Géographie physique et Quaternaire

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La physique glaciaire du lobe de Puget, au sud-ouest de l'Inlandsis de la Cordillère
Glaziale Physik des Puget-Lappens, südwestliche Kordilleren-Eisdecke

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Volume 45, numéro 3, 1991

L'Inlandis de la Cordillère
The Cordilleran Ice Sheet

URI : https://id.erudit.org/iderudit/032877ar
DOI : https://doi.org/10.7202/032877ar

Aller au sommaire du numéro

Éditeur(s)

Les Presses de l'Université de Montréal

ISSN

0705-7199 (imprimé)
1492-143X (numérique)

Découvrir la revue

Citer cet article

Booth, D. B. (1991). Glacier Physics of the Puget Lobe, Southwest Cordilleran Ice Sheet. *Géographie physique et Quaternaire*, *45*(3), 301–315. https://doi.org/10.7202/032877ar

Résumé de l'article

Le lobe de Puget, à la limite sud-ouest de l'Inlandsis de la Cordillère, permet d'étudier les liens entre la physique glaciaire et l'action glaciaire. L'action de l'eau à proximité et à l'intérieur des sédiments du lit glaciaire est un aspect particulièrement important de la géologie de l'inlandsis. Les données d'ordre physique et les calculs du bilan de masse permettent de faire une reconstitution assez fidèle du lobe et d'extrapoler les vitesses de glissement au delà de 500 m/a et les débits d'eau de près de 1 * 10" m3/a. L'écoulement de l'eau sous-glaciaire a créé un réseau de chenaux dentritiques que l'analyse statique de l'hydrologie sous-glaciaire a extrapolé avec justesse. Près de la marge glaciaire orientale, un important chenal isolé a assuré épisodiquement le drainage sous-glaciaire, alors que les lacs de barrage glaciaire emprisonnés dans les vallées adjacentes se déchargeaient par débâcles. L'eau de fonte basale a produit des pressions d'eau près de l'équilibre hydrostatique et de très faibles résistances du till à la base de l'inlandsis. La pression de l'eau ne diminuait que près de la marge glaciaire, laissant ainsi les taux de contrainte normaux s'élever jusqu'à des proportions importantes de la couverture totale de glace. Ainsi les conditions du lit sont très différentes selon qu'il se situe en zone interne ou en zone marginale. L'observation des dépôts sous-glaciaires révèle la phase finale en zone marginale, mais les conditions à l'intérieur même de l'inlandsis, d'une plus grande portée quant à l'évolution et au comportement du glacier, pourraient être très différentes.

Géographie physique et Quaternaire, 1991, vol. 45, n° 3, p. 301-315, 13 fig., 1 tabl.

GLACIER PHYSICS OF THE PUGET LOBE, SOUTHWEST CORDILLERAN ICE SHEET

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ABSTRACT The Puget lobe, the southwestmost extension of the Cordilleran ice sheet, provides an excellent opportunity to examine the connection between glacier physics and the resulting products of glaciation. The action of water, at and within the sediments of the glacier bed, is particularly significant for the geologic record of this ice sheet. Physical data and inferred mass balance relationships constrain lobe reconstruction and predict sliding velocities in excess of 500 m/a and water discharges of nearly 1×10^{11} m³/a. This subglacial water produced a dendritic channel pattern well predicted by static analysis of subglacial hydrology. Near to the eastern ice margin, a much larger single channel drained subglacially and episodically, with tributary ice-dammed lakes releasing their water as jokulhlaups. Basal meltwater generated near-hydrostatic water pressures and very low till strengths at the base of the ice sheet. Water pressure dropped only close to the ice margin, allowing normal stresses to rise to significant fractions of the total ice overburden. Thus marginal and interior zones impose contrasting bed conditions. Although observation of subglacial deposits will reflect the late-stage passage of the marginal zone, conditions within the ice-sheet interior, far more significant to glacier history and behavior, may be substantially different.

RÉSUMÉ *La physique glaciaire du lobe de Puget, au sud-ouest de l'Inlandsis de la Cordillère*. Le lobe de Puget, à la limite sud-ouest de l'Inlandsis de la Cordillère, permet d'étudier les liens entre la physique glaciaire et l'action glaciaire. L'action de l'eau à proximité et à l'intérieur des sédiments du lit glaciaire est un aspect particulièrement important de la géologie de l'inlandsis. Les données d'ordre physique et les calculs du bilan de masse permettent de faire une reconstitution assez fidèle du lobe et d'extrapoler les vitesses de glissement au delà de 500 m/a et les débits d'eau de près de 1×10^{11} m³/a. L'écoulement de l'eau sous-glaciaire a créé un réseau de chenaux dentritiques que l'analyse statique de l'hydrologie sous-glaciaire a extrapolé avec justesse. Près de la marge glaciaire orientale, un important chenal isolé a assuré épisodiquement le drainage sous-glaciaire, alors que les lacs de barrage glaciaire emprisonnés dans les vallées adjacentes se déchargeaient par débâcles. L'eau de fonte basale a produit des pressions d'eau près de l'équilibre hydrostatique et de très faibles résistances du till à la base de l'inlandsis. La pression de l'eau ne diminuait que près de la marge glaciaire, laissant ainsi les taux de contrainte normaux s'élever jusqu'à des proportions importantes de la couverture totale de glace. Ainsi les conditions du lit sont très différentes selon qu'il se situe en zone interne ou en zone marginale. L'observation des dépôts sous-glaciaires révèle la phase finale en zone marginale, mais les conditions à l'intérieur même de l'inlandsis, d'une plus grande portée quant à l'évolution et au comportement du glacier, pourraient être très différentes.

ZUSAMMENFASSUNG *Glaziale Physik des Puget-Lappens, südwestliche Kordilleren-Eisdecke*. Der Puget-Lappen, die südwestliche Ausdehnung der Kordilleren-Eisdecke, bietet eine ausgezeichnete Gelegenheit, die Beziehung zwischen glazialer Physik und den daraus resultierenden Produkten der Vereisung zu untersuchen. Die Wirkung von Wasser in der Nähe der Sedimente des Gletscherbettes und in ihnen ist besonders signifikant für die geologische Dokumentierung der Eisdecke. Physikalische Daten und abgeleitete Massenbilanz-Beziehungen erlauben eine getreue Rekonstruktion des Lappens und Voraussagen über Rutschgeschwindigkeiten von über 500 m/a und Wasserführungen von nahezu 1×10^{11} m³/a zu machen. Dies subglaziale Wasser bewirkte ein baumartiges Kanal-Muster, welches die statische Analyse der subglazialen Hydrologie richtig vorausgesagt hatte. In der Nähe des östlichen Eisrands bewirkte ein sehr viel breiterer einzelner Kanal gelegentlich eine subglaziale Dränage, und tributäre, durch Eis eingeschlossene Seen entließen ihr Wasser in Gletscherläufen. Basales Schmelzwasser bewirkte fast hydrostatische Wasserdruckwerte und sehr niedrige Widerstände des Tills an der Basis der Eisdecke. Der Wasserdruck sank nur in der Nähe des Eisrands, so daß normaler Druck bis zu bedeutenden Proportionen der totalen Eislast anstieg. So sind die bedingungen des Bettes sehr verschieden, je nachdem ob es sich in der Randzone oder der inneren Zone befindet. Obgleich die Beobachtung der subglazialen Ablagerungen die Endphase der Randzone spiegelt, könnten doch die Bedingungen im Innern der Eisdecke, die für die Geschichte und das Verhalten des Gletschers viel wichtiger sind, sehr verschieden sein.

