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Mass Balance and Sliding Velocity of the Puget Lobe of the Cordilleran Ice Sheet during the Last Glaciation

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An estimate of the sliding velocity and basal meltwater discharge of the Puget lobe of the Cordilleran ice sheet can be calculated from its reconstructed extent, altitude, and mass balance. Lobe dimensions and surface altitudes are inferred from ice limits and flow-direction indicators. Net annual mass balance and total ablation are calculated from relations empirically derived from modern maritime glaciers. An equilibrium-line altitude between 1200 and 1250 m is calculated for the maximum glacial advance (ca. 15,000 yr B.P.) during the Vashon Stade of the Fraser Glaciation. This estimate is in accord with geologic data and is insensitive to plausible variability in the parameters used in the reconstruction. Resultant sliding velocities are as much as 650 m/a at the equilibrium line, decreasing both up- and downglacier. Such velocities for an ice sheet of this size are consistent with nonsurging behavior. Average meltwater discharge increases monotonically downglacier to 3000 m^3/sec at the terminus and is of a comparable magnitude to ice discharge over much of the glacier's ablation area. Paleoclimatic inferences derived from this reconstruction are consistent with previous, independently derived studies of late Pleistocene temperature and precipitation in the Pacific Northwest.

INTRODUCTION

Most modern glaciers not frozen to their beds are believed to slide, with measured velocities ranging from a few meters to more than a kilometer per year (Paterson, 1981, Table 5.1). Common features of glaciated landscapes, such as striations and streamlined bedforms, indicate that erosion caused by the movement of Pleistocene ice masses has often dominated other geomorphic processes. Estimates of the rate of basal sliding is important to understanding these glacially sculpted forms. Estimates of basal sliding also provide information on the volume and longitudinal variation of subglacial meltwater and an independent basis for analyzing the climatic implications of fluctuations in the glacier's terminus.

The Puget Lowland of western Washington offers an excellent opportunity to develop and apply a useful technique for calculating basal sliding rates. Ice invaded the region most recently during the Vashon Stade of the Fraser Glaciation (Armstrong *et al.*, 1965) about 15,000–14,000 yr ago (Rigg and Gould, 1957; Mullineaux *et al.*, 1965). The Puget lobe of the Cordilleran ice sheet flowed southward in the broad trough between the Olympic Mountains and Cascade Range, which is now partly occupied by Puget Sound. The adjoining Juan de Fuca lobe flowed westward along the topographic low now occupied by the Strait of Juan de Fuca and terminated west of the modern coastline of northwest Washington (Fig. 1).

Calculation of the equilibrium mass balance of the ice sheet forms the basis of this analysis (cf. Pierce, 1979). For a glacier in equilibrium, net accumulation above any transect perpendicular to flow must be transferred by ice discharge through that transect to replenish the net ablation downglacier from it. This mass-balance method requires a reconstruction of the physical boundaries of the ice mass, topographic contours of its surface, and a relationship between specific net balance and

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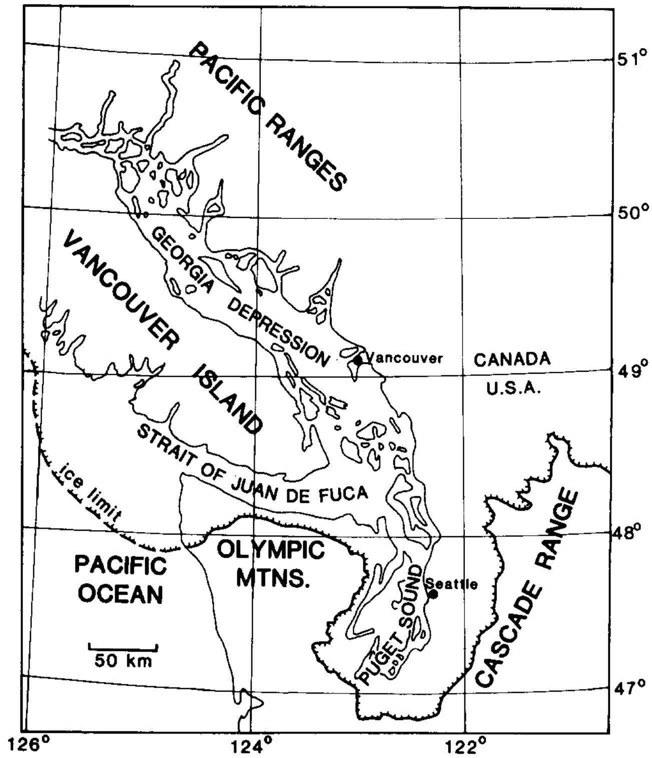

FIG. 1. Index map of the region covered by the southern part of the Cordilleran ice sheet at maximum stage (ca. 15,000 yr B.P.).

altitude on the ice sheet. An equilibrium-line altitude (ELA) can then be found that brings the reconstructed glacier as a whole into balance. The rate of ice flux through any cross section can then be simply calculated. An equivalent relationship between *total* ablation and altitude on the ice sheet also permits a quantitative estimate of meltwater discharge.

Although this reconstruction contains considerable uncertainties, the inferred order of magnitude of the sliding rate is rather insensitive to them. This is noteworthy, because the geologic significance of calculated basal sliding velocities depends less on their precise value than on their overwhelming dominance in relation to two other ice velocities intrinsic to any glacier or ice sheet.

