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 limit 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 nonsurfing behavior. Average meltwater discharge increases monotonically downglacier to 3000 m³/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
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]:

\[ \bar{u} = 2h \Lambda (\tau_b)^{n/(n + 2)}, \]  

(1)

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

\[ \tau_b = \rho gh \sin \alpha, \]  

(2)

where \( \rho \) is the ice density, \( g \) the acceleration due to gravity, and \( \alpha \) 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 \( \tau_b \) 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 $\tau_b = 50$ kPa and $h = 1500$ m, and flow parameters $n = 3$ and $A = 5.3 \times 10^{-15} \text{ sec}^{-1} \text{ kPa}^{-3}$ (Paterson, 1981, p. 39), Equations (1) and (2) give

$$\bar{u} = 11 \text{ m/a}$$

(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-
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

![Figure 3](image)

**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), $7.9 \times 10^{10}$ 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 $1.2 \times 10^8$ 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.

**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>
<thead>
<tr>
<th>Altitude interval (m)</th>
<th>Area of Juan de Fuca lobe ($10^3$ km$^2$)</th>
<th>Area of Puget lobe ($10^3$ km$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0–300</td>
<td>10.1</td>
<td>0.8</td>
</tr>
<tr>
<td>300–600</td>
<td>3.2</td>
<td>3.8</td>
</tr>
<tr>
<td>600–900</td>
<td>4.3</td>
<td>6.0</td>
</tr>
<tr>
<td>900–1200</td>
<td>6.5</td>
<td>7.8</td>
</tr>
<tr>
<td>1200–1500</td>
<td>23.9</td>
<td>8.6</td>
</tr>
<tr>
<td>&gt;1500</td>
<td>27.7</td>
<td>12.4</td>
</tr>
</tbody>
</table>
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

<table>
<thead>
<tr>
<th>Contour interval (m)</th>
<th>Ice width (10$^5$ m)</th>
<th>Average thickness (10$^3$ m)</th>
<th>Cross-sectional area (10$^4$ m$^2$)</th>
<th>Ice discharge (10$^6$ m$^3$/a of ice)</th>
<th>Ice velocity (m/a)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2000</td>
<td>1.1</td>
<td>1.0</td>
<td>1.1</td>
<td>4.8</td>
<td>430</td>
</tr>
<tr>
<td>1500</td>
<td>1.1</td>
<td>1.1</td>
<td>1.2</td>
<td>7.0</td>
<td>580</td>
</tr>
<tr>
<td>1200</td>
<td>1.0</td>
<td>1.2</td>
<td>1.2</td>
<td>7.9</td>
<td>660</td>
</tr>
<tr>
<td>900</td>
<td>1.3</td>
<td>0.9</td>
<td>1.2</td>
<td>6.4</td>
<td>540</td>
</tr>
<tr>
<td>600</td>
<td>1.1</td>
<td>0.7</td>
<td>0.8</td>
<td>3.8</td>
<td>470</td>
</tr>
<tr>
<td>300</td>
<td>1.0</td>
<td>0.4</td>
<td>0.4</td>
<td>1.2</td>
<td>310</td>
</tr>
<tr>
<td>Terminus</td>
<td>1.0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

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

<table>
<thead>
<tr>
<th>Contour (m)</th>
<th>Ice discharge (10^10 m³/a of ice)</th>
<th>Cumulative water discharge (10^10 m³/a)</th>
<th>Water discharge/ice discharge (by volume)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2000</td>
<td>4.8</td>
<td>0.0</td>
<td></td>
</tr>
<tr>
<td>1500</td>
<td>7.0</td>
<td>0.4</td>
<td>1/17</td>
</tr>
<tr>
<td>1200</td>
<td>7.9</td>
<td>1.4</td>
<td>1/6</td>
</tr>
<tr>
<td>900</td>
<td>6.4</td>
<td>3.5</td>
<td>1/2</td>
</tr>
<tr>
<td>600</td>
<td>3.8</td>
<td>6.1</td>
<td>1.6/1</td>
</tr>
<tr>
<td>300</td>
<td>1.2</td>
<td>8.6</td>
<td>7/1</td>
</tr>
<tr>
<td>Terminus</td>
<td>0</td>
<td>9.7</td>
<td>x</td>
</tr>
</tbody>
</table>

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, analogous 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, 14C sample W-1305) and 13,650 ± 550 yr B.P. (Rigg and Gould, 1957, 14C 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 km² (Table 1) divided into the predicted yearly water discharge of 97 km³/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 topo-
graphy and exposure strongly influence single measurements far beyond the effects of altitude and increasing continentality, which for ocean-facing slopes have op-
posing 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 ac-
cumulation rate can be set, suggested by some observational and theoretical evi-
dence 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 sug-
gests a lower limit for the accumulation rate far upglacier.

**Sliding Velocity**

Calculated sliding velocities are mostly insensitive to any plausible range in recon-
structed 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 plau-
sible ELAs (1200–1500 m) is constrained
by geologic evidence (see above) inde-
pendent of any assumed mass-balance rela-
tionship. For an ELA = 1200 m, the flux
through the equilibrium line is $7.3 \times 10^{10}$
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 $10.8 \times
10^{10}$ m$^3$/a, giving a velocity at the equilib-
rium line of 890 m/a (a 40% increase). Even

<table>
<thead>
<tr>
<th>ELA (m)</th>
<th>Ice velocity due to internal deformation (m/a)</th>
<th>% of total flow due to basal slip</th>
</tr>
</thead>
<tbody>
<tr>
<td>1125</td>
<td>11</td>
<td>98</td>
</tr>
<tr>
<td>1200</td>
<td>11</td>
<td>98</td>
</tr>
<tr>
<td>1225</td>
<td>11</td>
<td>98</td>
</tr>
<tr>
<td>1500</td>
<td>13</td>
<td>99</td>
</tr>
</tbody>
</table>

using an ELA of 1125 m, required by a 2
m/a limit on accumulation rates, the ve-
locity 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 regard-
less 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 $5 \times 10^9$ m$^3$ (cf. Table 2) would be equiva-
lent 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.

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 \( \times \) 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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