Manuscrit reçu le 7 février 1991; manuscrit révisé accepté le 20 septembre 1991

INTRODUCTION

The Puget lobe of the Cordilleran Ice Sheet formed the southwest-most extention of ice during the last glaciation of North America (Fig. 1). Beginning as coalescing mountain ice caps in British Columbia, the lobe advanced first as a tongue of ice along the Georgia Depression and then extended farther south along the trough of the Puget Lowland in western Washington state. At ice maximum, the lobe was about 100 km wide; its lateral extent was largely confined by the Olympic Mountains to the west and the Cascade Range to the east. In contrast, its southernmost limit reflected more the balance of ice-sheet accumulation and ablation than any significant topographic barriers.

Over a century of study has developed a picture of ice-sheet chronology, erosion, and deposition. Yet many of the basic questions of ice-sheet behavior, particularly how the landforms of the region have been shaped, are addressed only imperfectly by such studies. I have sought to approach some of these questions, particularly those pertaining to subglacial water flow and the properties of subglacial sediment, by using the rapidly increasing understanding of glacier physics. The results of similar efforts from other geographic regions also are relevant, because the underlying mechanical and material laws are universal.

For a number of reasons, the Puget lobe provides an exceptional opportunity to examine the connection between glacier physics and the geomorphic products of the glacier system. Most importantly, the Puget lobe forms a well-constrained system, with good age control, clearly recognized boundaries, and moderately definitive source area. The geomorphic products of glaciation are well-displayed across the Puget Lowland, despite the region's luxuriant vegetation cover. Rapid urbanization of this region has provided good access, widespread exposures, and relatively dense subsurface data. Finally, a number of ancillary studies of late Pleistocene environments in the region, particularly those reflecting climatic variation from local glaciers (Porter, 1977), vegetation (Barnosky *et al.*, 1987), and global climate (Kutsbach, 1987), establish an independent environmental context and provide invaluable checks on the results of ice-sheet reconstruction and modeling.

The following discussion of the physics and the physical processes of the Puget lobe draws heavily on recent studies, not only those of the Puget lobe itself (*e.g.*, Booth, 1986a,b; Brown *et al.*, 1987; Thorson, 1989) but also those of other regions using compatible analyses (*e.g.*, Boulton and Paul, 1976; Pierce, 1979; Shreve, 1985; Ridky and Bindschadler, 1990). In this region I have focused on the subglacial water system, because the movement of water at and within the glacier appears to have controlled the most significant aspects of the Puget lobe and its geomorphic effects. This significance is probably not unique to the Puget lobe; instead, it may be pervasive wherever large temperate glaciers sustained high discharges of both water and ice (Gustavson and Boothroyd, 1987).

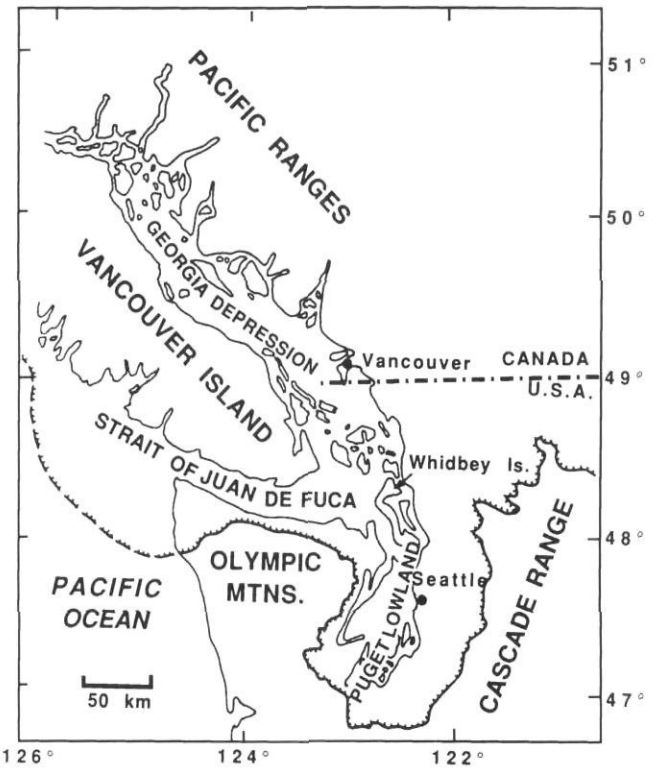


FIGURE 1. The Puget Lobe of the Cordilleran Ice Sheet, showing maximum extent (*ca.* 15 ka BP).

Étendue maximale du lobe de Puget de l'Inlandsis de la Cordillère, vers 15 ka BP.

REGIONAL SETTING

GEOGRAPHY

The southwest part of the Cordilleran Ice Sheet occupied a distinctive geographic region. A broad topographic basin in British Columbia, the Georgia Depression and the Fraser Lowland, extends southward into Washington state. The depression splits at the northeast corner of the Olympic Peninsula. One branch extends west, between the Olympic Mountains and Vancouver Island, as the Strait of Juan de Fuca. The other branch continues south, forming the Puget Lowland between the Olympic Mountains and Cascade Range. Although low hills at about 46°45′ define the southern limit of ice advance in the Puget Lowland, the lowland province itself continues south for at least several hundred kilometers.

The southern Cordilleran Ice Sheet has expanded into the Puget Lowland during several glaciations during the Pleistocene (Crandell *et al.*, 1958; Easterbrook *et al.*, 1967). The most recent advance, the Vashon stade of the Fraser glaciation of Armstrong *et al.* (1965), provides the best picture of ice-sheet growth and decay. Ice caps on the mountains of Vancouver Island and the British Columbia mainland expanded and coalesced, gradually extending into the lowland valleys and the Georgia Depression (Clague, 1981). As the ice tongue moved into the Puget Lowland, it received no additional lateral input — mountain glaciers in the Olympics and Cascades had

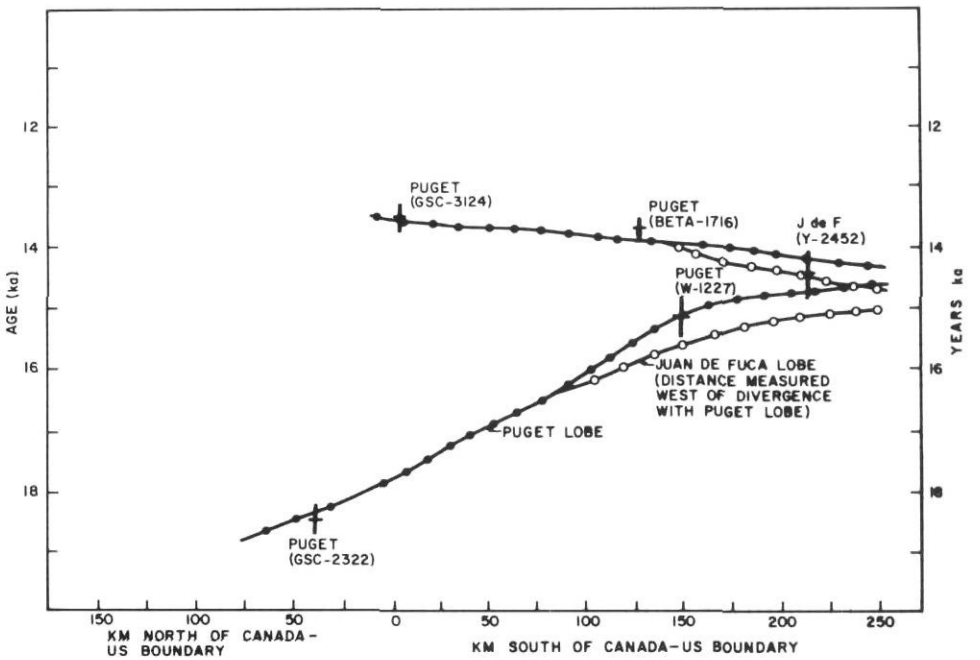


FIGURE 2. Advance and retreat of the Puget and Juan de Fuca lobes, with key limiting radiocarbon dates and standard errors (adapted from Booth, 1987).