The first such velocity is the average rate of ice flow solely from internal deformation caused by viscous creep. The vertically averaged longitudinal velocity for a non-sliding glacier is determined by integrating the equation for internal deformation at any level in the ice [Nye, 1952, Eq. (4); Paterson, 1981, p. 87]:

$$\bar{u} = 2hA\,(\tau_b)^n/(n + 2), \qquad (1)$$

where $\bar{u}$ is the average ice velocity, A and n are flow-law parameters (Glen, 1954), h is the ice thickness, and τ_b the basal shear stress. The value of τ_b for any ice sheet can be calculated from its reconstructed longitudinal profile by

$$\tau_b = \rho\, gh \sin \alpha, \qquad (2)$$

where ρ is the ice density, g the acceleration due to gravity, and α the surface slope. Analysis of a detailed reconstruction of the Puget lobe below the 1300-m contour by Thorson (1980, Fig. 3) shows a value of τ_b slightly below 50 kPa (0.5 bar). Values determined from the regional reconstruction of the Cordilleran ice sheet (see below) average about 10% less. Using likely maximum values of τ_b = 50 kPa and h = 1500 m, and flow parameters n = 3 and A = 5.3 $\times$ 10^{-15} sec^{-1} kPa^{-3} (Paterson, 1981, p. 39), Equations (1) and (2) give

$$\bar{u} = 11 \text{ m/a} \quad (3)$$

as the maximum averaged horizontal velocity caused solely by viscous creep.

The second intrinsic ice velocity of geologic importance is the rate of vertical flow at the bed caused by basal melting. Various subglacial processes, such as the abrasion of bedrock and the lodgment of clasts on rock surfaces, depend on whether the ratio of horizontal to vertical ice velocity at the bed is close to, or much larger than, unity (Hallet, 1979). Causes of basal melting include goethermal heat flow, friction from sliding, and flow of subglacial water. Spatially averaged, their combined magnitude is at most a few tenths of a meter per year (Rothlisberger, 1968).

RECONSTRUCTION OF THE ICE SHEET (FIG. 2)

Ice Limits

Ice limits for the southern boundary of the ice sheet on land have been rather ac-

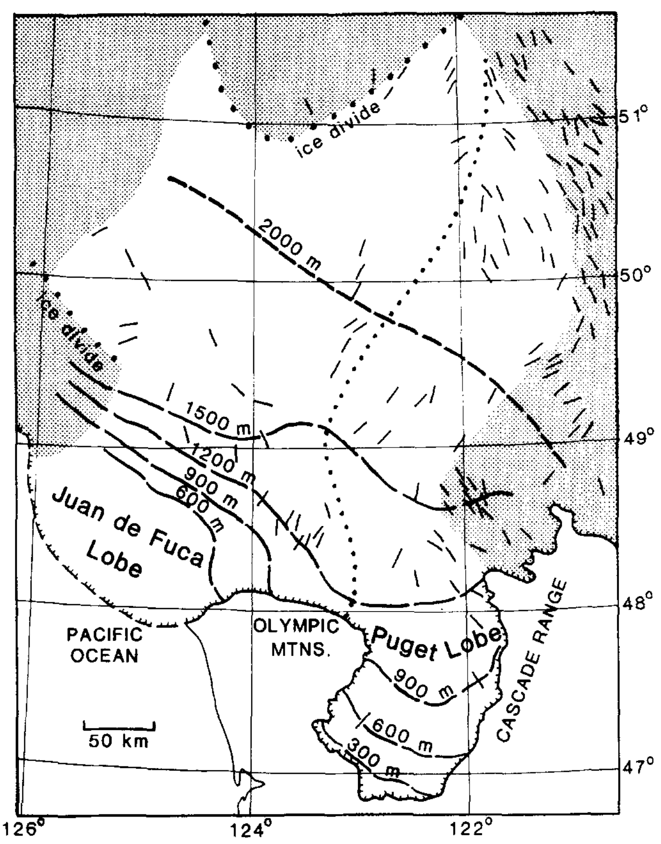

FIG. 2. Reconstruction of the Puget and Juan de Fuca lobes of the Cordilleran ice sheet at maximum stage. Short lines indicate the orientation of some representative striations and other glacially elongated topography; the hachured line shows maximum extent of ice. Heavy lines are contours of ice-surface altitude above modern sea level, uncorrected for glacial-age sea level or isostatic depression. The unshaded ice-covered area includes all ice of the Puget and Juan de Fuca lobes. Ice of these two lobes is inferred to have been separated along the dotted line. Sources of data include Wilson *et al.* (1958), Prest *et al.* (1968), Thorson (1980), Heller (1980), Dethier *et al.* (1981), Fred Pessl, Jr. (written communication, 1983), and Booth (1985, 1986, and unpublished data).

curately determined from extensive mapping. Thorson (1980) and Waitt and Thorson (1983) have most recently compiled data for the Puget lobe and the south boundary of the Juan de Fuca lobe. The Puget lobe terminated above sea level, whereas the Juan de Fuca lobe terminus was probably tidewater. The now-submerged western boundary of the Juan de Fuca lobe (Clague, 1981) has been inferred (Alley and Chatwin, 1979) to coincide with the edge of the continental shelf southwest of Vancouver Island (U.S. Department of Commerce, National Ocean Survey Chart No. 18480). The presence of submerged lobate ridges (e.g., La Perouse Bank) on the outer portion of the continental shelf is consistent with this interpretation.

