Puget Sound basin, Washington: Geol. Soc. of America Bull., v. 68, p. 1849-1850.

Walters, K.L., and Kimmell, G.E., 1968, Ground water occurrence and stratigraphy of unconsolidated deposits, central Pierce County, Washington: Wash. Dept. of Water Resources, Water-Supply Bull. 22, 428 p.

Washburn, A.L., and Stuiver, Minze, 1962, Radiocarbon-dated post-glacial deleveling in Northeast Greenland and its implications: Arctic, v. 15, p. 66-72.

Weertman, Johannes, 1964, The theory of glacier sliding: Jour. of Glaciology, v. 5, p. 287-303.

Williams, V.S., 1971, Glacial geology of the drainage basin of the Middle Fork of the Snoqualmie River: Unpub. M.S. thesis, Univ. Wash., 45 p.