Avancée et retrait des lobes de Puget et de Juan de Fuca ainsi que les principales dates au radiocarbone et leurs marges d'erreur (adapté de Booth, 1987).

already retreated from their late Pleistocene maximum limits and probably did not contribute any additional ice to the Puget lobe (Porter, 1976; Booth, 1987).

CHRONOLOGY

During the Fraser Glaciation, the advancing and retreating Puget lobe was among the more rapid-moving of the North American ice sheets. Only a few thousand years span the advance of the lobe across the Canadian border, attainment of maximum southern position, and retreat back to the foothills of the British Columbia mountains. A movement of the terminus of over 200 km in each direction was therefore accomplished in this period, for an average velocity at the terminus of at least 100 m/a. Yet the geomorphic record in the Puget Lowland is primarily that of an ice sheet at maximum stage (see also Sugden, 1979), with relatively well-defined ice limits and ice-flow patterns that are consistent with a sustained ice-maximum position (Thorson, 1980; Booth, 1984b). Thus the lobe undoubtedly maintained its maximum position for at least some fraction of its total history, with limiting radiocarbon dates (see below) permitting 500-1000 years of maximum or near-maximum conditions. Advance and retreat rates were therefore even higher than their minimum, averaged value of 100 m/a.

Abundant radiocarbon dating constrains both the advance and the retreat chronology of the Puget lobe and the neighboring Juan de Fuca lobe (Fig. 2). The ice sheet entered the Fraser Lowland some time after 18.3 ± 0.17 ka (GSC-2322; Armstrong and Clague, 1977, Fig. 2). The advance of the Puget lobe is further constrained in the Seattle area, 160 km south of the International Boundary, by a pre-glacial age of 15.0 ± 0.4 ka (W-1227; Mullineaux *et al.*, 1965).

Final advance to the ice-maximum position and subsequent retreat are also detailed for both lobes. The Juan de Fuca lobe has one dated locality near its terminus, a post-retreat date of 14.46 ± 0.20 ka (Y-2452; Heusser, 1973) 50 km upglacier of its ice-maximum limit. The Puget lobe advanced an additional 100 km south of Seattle and then retreated a like amount between 15.0 ka (see above) and 13.65 ± 0.55 ka (L-346a; Rigg and Gould, 1957). Yet nearly equivalent dates of 13.60 ± 0.15 ka and 13.65 ± 0.35 ka are also reported from shells on Whidbey Island, an additional 30 to 50 km north of Seattle (BETA-1716 and BETA-1319; D. P. Dethier *et al.*, written communication, 1986). Grounded ice of both the Puget and Juan de Fuca lobes therefore must have retreated at least this far upglacier by that time. Required minimum rates of terminal advance and retreat are about 200 m/a during this period; allowing for slower advance rates (Weertman, 1964) and sufficient time to achieve equilibrium at ice-maximum position, actual retreat rates were probably closer to 500 m/a.

The character of the final retreat and decay of the combined Puget and Juan de Fuca lobes is suggested by both radiocarbon dates and the types of recessional deposits. In contrast to the southern Puget Lowland, where dead-ice topography is relatively rare and active ice margins systematically diverted glacial meltwater at the ice margin (Thorson, 1980; Booth, 1987, 1990), the northern lowland and eastern Strait of Juan de Fuca display either irregular recessional deposits (Anderson, 1968; Chrzastowski, 1980), glaciomarine drift (Easterbrook, 1963; Pessl *et al.*, 1987), or no deposits at all. This irregularity of deposition is matched by a broad suite of dates in this region, spanning in time from 13.65 ± 0.35 ka (BETA-1319) to 11.30 ± 0.07 ka (USGS-124; Dethier *et al.*, written communication, 1986) and in space from Whidbey Island north to the International Boundary. In this region, the ice probably retreated primarily by calving, with local backwasting and stagnation (Pessl *et al.*, 1987). Marine sedimentation would have continued until isostatic rebound finally lifted much of the area above sea level (Thorson, 1989).

RECONSTRUCTION OF THE PUGET LOBE

Determining the physical behavior of an ice sheet first requires some knowledge of its physical dimensions. The flux of both ice and water must be estimated as well, because the passage of these agents over the glacier bed is what yields geomorphic changes. Unfortunately, most glacier systems are ill-suited to such a reconstruction. If applied to an existing glacier, the necessary parameters are easily measured but their effects on the glacier bed are obscured. Pleistocene ice sheets suffer from the inverse problem — typically, the record of their passage is grossly incomplete or their mass-balance regime is completely unknown. The Puget lobe does not avoid completely these shortcomings of other vanished ice sheets, but sufficient data are available and constrained by independent checks that reconstruction is both feasible and instructive. The discussion of method and results that follow are summarized from Booth (1986a), which includes a more extensive review of techniques, data sources, errors, and independent verifications.

METHODS

This reconstruction is based first on the compilation of available geologic data on the physical extent of the ice sheet (Fig. 3) and second on the calculation of an equilibrium mass balance for the Puget lobe itself. The southern boundary of the ice sheet, along the boundary of the Puget lobe, is rather well-constrained; the southwestern boundary was formed by the tidewater part of the Juan de Fuca lobe and so is far less certain. Ice-flow direction indicators discriminate the source area of the two lobes in southern and central British Columbia; the altitude of marginal ice limits and nunataks provide most of the data for ice-surface contours. This compilation relies heavily on Thorson (1980) in the south and Wilson *et al.* (1958) in the north. I have sought to reconcile these and other subsequent reconstructions using several basic assumptions: ice-surface contours lie perpendicular to flow indicators, flow lines do not converge or diverge without commensurate changes in ice thickness or net mass balance, and longitudinal stress gradients (*i.e.* the downglacier change in the depth-slope product) are low at the scale of the reconstruction.

The mass balance of an ice sheet during the Pleistocene cannot be known with certainty, but analogy with existing glaciers provides a reasonable working estimate. In the Pacific Northwest, long-term study of a number of modern maritime glaciers yields a local mass-balance relationship (Meier *et al.*, 1971; Porter *et al.*, 1983), expressing net accumulation or ablation as a function of elevation above or below the equilibrium line altitude, or ELA (Fig. 4). The *total* ablation provides a measure of the water production that discharges through the glacier, irrespective of whether that melting is compensated by snow accumulation or not. This relationship is compiled from high-quality data from two Norwegian maritime glaciers (IAHS, 1977).

RESULTS

With a reconstructed ice sheet and a relationship between relative altitude and mass balance, an ELA can be determined by trial and error that brings the lobe into balance. Conversely, if the ELA is known independently, the condition of net growth or decay can be easily calculated. In either case, the flux of both ice and water can be established as a function of position downglacier.