Flow Boundaries

Flow-direction indicators, in the form of striations and glacially elongated topography, define a consistent pattern of ice flow throughout the area. Flow boundaries can be drawn between ice of the Puget–Juan de Fuca system and ice to the north, west, and east. The Puget and Juan de Fuca lobes are not as well differentiated by geologic data upglacier from the obvious divergence at the northeast corner of the Olympic Mountains. Consequences of this uncertainty are discussed below.

Ice-Surface Contours

Contours on the southern part of the Puget lobe follow Thorson's (1980) reconstruction and are mostly well constrained by ice-margin altitudes and abundant ice-flow direction indicators throughout the lowland. Contours in Canada are largely from the Glacial Map of Canada (Wilson *et al.*, 1958). Waitt and Thorson (1983) show ice-surface contours in the northern Cascade Range that project into Canada at altitudes up to a few hundred meters higher. The compilation of Figure 2 attempts to reconcile these sources, following the basic principle that the ice-surface contours should be perpendicular to the direction of indicated flow. In addition, flow lines should not converge or diverge without commensurate changes in ice thickness or net balance (in order to conserve mass); and the local basal shear stress, determined by the product of the reconstructed ice thickness and surface slope [Eq. (2)], should vary only gradually along the glacier's length.

MASS-BALANCE CALCULATIONS

Height/Mass-Balance Relationship

Local mass balance of modern glaciers is plotted in Figure 3 as a function of height above or below the equilibrium line. This curve was derived from the averaged

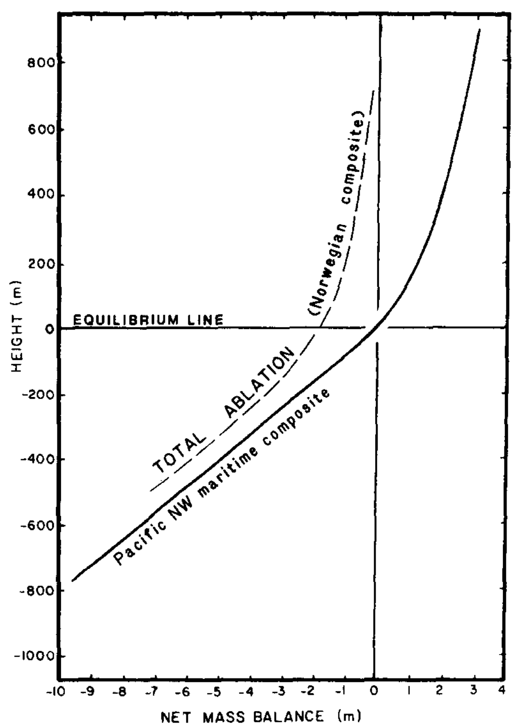

FIG. 3. Height mass-balance relationship for modern maritime glaciers. Height is referenced to the equilibrium-line altitude; specific balance is in cubic meters of equivalent water volume per square meter of ice surface per year. Original data for net balance are from Pacific Northwest glaciers (Meier *et al.*, 1971), summarized in Porter *et al.* (1983); total ablation data are adapted from two Norwegian maritime glaciers (Nigardsbreen and Tunsbergdalsbreen) (IAHS, 1977, pp. 170–175).

values of seven modern maritime glaciers in the Pacific Northwest and Alaska (Meier *et al.*, 1971; Porter *et al.*, 1983). The good correlation among these data, representing a variety of local environments, lends some confidence to their application to an obviously larger Pleistocene ice mass of unknown mass-balance gradient. An additional relationship is provided in Figure 3 between total ablation and height in relation to the ELA, derived from good but limited data from two Norwegian maritime glaciers (IAHS, 1977). These data allow a quantitative estimate of meltwater discharge through transverse cross sections of the glacier as well as the corresponding values of ice discharge.

Equilibrium-Line Altitude Calculations

The areas of each region shown in Figure 2, bounded by contour lines, ice limits, and lobe-dividing flow lines, are given in Table 1. The altitude of each region was specified for all subsequent calculations by the average of its limiting values. The region above 2000 m was assigned a representative altitude of 2300 m, based on nunatak altitudes (Wilson *et al.*, 1958) and plausible ice-surface profiles. Total annual accumulation and ablation volumes are calculated for three values of the ELA (Table 2). Equilibrium, which is assumed to approximate conditions during ice-maximum time, is attained with the ELA equal to ca. 1225 m. The predicted accumulation-area ratio (0.7) is within the range reported for modern glaciers (Meier and Post, 1962; Porter, 1975).

Sliding Velocity

On any glacier the ice accumulated above the equilibrium line must be transported into the ablation area. The flux will be greatest at the equilibrium line and decrease through successive downglacier cross sections, because increasing volumes of glacier mass are lost through surface ablation. The inferred pattern of ice flow in the Juan de Fuca lobe (Fig. 2) is sufficiently complex to introduce large lateral variability in this flux. The simpler flow pattern of the Puget lobe, however, is well suited to flow calculations based on accumulation and ablation volumes. Using an ELA of 1225 m (Table 2), 7.9×10^{10} m^3 of ice must cross the equilibrium line each year for the Puget lobe to remain in a balanced state (Table 3). The cross-sectional area at the equilibrium line is 1.2×10^8 m^2 (approximately the same area as at 1200 m; cf. Table 3), giving an average transport rate of 660 m/a. As only 11 m/a can be accounted for by internal deformation of the ice [Eq. (3)], the sliding velocity is approximately 650 m/a, or 98% of the total velocity. Average rates of this order are estimated downglacier as well, for although flux is decreasing, the cross-sectional area is also shrinking. For example, where the ice-surface altitude is 900 m (about the latitude of Seattle), the average transport rate is still over 500 m/a.

TABLE 1. AREA OF ICE LOBES, DETERMINED BY PLANIMETRY (FIG. 2)

Altitude interval (m)	Area of Juan de Fuca lobe (10^3 km^2)	Area of Puget lobe (10^3 km^2)
0–300	10.1	0.8
300–600		2.4
600–900	3.2	3.8
900–1200	4.3	6.0
1200–1500	6.5	7.8
1500–2000	23.9	8.6
>2000	27.7	12.4

Water Flux

Estimated water production (calculated from the total-ablation curve in Fig. 3) and cumulative downglacier flux are shown in Table 4. These values increase from a negligible value, consistent with the dry conditions observed high on modern glaciers (Mayo, 1984), to values well in excess of the ice flux. Because this water will tend to flow in channels (Shreve, 1972), its geomorphic effects probably will be locally more pronounced than those of the sliding

TABLE 2. NET BALANCE VALUES IN 10^{10} m^3/a OF EQUIVALENT WATER VOLUME

Altitude interval (m)	ELA = 1250 m		ELA = 1200 m		ELA = 1150 m	
	J de F	Puget	J de F	Puget	J de F	Puget
0–300	−12.9	−1.2	−12.3	−1.1	−11.6	−1.0
300–600		−2.4		−2.2		−2.1
600–900	−2.0	−2.4	−1.8	−2.1	−1.6	−1.9
900–1200	−1.1	−1.5	−0.8	−1.1	−0.5	−0.8
1200–1500	0.7	0.8	0.8	0.9	0.9	1.1
1500–2000	5.4	2.0	5.6	2.1	5.9	2.1
>2000	9.5	4.3	9.8	4.4	10.1	4.5
Net balance	−0.4	−0.4	+1.3	+0.9	+3.2	+1.9

Note. Positive values indicate net accumulation, negative values net ablation. Calculations are based on areas between contours (Table 1) and specific net mass balance relative to ELA shown in Figure 3. Both lobes are brought into balance by an ELA between 1250 and 1200 m. An ELA value of 1225 m is therefore assumed for all subsequent calculations. Note that these values differ from those used in the discussions of ice volume by a density factor of approximately 10%.

ice regardless of relative discharges (Booth, 1984a, 1984b).

CORROBORATING EVIDENCE

Estimated Equilibrium-Line Altitude

The glacier-wide pattern of ice flow, as expressed by directional indicators, should show marked differences above and below the equilibrium line. Above the equilibrium line, flow should diverge from all ice-sheet boundaries, as all points over the ice surface contribute to the net increase in mass. Only locally, such as where an ice tongue projects up an ice-free alpine valley, can an additional source of heat (i.e., alpine rivers) yield a net loss of mass. Below the equilibrium line, all points on the glacier experience a net loss, and so flow must converge with the ice margins. Lateral ice-surface gradients, expressed by convex downglacier contours, drive this flow. The association of compressive flow with the ablation area (Nye, 1957) also dictates this transverse profile, in order to satisfy mass conservation over a glacier with stable margins. Inspection of the data compiled for Figure 2 shows generally convex-down contours and margin-convergent flow directions up to at least 1100 m in altitude along the eastern edge of the Puget lobe, and less definitively up to 1200 m over the entire ice sheet.

The longitudinal change in glacier mass

TABLE 3. ICE DISCHARGE AND VELOCITY THROUGH TRANSVERSE SECTIONS OF THE PUGET LOBE, ELA = 1225 m

Contour interval (m)	Ice width (10^5 m)	Average thickness (10^3 m)	Cross-sectional area (10^8 m^2)	Ice discharge (10^{10} m^3/a of ice)	Ice velocity (m/a)
2000	1.1	1.0	1.1	4.8	430
1500	1.1	1.1	1.2	7.0	580
1200	1.0	1.2	1.2	7.9	660
900	1.3	0.9	1.2	6.4	540
600	1.1	0.7	0.8	3.8	470
300	1.0	0.4	0.4	1.2	310
Terminus	1.0	0	0	0	0

Note. Sections are taken along contour lines shown on Figure 2. Ice velocity through these sections will be overwhelmingly by basal sliding (see text).