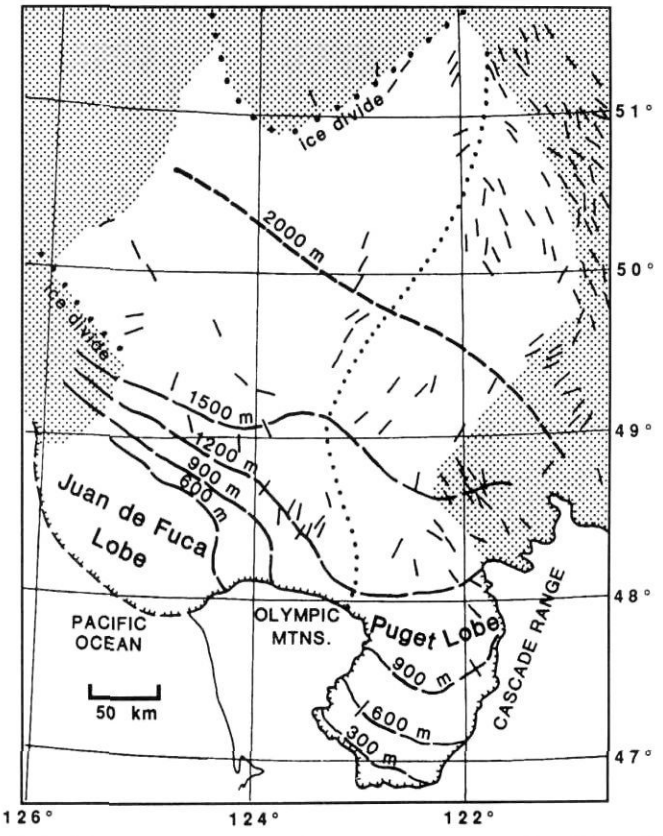


FIGURE 3. Reconstruction of the Puget and Juan de Fuca lobes of the Cordilleran Ice Sheet at maximum stage. Short lines show orientation of representative striations and other glacial lineations; hatchured line shows maximum extent of ice. Heavy lines are contours of ice-surface altitudes above modern sea level, uncorrected for glacial-age sea level lowering or isostatic depression. Unshaded ice-covered area included all ice of the Puget and Juan de Fuca lobes, with an inferred separation between the lobes indicated by the dotted line (adapted from Booth, 1986a).

Reconstitution des lobes de Juan de Fuca et Puget de l'Inlandsis de la Cordillère à l'optimum glaciaire. Les petits traits donnent l'orientation des stries et autres marques glaciaires représentatives: la ligne à barbules montre l'étendue maximale des glaces. Les lignes discontinues montrent l'altitude des surfaces glaciaires au-dessus du niveau actuel de la mer, non corrigée en fonction de l'abaissement isostatique de l'époque glaciaire. La région glaciaire non tramée comprend l'ensemble des deux lobes; une séparation hypothétique entre les deux lobes est indiquée par le pointillé (à partir de Booth, 1986 a).

Using these procedures, the Puget lobe is calculated to be in balance with an ELA between 1200 and 1250 m. Table I shows the resulting values of the ice and water fluxes for an ELA between these two levels. The contribution to ice velocity from internal ice deformation can be calculated (Paterson, 1981) and accounts for less than 2 percent of the total flux. Thus basal sliding must account for virtually all of the predicted motion, which is several hundred meters per year over nearly

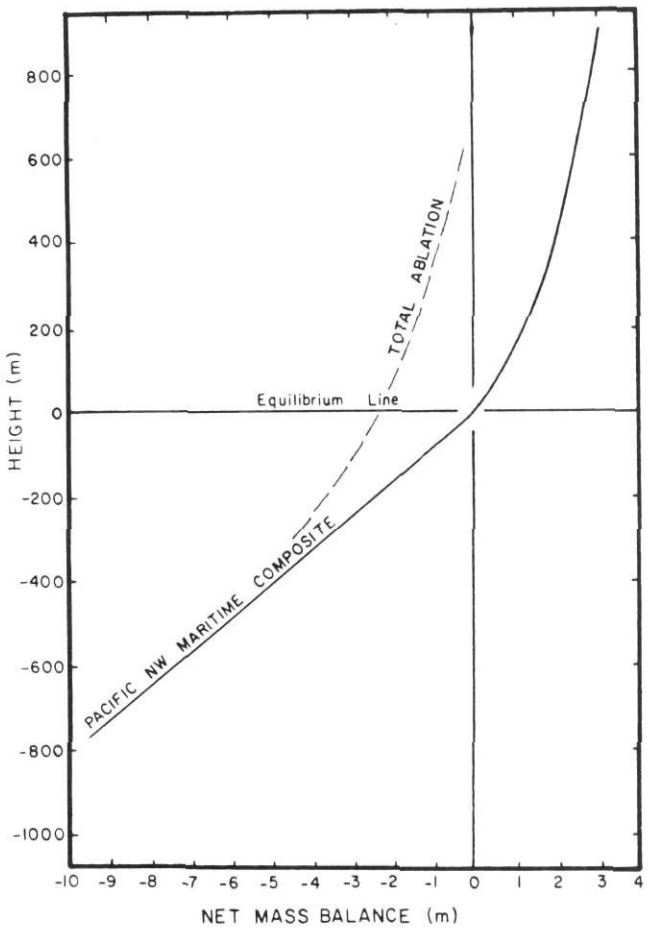


FIGURE 4. *Height-mass balance relationships for modern maritime* glaciers. Height is referenced to the equilibrium-line altitude; specific balance is in cubic meters of equivalent water volume per square meter of ice surface per year (adapted from Booth, 1986a, with original data from Meier *et al.*, 1971, and IAHS, 1977).

Relations entre la hauteur et le bilan de masse pour des glaciers maritimes modernes. La hauteur est en fonction de l'altitude de la ligne d'équilibre; le bilan spécifique est en mètres cubes de volume d'eau équivalent par mètre carré de surface de glace par année (à partir de Booth, 1986a et des données originales de Meier et al., *1971 et IAHS, 1977).*

TABLE I

Ice and water discharge through transverse sections of the Puget Lobe, ELA = 1225 m (adapted from Booth, 1986a)

Contour Interval (m)	Cross-Sectional Area (10^8 m)	Water Discharge (10^{10} m³/a)	Ice Discharge (10^{10} m³/a)	Ice Velocity (m/a)
2000	1.1	0.0	4.8	430
1500	1.2	0.4	7.0	580
1200	1.2	1.4	7.9	660
900	0.9	3.5	6.4	540
600	0.7	6.1	3.8	470
300	0.4	8.6	1.2	310
Terminus	0	9.7	0	0

the entire source and ablation areas of the lobe. The ice flux peaks at the ELA, in contrast to the monotonic increase in water flow downglacier (Fig. 5).

Nearly every element of this reconstruction is constrained by independent data; the net result, that of basal sliding of hundreds of meters per year, is unaffected by the uncertainties inherent in such an analysis. The mass-balance relationships are least verifiable, but the consequences of reduced accumulation-area snowfall or terminal ablation can be evaluated (Booth, 1986a). In both cases, any significant change in the mass-balance relationship raises or lowers the location of the ELA to geologically implausible levels; namely, moving the ELA onto areas of the lobe where flow lines either converge or diverge strongly.

ICE-SHEET RECONSTRUCTION AND GLACIAL PROCESSES

The effects of glaciation depend on certain key elements of the ice-sheet reconstruction. From lobe dimensions, the basal shear stress of the ice is readily calculated and ranges between 40-50 kPa (Booth, 1986a; Brown *et al.*, 1987). Despite this shear stress, low by the standards of modern valley glaciers but apparently quite typical of ice streams and Pleistocene ice lobes (Blankenship *et al.*, 1987; Mathews, 1974; Paterson, 1981), sliding velocities were many hundreds of meters per year. Ablation generated much meltwater, which approached and finally exceeded the flux of ice downglacier in the ablation zone. The system thus described is one of rapid mass transport under a rather low driving stress, traversing a bed of mainly unconsolidated sediment.

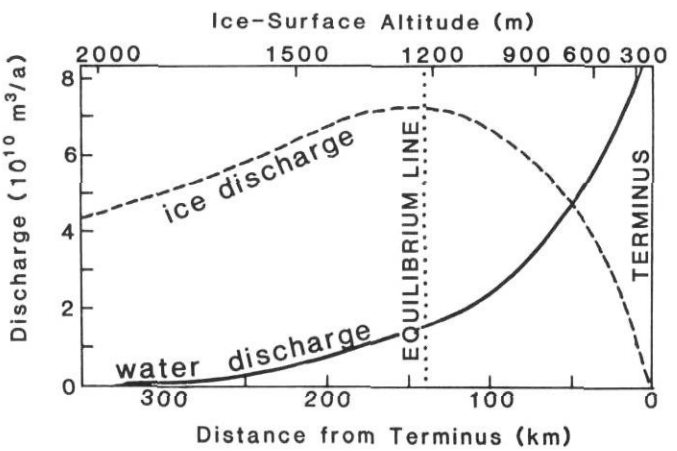


FIGURE 5. Variation of ice and water fluxes along the length of the Puget lobe, using an ELA of 1225 m.

Variations des débits de glace et d'eau tout au long du lobe de Puget, selon une ligne d'équilibre à 1225 m d'altitude.

THE MOVEMENT OF SUBGLACIAL WATER

Theory

Water flow through a glacier can be described by the same mathematical formulations developed for groundwater. Water moves from areas of high total potential, or total head, to areas of low total head. Flow is down the gradient of the potential field, analogous to the flow of surface water over the topography of a hillslope (Shreve, 1972; Rothlisberger, 1972).

Total head at a point has two significant static components: the elevation head, measured as the height of a point in the water above a chosen datum; and the pressure head, which for a steady-state water-filled tunnel is nearly equal to the overburden pressure due to the overlying ice (Shreve, 1972).

The total head (H) is expressed in units of length and is the sum of the elevation head (z) and the pressure head. Because:

$$p_i = \rho_i g h,$$

the total head is:

$$H = z + \left(\frac{\rho_i}{\rho_w}\right) h, \qquad (1)$$

where p_i = the ice overburden pressure,
ρ_w = water density,
ρ_i = ice density,
g = gravitational acceleration, and
h = the overlying thickness of ice.

Because $z + h$ is constant within any vertical column in the ice and $\rho_i < \rho_w$, total head will be lowest at the bed of the glacier (where $z = z_b$), and thus subglacial flow is favored over englacial flow (Shreve, 1972).

The gross pattern of water flow beneath the glacier is readily predicted from equation (1). Differentiation of that equation yields:

$$\frac{\partial H}{\partial x} = \frac{\partial z_b}{\partial x} + \left(\frac{\rho_i}{\rho_w}\right) \times \frac{\partial (z_s - z_b)}{\partial x}$$

where x = distance downglacier, and
z_s = the ice-surface elevation, $z_b + h$.

If two points on the glacier bed have the *same* total head,

$$\frac{\partial H}{\partial x} = 0$$

and so:

$$\frac{\partial z_s}{\partial x} = -0.09 \left(\frac{\partial z_b}{\partial x}\right)$$

So long as the bed does not slope upglacier more steeply than eleven times the downglacier ice slope, total head will decrease downglacier and water will follow the ice slope direction, flowing up or down the bed topography with only minor deflection (Shreve, 1972, 1985).

Reconstruction

Using this static analysis of subglacial water flow, the potential field at the bed of an ice sheet can be developed. The approach used here is conceptually identical to that followed by Shreve (1985) for a portion of the Laurentide Ice Sheet in the state of Maine. It requires two reconstructed elements: the topography of the bed and the topography of the ice-sheet surface. The reconstructions are detailed for a part of the eastern Puget lobe (Fig. 6), using digitized topography from U.S. Geological Survey 1:24,000 and 1:62,500 topographic maps (Fig. 7) and a more detailed version of the ice-sheet reconstruction shown in Figure 3 (Fig. 8). No correction is made for postglacial modification of the bed topography, undulations in the ice surface from bed irregularities (Robin, 1967), dynamic pressures imposed by the flowing ice, or isostatic rebound following deglaciation, as the overall results are insensitive to these complications (Booth, 1984b).

The subglacial potential field is calculated with these data and equation (1). The resulting equipotentials, or contours of equal total head, define the conceptual "topography" down

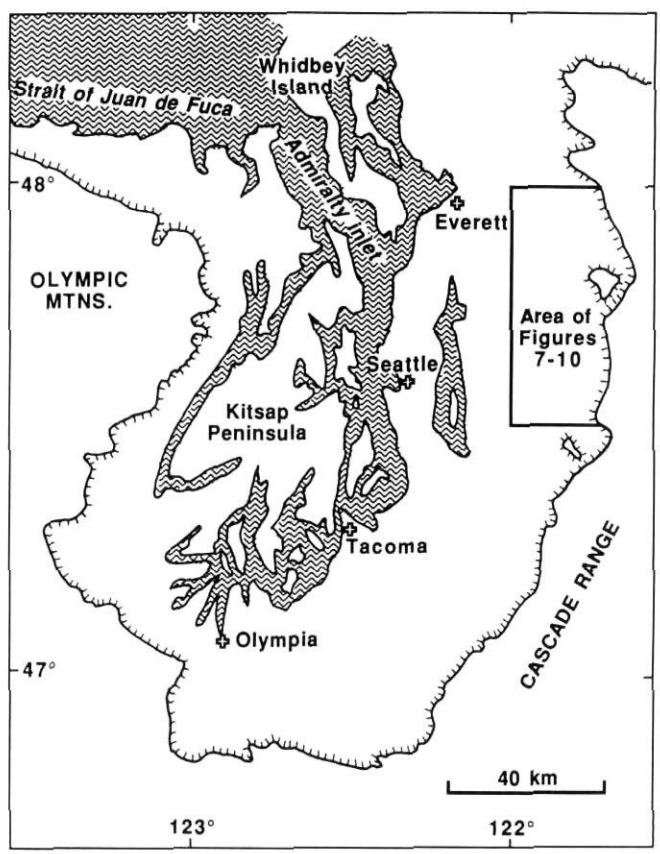


FIGURE 6. Index map of the Puget lobe, showing the area of reconstructed subglacial hydraulic equipotentials along the eastern margin in the Skykomish-Snoqualmie region (Figs. 7-10).

Carte index du lobe de Puget, montrant la région où l'on a fait la reconstitution des zones des équipotentiels hydrauliques sous-glaciaires le long de la marge orientale, dans la région de Skykomish-Snoqualmie (fig. 7 à 10).