TABLE 4. ICE AND WATER DISCHARGES THROUGH TRANSVERSE SECTIONS OF THE PUGET LOBE, ELA = 1225 m

Contour (m)	Ice discharge (10^{10} m^3/a of ice)	Cumulative water discharge (10^{10} m^3/a)	Water discharge/ice discharge (by volume)
2000	4.8	0.0	0
1500	7.0	0.4	1/17
1200	7.9	1.4	1/6
900	6.4	3.5	1/2
600	3.8	6.1	1.6/1
300	1.2	8.6	7/1
Terminus	0	9.7	∞

Note. Ice-discharge data from Table 3. Ablation thickness from total ablation of "Norwegian composite" (Fig. 3) and ignores the opposing effects of evaporation and rainfall on the water-discharge values.

or in cross-sectional area, in those portions of an ice sheet with near-parallel lateral boundaries, also suggests the region in which net accumulation replaces net ablation. Because discharge must reach a maximum at the equilibrium line, cross-sectional areas (as well as velocity) should also tend to follow this pattern, analagous to the principle of minimum variance in river discharge (Leopold and Langbein, 1962). From Figure 2 and the underlying topography, cross-sectional areas of the Puget lobe do show a maximum between the 1200- and 1500-m contours. This suggests an ELA within this range (the apparent narrowing of the Puget lobe just south of 1500 m reflects the steep southward descent of the Pacific Ranges in this area and not a decreasing cross-sectional area).

Estimated Sliding Velocity

Constraint of the Puget lobe ELA by glacial–geologic inferences and data confirms broadly the more detailed calculations made in the previous section based on the height/mass balance curve (Fig. 3). The advantage of the quantitative mass-balance approach is that it provides an estimate of sliding velocity as well. This estimated sliding velocity can itself be compared with other indirect geological data. Near the latitude of Seattle (47°30′), bounding dates on the Vashon-age advance of 15,100 ± 300 yr B.P. (Mullineaux *et al.*, 1965, ^{14}C sample W-1305) and 13,650 ± 550 yr B.P. (Rigg and Gould, 1957, ^{14}C sample L-346) require that the ice sheet advanced and retreated on average more than 80 km in each direction during the intervening 870–2130 yr. Assuming an advance rate half that of the retreat (Weertman, 1964), minimum ice-margin advance rates of 50 to more than 200 m/a are required (equal advance and retreat rates increase this value by approximately one-third). Because ice ablated rapidly at low altitudes before even reaching the terminus, the total flux of ice at the equilibrium line must have exceeded significantly this value to sustain the advance.

Water Flux

Under equilibrium conditions, the average yearly water discharge for the entire ice sheet should nearly equal the total precipitation that falls on its surface. Although direct evidence of Pleistocene precipitation values are absent, Porter (1977) has used pollen data and snow-line-depression reconstructions to argue for a change of less than 30% in relation to modern values. In the region previously occupied by the Puget lobe, precipitation now ranges from 1 to nearly 3 m/a (U.S. Dept. of Commerce, 1979). During the glacial maximum the increased altitude of the ice surface probably resulted in greater precipitation than occurs at modern sea-level stations. Therefore, the best modern analogs are probably those stations in the Cascade Range on the seaward (west-facing) slopes. Their reported values are 2.1 ± 0.7 m/a (Porter, 1977, p. 106). The reconstructed Puget-lobe area of 41,800 km^2 (Table 1) divided into the predicted yearly water discharge of 97 km^3/a (Table 4) gives a past precipitation rate of 2.3 m/a, remarkably consistent with modern values and well within the limits required by the independent pollen and snow-line data.

SENSITIVITY

Calculations of Equilibrium-Line Altitude

Ice-Sheet Reconstruction

The boundaries and contours of the Puget lobe below about 1200 m are well constrained by several geologic studies along the ice-sheet margin (Crandell, 1963; Rosengreen, 1965; Carson, 1970; Thorson, 1980; Booth, 1986). The weak dependence of accumulation on altitude (Fig. 3; Mayo, 1984) deemphasizes the large uncertainty in ice-surface altitude above 2000 m (e.g., Waitt and Thorson, 1983). The flow boundaries chosen for the east and west sides of the Puget lobe upglacier of 1200-m altitude are consistent with flow-directional indicators. The position of the western boundary is further constrained by requiring that calculated ELAs of the reconstructed Puget and Juan de Fuca lobes are equal, which ignores a possible contribution to the total Juan de Fuca mass budget from calving (see below).

The likelihood of temperature and precipitation gradients across the ice sheet is ignored by the assumed equivalence of Puget and Juan de Fuca ELAs. The consequences of this assumption, however, are not severe. An east–west precipitation gradient, in particular, is expected from modern data across topographic barriers. However, a rise of only 25 m in the Puget-lobe ELA in relation to the Juan de Fuca lobe requires a decrease of 20% in the area of the Puget lobe (by adjustment of its western boundary with the Juan de Fuca lobe) to maintain equilibrium. More extreme changes do not appear possible from the flow-direction indicators. An equivalent increase in the Puget-lobe area, at the expense of the Juan de Fuca lobe, also appears implausible from the pattern of directional indicators. Its effect would be an equally trivial, and somewhat anomalous, decrease in ELA of the more landward lobe. Porter (1977) shows that accumulation-season precipitation and mean annual temperature are strongly correlated, so that the anticipated east–west gradient in the first may be closely compensated by corresponding changes in the second.

Alternatively, calving at the terminus of the Juan de Fuca lobe may have partly compensated for an eastward decrease in precipitation. A crude estimate of calving rates, based on those measured at modern tidewater glaciers in Alaska (Brown *et al.*, 1982), suggests that the order of magnitude of this process could have been equivalent to that of the excess yearly budget due to a 25-m drop in the ELA of the Juan de Fuca lobe (Table 2). The inflexibility of the reconstructed Puget-lobe ELA may therefore reflect a balance between these tendencies rather than an absence of spatial variability.

Mass-Balance Relationship

The relation between altitude and net mass balance critically determines both the ELA and the mass flux through the equilibrium line. Use of a curve (Fig. 3) assembled from relatively small glaciers must only approximate the conditions over a continental ice sheet, particularly the consequences of increasing distances between the accumulation area and the source of moisture. The estimates derived from it, however, may still prove satisfactory. As noted previously, the balance curves of these glaciers cluster closely about a single set of values (Meier *et al.*, 1971) despite their variety of geographic locations and ELAs, lending some confidence to the generality of this relationship.

Additional data from the 150-km-long Malaspina–Seward–Hubbard glacier system can address qualitatively the problems of scaling up this mass-balance relationship. Investigations by Marcus and Ragle (1970) on these glaciers show that total precipitation at the ice divide, 150 km from the coastline, is 15% less than at a point 90 km closer to the ocean but nearly twice the value recorded at a station only 12 km closer to the coast. Thus, local topography and exposure strongly influence single measurements far beyond the effects of altitude and increasing continentality, which for ocean-facing slopes have opposing influence. Thus a composite curve, representing multiple measurements in a variety of microenvironments, is probably more trustworthy at all scales.

A final check can be made on this source of uncertainty. An upper bound on the accumulation rate can be set, suggested by some observational and theoretical evidence of an "elevation desert" high on an ice sheet (e.g., Budd and Smith, 1981) and by the paucity of data for these regions compiled for Figure 3. If a limit of 2 m/a is imposed, the ELA of the Puget lobe must drop by 100 m to remain in equilibrium. This appears to fall outside the range of geologically plausible values and thus suggests a lower limit for the accumulation rate far upglacier.

Sliding Velocity

Calculated sliding velocities are mostly insensitive to any plausible range in reconstructed ELAs. This can be demonstrated by computing alternative sliding velocities using a range of assumed ELAs and the height/mass-balance relationship (Fig. 3) for the ablation area only. The latter is likely to be influenced only slightly by the total length of the ice sheet, reducing the potential inaccuracy of data from smaller modern glaciers. Ice-sheet boundaries and contours are also much better constrained at these lower altitudes. The range of plausible ELAs (1200–1500 m) is constrained by geologic evidence (see above) independent of any assumed mass-balance relationship. For an ELA = 1200 m, the flux through the equilibrium line is 7.3×10^{10} m^3/a, giving a sliding velocity of 600 m/a (92% of the value for an ELA = 1225 m). For an ELA = 1500 m, the flux is 10.8×10^{10} m^3/a, giving a velocity at the equilibrium line of 890 m/a (a 40% increase). Even

TABLE 5. ICE VELOCITIES AT THE EQUILIBRIUM LINE OF THE PUGET LOBE FOR VARIOUS ESTIMATES OF THE ELA

ELA	Average ice velocity (m/a)	Ice velocity due to internal deformation [Eq. (3)] (m/a)	% of total flow due to basal slip
1125	500	11	98
1200	610	11	98
1225	660	11	98
1500	900	13	99

using an ELA of 1125 m, required by a 2 m/a limit on accumulation rates, the velocity is still 500 m/a. Sliding velocities therefore exceed the other characteristic velocities in the glacier, namely the basal melt rate and the internal deformation rate, by one or more orders of magnitude regardless of the estimate used (Table 5).

DISCUSSION

Climatic Implications

Predicted high mass flux, expressed as high sliding rates, provides insight into the relatively brief occupancy of the Lowland by the Puget lobe during the Vashon Stade. Because most of the ice that accumulated above the equilibrium line merely replaced ice lost below, fluctuations in the ice front at rates of 100 m/a represent only minor perturbations in the total budget of the ice sheet. For example, if the ELA were to move from the hypothesized 1225-m level to the 1250-m level, a change of only 3% in the 900-m total Fraser-age ELA depression (Porter, 1977), the subsequent yearly deficit of 5×10^9 m^3 (cf. Table 2) would be equivalent to an averaged retreat rate of ca. 50 m/a (deficit divided by average cross-sectional area). Although this ignores all complexities of an ice sheet's response to nonequilibrium conditions (Nye, 1963), it clearly demonstrates the sensitivity of this body to small climatic changes and the plausibility of rapid movement of the ice front without equally precipitous climatic changes.

This analysis also lends credence to several prior studies of late Pleistocene climate in the Pacific Northwest based on completely independent data. The predicted rate of transfer of ice and water by the ice sheet requires basal sliding and thus temperate conditions over all of its bed. The thermal regime is therefore not analogous to the cold-based margin assumed for the mid-continent Laurentide ice sheet (Moran *et al.*, 1980). Mean annual temperatures are predicted by pollen assemblages just south of the Juan de Fuca lobe (Heusser, 1972) to have been ca. 5°C colder than at present (which average about 10°C throughout the lowlands of western Washington). The observed correlation of increased precipitation with increased temperature (Porter, 1977), discussed above, is supported by the physical constraints on any significant east–west change in reconstructed ELAs for the Puget and Juan de Fuca lobes. Finally, the close approximation of past and present precipitation values, previously hypothesized from the depression of Cascade snow lines during the late Pleistocene (Porter, 1977), is further supported by the apparent coincidence of modern precipitation rates and predicted ice-sheet runoff.