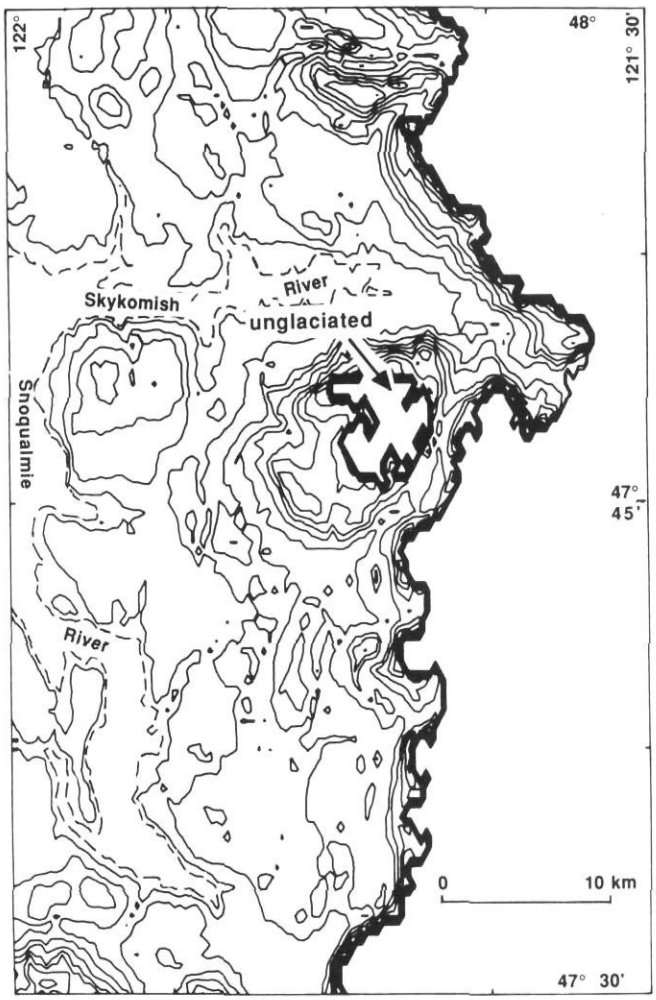


FIGURE 7. Ground topography of the Skykomish-Snoqualmie region outlined in Figure 6. Contour interval 100 m (dashed contour = 50 m), digitized from USGS topographic maps. Dark line at eastern border is the Vashon-age ice limit, approximated at the mouth of each alpine valley.

Topographie de la région de Skykomish-Snoqualmie (voir fig. 6). L'équidistance des courbes est de 100 m (courbes interrompues = 50 m), digitalisées à partir des cartes topographiques du USGS. Le trait gras à la marge orientale représente la limite glaciaire de Vashon, approximative à l'embouchure des vallées alpines.

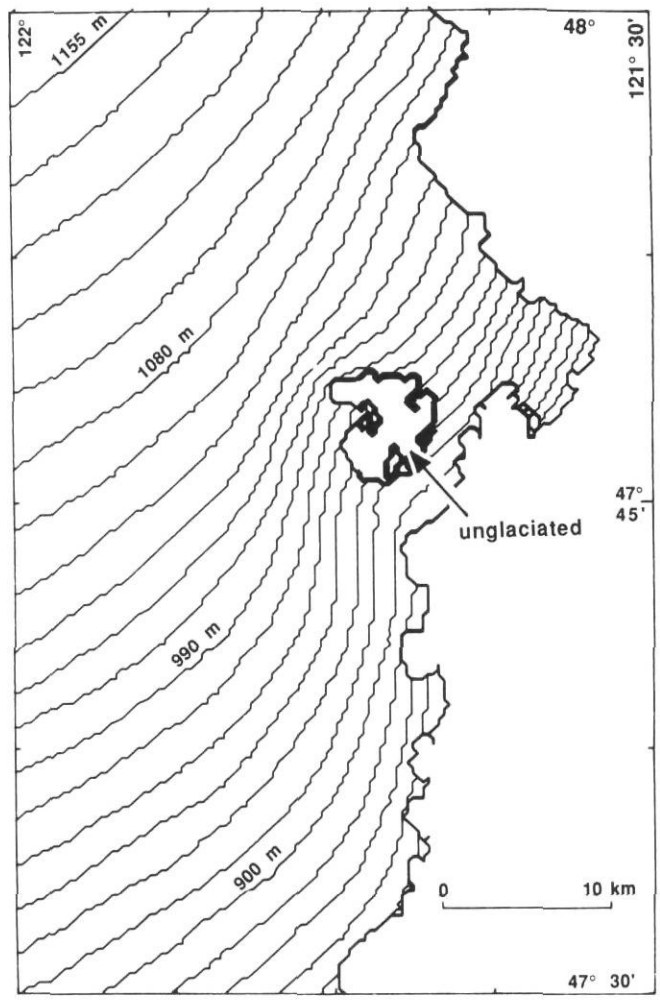


FIGURE 8. Ice-sheet topography of the Puget lobe at maximum stage in the Skykomish-Snoqualmie region (Fig. 6). Contour interval is 15 m.

Topographie du lobe de Puget à son optimum dans la région de Skykomish-Snoqualmie (fig. 6). L'équidistance est de 15 m.

which subglacial water would flow (Fig. 9). The channelized features are highlighted in Figure 10 by following all lows in the potential surface from the upglacier boundaries of the map. The interconnection of these valleys, not always obvious in the modern topography, become apparent when viewed in their proper subglacial context.

The potential field defines two distinct categories of subglacial channels. The first category of channels forms the crudely dendritic network across the region in Figure 10. Water flow follows routes of monotonically decreasing gradient, with divergent flow paths drawn where the resolution of the reconstruction is inadequate to specify a singular route. The net flow is to the south and west, reflecting the overall control of the ice-surface slope in the downglacier direction. Variations in that overall pattern, however, are imposed by the bed topography. In turn, the channelizing of subglacial flow could have enhanced this topography by fluvial erosion.

The second category of channel follows a single submarginal valley near the eastern (lateral) boundary of the reconstructed ice sheet. This channel does not, however, experience monotonically decreasing hydraulic gradients. Instead, local saddles in the potential field block the flow. As with subaerial spillways, water would be impounded behind them; here, that water would have formed subglacial and submarginal lakes in the alpine valleys immediately upglacier of each saddle (Booth, 1986b).

Yet unlike subaerial spillways, the control of lake elevation depended not only on saddle elevation but also on ice pressure. As the lakes filled from subglacial influx and drainage from Cascade Range rivers, the elevation of the lake would have exceeded the potential of the spillway and subglacial drainage would begin. Rapid melting of a tunnel to accommodate this drainage would remove the component of the hydraulic poten-

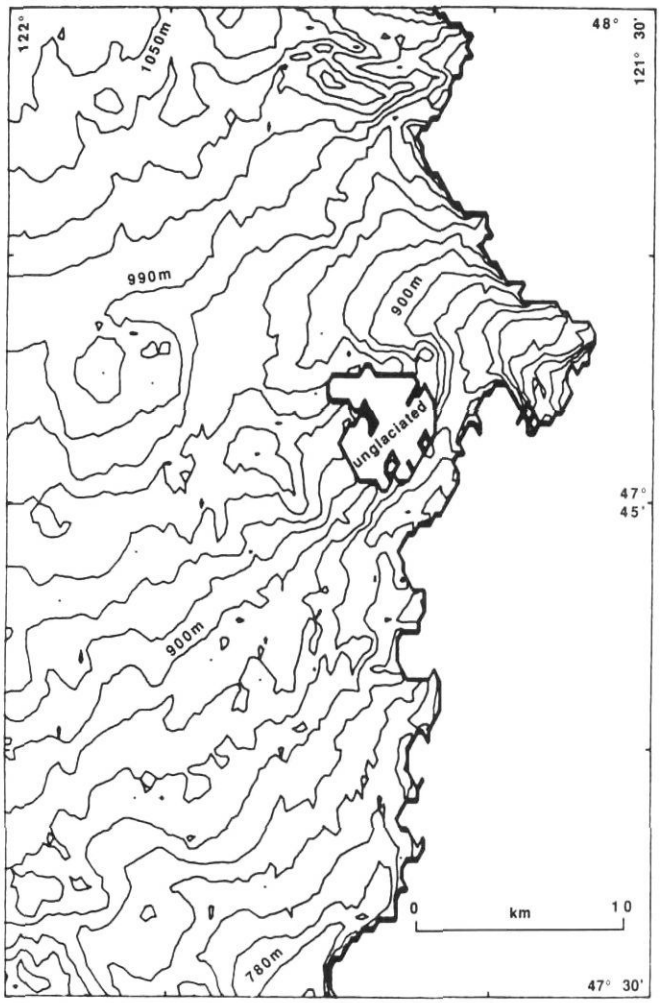


FIGURE 9. Calculated subglacial hydraulic equipotentials in the Skykomish-Snoqualmie region (Fig. 6). Equipotentials are defined from equation (1), such that water of a given potential will stand at that altitude in a borehole or adjacent lake. Contour interval is 15 m; alpine valleys labeled along the eastern ice margin (adapted from Booth, 1986b).