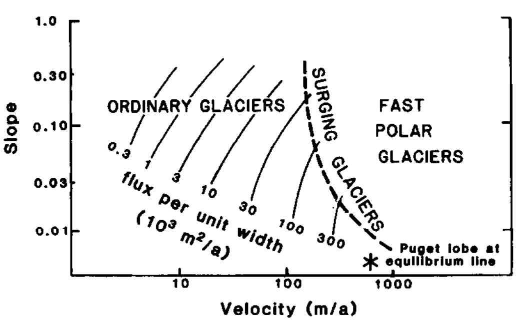

FIG. 4. Comparison of the Puget lobe with a compilation of velocity measurements on modern glaciers, modified from Budd (1975). "Ordinary" (nonsurging) and surging glaciers can be discriminated on this graph by a line (dashed) of power per unit bed area (= basal shear stress × velocity = 5.4×10^7 J/m² a). The reconstructed parameters of the Puget lobe show an energy-dissipation rate that is only two-thirds of this empirically determined transition value.

Surging of the Puget Lobe

The rapid advance and predicted high sliding velocities of the Puget lobe invite speculation that the ice sheet may have moved erratically, analogous on a vast scale with modern surging valley glaciers (Post, 1969). Although these reconstructed rates are far higher than those reported for most modern nonsurging glaciers, they apparently do not require surging behavior, based on admittedly limited understanding of the controlling glacial mechanics. Budd (1975) has classified glaciers as "ordinary," "fast," and "surging," depending on whether they can continuously resupply the volume of ice transported downglacier by sliding. Within this framework, he finds that modern glaciers can be divided into "ordinary" (nonsurging) and surging types both observationally and by a line of constant energy-loss rate (shear stress × velocity) over a wide range of surface slopes, velocities, and ice thicknesses. These data, reproduced in Figure 4, include the reconstructed value of the Puget lobe at the equilibrium line. They show that in spite of high predicted velocity, this ice sheet lies well within the field of "ordinary" glaciers, having only two-thirds of the energy-loss rate of observed surging glaciers.

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REFERENCES

Alley, N. F., and Chatwin, S. C. (1979). Late Pleistocene history and geomorphology, southwestern Vancouver Island, British Columbia. *Canadian Journal of Earth Sciences* **16**, 1645–1657.

Armstrong, J. E., Crandell, D. R., Easterbrook, D. J., and Noble, J. B. (1965). Late Pleistocene stratigraphy and chronology in southwestern British Columbia and northwestern Washington. *Geological Society of America Bulletin* **76**, 321–330.

Booth, D. B. (1984a). "Glacier Dynamics and the Formation of Glacial Landforms in the Eastern

Puget Lowland, Washington." Unpublished Ph.D. dissertation, University of Washington.

Booth, D. B. (1984b). Ice-sheet reconstruction and erosion by subglacial meltwater in the eastern Puget lowland, Washington. *Geological Association of America Abstracts with Programs* **16**, 450.

Booth, D. B. (1985). "Surficial Geologic Map of the Granite Falls 15-Minute Quadrangle, Washington." U.S. Geological Survey Open-File Report 85–517.

Booth, D. B. (1986). "Surficial Geology of the Skykomish and Snoqualmie Rivers Area, Snokomish and King Counties, Washington." U.S. Geological Survey Miscellaneous Investigations Map I-1745.

Brown, C. S., Meier, M. F., and Post, Austin (1982). "Calving Speed of Alaska Tidewater Glaciers, with Application to Columbia Glacier." U.S. Geological Survey Professional Paper 1258C.

Budd, W. F. (1975). A first simple model for periodically self-surging glaciers. *Journal of Glaciology* **14**, 3–21.

Budd, W. F., and Smith, I. N. (1981). The growth and retreat of ice sheets in response to orbital radiation changes. *International Association of Hydrologic Sciences* **131**, 369–409.

Carson, R. J., III (1970). "Quaternary Geology of the South-Central Olympic Peninsula, Washington." Unpublished Ph.D. dissertation, University of Washington.

Clague, J. J. (1981). "Late Quaternary Geology and Geochronology of British Columbia, 2." Geological Survey of Canada, paper 80–35.

Crandell, D. R. (1963). "Surficial Geology and Geomorphology of the Lake Tapps Quadrangle, Washington." U.S. Geological Survey Professional Paper 388A.

Dethier, D. P., Safioles, S. A., and Pevear, D. R. (1981). "Composition of Till from the Clear Lake Quadrangle, Skagit and Snohomish Counties, Washington." U.S. Geological Survey Open-File Report 81–517.

Glen, J. W. (1954). The stability of ice-dammed lakes and other water-filled holes in glaciers. *Journal of Glaciology* **2**, 316–318.

Hallet, B. (1979). A theoretical model of glacial abrasion. *Journal of Glaciology* **23**, 39–50.