Equipotentiels hydrauliques sous-glaciaires calculés dans la région de Skykomish-Snoqualmie. Les équipotentiels sont définis à partir de l'équation (1) de telle façon que l'eau d'un potentiel donné se maintiendra au même niveau dans un trou de sondage ou un lac adjacent. L'équidistance est de 15 m (adapté de Booth, 1986b).

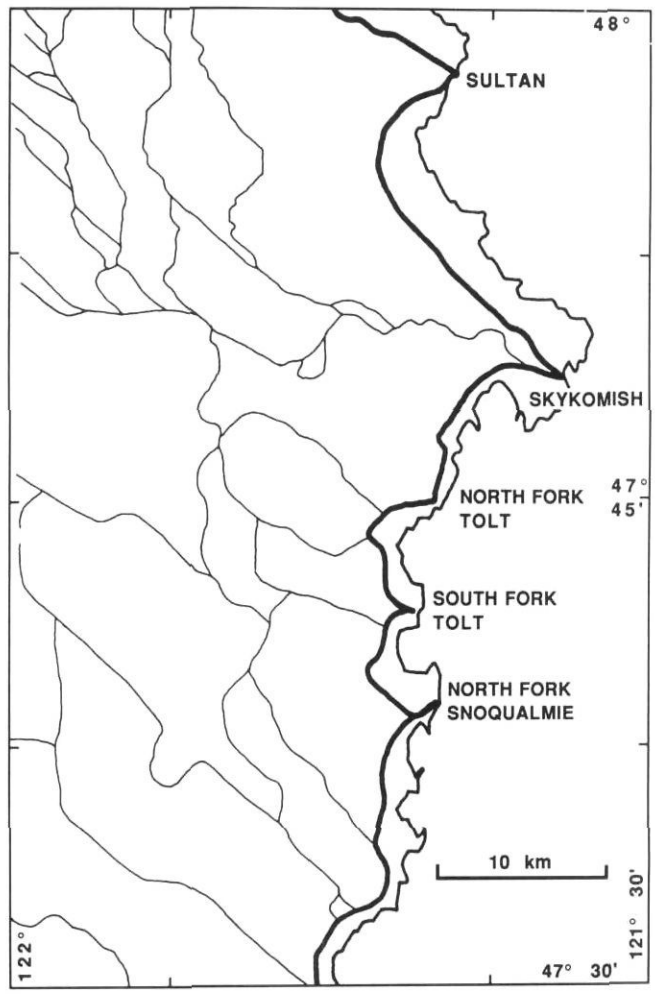


FIGURE 10. Channelized features in the equipotential surface of Figure 9. The thin lines trace a dendritic network of monotonically decreasing hydraulic potential; the thick line marks a single marginal channel where flow was diverted by saddles into adjacent ice-free valleys (named on map), forming ice-marginal lakes that drained episodically.

Chenaux établis à partir de la surface des équipotentiels de la figure 9. Les traits fins identifient un réseau dendritique au potentiel hydraulique décroissant de façon uniforme; le trait gras montre un chenal isolé où l'écoulement était dérivé vers les vallées déglacées adjacentes par des cols formant ainsi des lacs de marge glaciaire qui se vidangeaient épisodiquement.

tial at the spillway due to the ice overburden (Nye, 1976). This radical reduction in potential would precipitate catastrophic discharge of the marginal lake as a jokulhlaup, observed on numerous modern ice-dammed lakes (Stone, 1963; Post and Mayo, 1971). The submarginal flow path of Figure 10 was therefore episodic in nature.

The ice-marginal lakes also lie in sequence — a discharge of one would drain, via the submarginal channel and diversion by the next subglacial spillway downglacier, into the next lake to the south. Drainage of the largest lake (that occupying the modern Skykomish River valley, with an estimated volume of 120 km^3) probably would have triggered all others downglacier, because it contained over five times the combined volume of its downstream neighbors. Its refilling time, based on modern precipitation rates (U.S. Geological Survey, 1983; see Porter, 1977, for a discussion of their applicability) would have been only a few decades (Booth, 1984b, Table 3.1). Thus multiple discharges along the Cascade margin probably were inevitable during the occupation of the Puget lobe.

LANDFORMS OF SUBGLACIAL WATER FLOW

Across the Puget Lowland, the effects of channelized subglacial water are evident. They are particularly prominent towards the lateral margins of the ice sheet's ablation zone. Here, the cumulative effect of upglacier melting increased the total available water, and the convexity of the ice surface drove water into this region, away from the center of the lowland. Erosional features have persisted best in the bedrock uplands,

above the level of recessional lakes and outwash deposits that now fill most of the lower valleys (Bretz, 1913; Thorson, 1980; Booth, 1990).

The erosional channels share a number of characteristics reported elsewhere in regions covered by Pleistocene ice sheets (*e.g.*, Mannerfelt, 1949; Sissons, 1963; Clapperton, 1968; Ehlers; Sharpe and Shaw, 1989). They are sinuous to linear valleys, commonly occupied now by bogs or underfit streams, with steep sidewalls of 10s to over 100 meters high. Small-scale polishing and smoothed sculpted "p-forms" (Dahl, 1965; Sharpe and Shaw, 1989) are sporadically exposed along the bedrock walls. Their modern drainage areas are typically trivial and their gradients are very low, precluding significant postglacial incision. Conversely, their orientations are commonly oblique to independent indicators of ice flow, particularly striae on adjacent uplands, limiting the amount of ice scour possible during their formation.

The episodic, submarginal channel near the eastern ice margin is a unique display of the power of subglacial water processes. The present valley is incised over 150 m into the glaciated upland surface to the west (towards the center of the ice lobe) and lies at the base of a 400-meter escarpment to the east (towards the ice margin; Fig. 11). Although early reconstruction of the Puget lobe (Mackin, 1941) recognized that an immense flux of meltwater must have filled this channel, only its relatively brief function as a subaerial, ice-marginal channel during the ice recession had been understood. As a result, the channel's long-term occupation by subglacial flow and episodic, catastrophic jokulhlaups (Booth, 1984a,b, 1990) remained uninvestigated until the present reconstruction of subglacial water flow (Fig. 9).