Heller, P. L. (1980). Multiple ice-flow directions during the Fraser Glaciation in the lower Skagit River drainage, North Cascade Range, Washington. *Arctic and Alpine Research* **12**, 299–308.

Heusser, C. J. (1972). Palynology and phytogeographical significance of a late-Pleistocene refugium near Kalaloch, Washington. *Quaternary Research* **2**, 189–202.

International Association of Hydrological Sciences (1977). "Fluctuations of glaciers," Vol. 3. International Commission on Snow and Ice, Paris, IAHS-UNESCO.

Leopold, L. B., and Langbein, W. B. (1962). "The Concept of Entropy in Landscape Evolution." U.S. Geological Survey Professional Paper 500A.

Marcus, M. G., and Ragle, R. H. (1970). Snow accumulation in the Icefield Ranges, St. Elias Mountains, Yukon. *Arctic and Alpine Research* **2**, 277–292.

Mayo, L. R. (1984). Glacier mass balance and runoff research in the U.S.A. *Geografisker Annelar* **66A**, 215–227.

Meier, M. F., and Post, A. S. (1962). "Recent Variations in Mass Net Budgets of Glaciers in Western North America." International Association of Scientific Hydrology Publication 58, pp. 63–77.

Meier, M. F., Tangborn, W. V., Mayo, L. R., and Post, A. S. (1971). "Combined Ice and Water Balances of Gulkana and Wolverine Glaciers, Alaska, and South Cascade Glacier, Washington, 1965 and 1966 Hydrologic Years." U.S. Geological Survey Professional Paper 715A.

Moran, S. R., Clayton, L., Hooke, R. LeB., Fenton, M. M., and Andriashek, L. D. (1980). Glacier-bed landforms of the prairie region of North America. *Journal of Glaciology* **25**, 457–476.

Mullineaux, D. R., Waldron, H. H., and Rubin, M. (1965). "Stratigraphy and Chronology of Late Interglacial and Early Vashon Glacial Time in the Seattle Area, Washington." U.S. Geological Survey Bulletin 1194-0.

Nye, J. F. (1952). The mechanics of glacier flow. *Journal of Glaciology* **2**, 82–93.

Nye, J. F. (1957). The distribution of stress and velocity in glaciers and ice-sheets. *Proceedings of the Royal Society of London Ser. A* **239**, 113–133.

Nye, J. F. (1963). The response of a glacier to changes in the rate of nourishment and wastage. *Proceedings of the Royal Society of London Ser. A* **275**, 87–112.

Paterson, W. S. B. (1981). "The Physics of Glaciers." Pergamon, Oxford.

Pierce, K. L. (1979). "History and Dynamics of Glaciation in the Northern Yellowstone National Park Area." U.S. Geological Survey Professional Paper 729F.

Porter, S. C. (1975). Equilibrium-line altitudes of late Quaternary glaciers in the southern Alps, New Zealand. *Quaternary Research* **5**, 27–47.

Porter, S. C. (1977). Present and past glaciation threshold in the Cascade Range, Washington, U.S.A.: Topographic and climatic controls, and paleoclimatic implications. *Journal of Glaciology* **18**, 101–116.

Porter, S. C., Pierce, K. L., and Hamilton, T. D. (1983). Late Wisconsin mountain glaciation in the western United States. *In* "Late-Quaternary Environments of the United States" (H. E. Wright, Jr., and S. C. Porter, Eds.), pp. 71–111. Univ. of Minnesota Press, Minneapolis.

Post, A. S. (1969). Distribution of surging glaciers in western North America. *Journal of Glaciology* 8, 229–240.

Prest, V. K., Grant, D. R., and Rampton, V. N. (compilers) (1968). "Glacial Map of Canada."

Rigg, G. B., and Gould, H. R. (1957). Age of Glacier Peak eruption and chronology of postglacial peat deposits in Washington and surrounding areas. *America Journal of Science* 255, 341–363.

Rosengreen, T. E. (1965). "Surficial Geology of the Maple Valley and Hobart Quadrangles, Washington." Unpublished M.S. thesis, University of Washington.

Rothlisberger, H. (1968). Erosive processes which are likely to accentuate or reduce the bottom relief of valley glaciers. *International Association of Scientific Hydrologists* 79, 87–97.

Shreve, R. L. (1972). Movement of water in glaciers. *Journal of Glaciology* 11, 205–214.

Thorson, R. M. (1980). Ice sheet glaciation of the Puget lowland, Washington, during the Vashon Stade. *Quaternary Research* 13, 303–321.

U.S. Department of Commerce (1979). "World Weather Records 1961–1970," Vol. 1. U.S. Government Printing Office.

Waitt, R. B., Jr., and Thorson, R. M. (1983). The Cordilleran ice sheet in Washington, Idaho, and Montana. *In* "Late-Quaternary Environments of the United States" (H. E. Wright, Jr., and S. C. Porter, Eds.), pp. 53–70. Univ. of Minnesota Press, Minneapolis.

Weertman, J. (1964). Rate of growth or shrinkage of non-equilibrium ice sheets. *Journal of Glaciology* 5, 145–158.

Wilson, J. T., Falconer, G., Mathews, W. H., and Prest, V. K. (compilers) (1958). "Glacial Map of Canada." Geological Association of Canada.