The magnitude of discharges along this channel can be estimated from empirical relationships between lake volume and jokulhlaup discharge. The formula of Clague and Mathews (1973), with modification by Begét (1986) —

$$Q_{max} = 0.0065\ V^{0.69}$$

where Q_{max} = maximum jokulhlaup discharge in m/sec, and
V = initial lake volume in m^3 —

predicts maximum discharges from the impounded ice-marginal lakes (with a release volume of 10^{11} m^3) of order 10^5 m^3/sec. By comparison, this is one to two orders of magnitude less than predictions of Lake Missoula floods (Baker, 1973; Clarke *et al.*, 1984; Begét, 1986) and of inferred sheet floods beneath the Laurentide Ice Sheet (Shaw and Gilbert, 1990), five to ten times greater than reported from computed paleodischarges in the Rocky Mountain Trench (Clague, 1975), and one to two orders of magnitude greater than reported from modern Alaskan ice-dammed lakes (Stone, 1963). Deposits associated with these Puget-lobe jokulhlaups would be largely covered by later recessional-outwash deposits that occupied the same channel. However, record of at least one and probably two discrete episodes is preserved and exposed by layers of 1- to 3-m boulders, located about 2 km downstream of the controlling spillway for the largest lake in the marginal system, Glacial Lake Skykomish (Fig. 12).

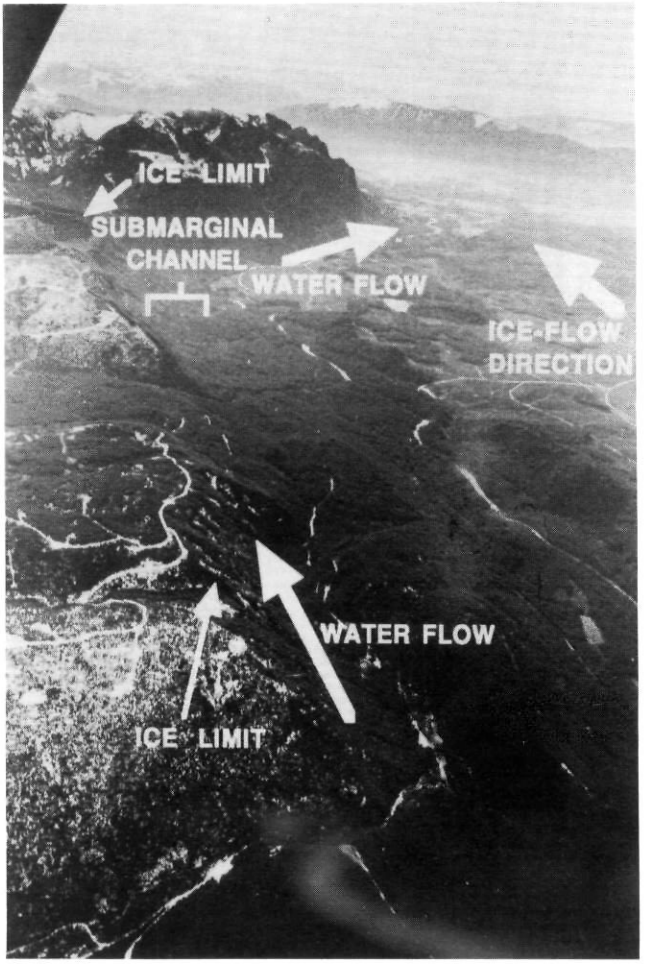


FIGURE 11. Aerial view of the submarginal channel of the eastern Puget lobe, looking south. The left (eastern) escarpment is about 400 m high, and the valley of the submarginal flow is at most about 2 km wide. View covers the southeast corner of Figures 7-10, with the valley of the North Fork Snoqualmie River in the lower left corner.

Vue aérienne vers le sud du chenal de la marge orientale du lobe de Puget. L'escarpement à gauche a une puissance de 400 m et la vallée du chenal mesure au plus 2 km de large. La photo couvre le coin sud-est des figures 7 à 10, avec dans le coin inférieur gauche, la vallée de North Fork Snoqualmie River.

WATER PRESSURE AND THE PROPERTIES OF SUBGLACIAL SEDIMENT

Basal drag imposed by sliding

The dual conditions of high sliding velocities and low basal shear stresses, reconstructed for most of the Puget lobe's length, constrain any hypothesis on the state of the ice-bed interface. Calculations of the drag imposed by a rigid bed of till (Brown *et al.*, 1987) show that even with generous assumptions of ice viscosity, cavitation, and interference between the projection of till clasts extending up into the ice, the drag on ice sliding at 500 m/a is about one order of magnitude greater than the driving shear stress of the ice. Thus bed deformation or ice-bed separation, both implying pervasively high pore-water pressures at the glacier bed, must have occurred.

FIGURE 12. Gravel deposits of one or more jokulhlaups, deposited downstream of Glacial Lake Skykomish above the valley of the North Fork Tolt River (Booth, 1986b). Largest boulders are about 3 m in diameter; note figure (circled) for scale.

Dépôts de graviers résultant d'une ou plusieurs débâcles glaciaires, accumulés en aval du Lac glaciaire Skykomish, au-dessus de la vallée de North Fork Tolt River (Booth, 1986b). Les plus gros blocs mesurent 3 m de diamètre.

Sufficient reduction of basal drag can be achieved by allowing larger clasts to plough along the bed as they melt out. To allow ploughing, however, the strength of the subglacial sediment must be substantially lower than if the full weight of the ice rested on the bed sediments. Based on the reconstructed thickness of the ice sheet, a reduction of at least 90% is required and could have occurred only by a commensurate increase in the subglacial water pressure (Brown *et al.*, 1987). Pervasive till deformation is also a possible mechanism for reduced drag and requires even greater water pressures (see below). Thus under any explaination for the rapid reconstructed sliding rates, the ice lobe must have been nearly floating.

Subglacial Water Pressure in the Substrate

The generation of high subglacial water pressures has been modeled in a number of settings using a variety of techniques (*e.g.*, Boulton *et al.*, 1974; Boulton and Jones, 1979; Shoemaker, 1986). In the subglacial environment, high pressures can result from either transient loading, by advance of the ice margin over saturated sediment, or by steady-state production of basal meltwater within the ice-sheet interior. Using the general solutions for groundwater flow based on the analogy with heat conduction (Bredehoeft and Hanshaw, 1968; Hanshaw and Bredehoeft, 1968), the transient condition results in a zone of influence of only a few tens of meters in width (Booth, 1984b). The steady-state condition, however, is more significant to the glacier as a whole.

The basis of any model of water movement in subglacial sediment is relatively simple. Assuming a source of basal melting only, water enters the till layer at a rate q_o (the volume per unit time per unit area) and drains into a continuous sub-till aquifer of high (but finite) permeability k_s. In the Puget Lowland, this aquifer layer corresponds to sand-dominated outwash deposits that are widespread beneath the overlying till of the last glaciation (Liesch *et al.*, 1963; Garling and Molenaar, 1965; Luzier, 1969). The meltwater is assumed to enter this high-permeability layer along its entire length x, from $x = 0$ at the glacier terminus to $x = L$, the total glacier length. The total horizontal flux of water through this layer per unit width is therefore the product of q_o and L-x, and the flux per unit area of aquifer is q_o (L-x)/D, where D is the aquifer thickness. Integrating Darcy's Law,

$$q = k_s \left(\frac{dp}{dx}\right)$$

with the boundary condition that the pore pressure drops to zero at the terminus, yields:

$$p = \frac{q_o}{Dk_s} \times \frac{Lx-x^2}{2}$$

where q is the local water flux through a unit area of aquifer and p is the pore pressure at position x. Using values appropriate to the Puget lobe of $q_o = 0.05$ m/a, $D = 100$ m, $k_s = 1 \times 10^{-4}$ m/sec, and $L = 200$ km (Rothlisberger, 1968; Olmstead, 1969; Booth, 1984b, 1986a), calculated values of p exceed the overburden pressure for all $x > 0$. This condition is physically impossible: so groundwater flow alone cannot even drain the basal meltwater, let alone the much greater volume of surface meltwater that may also have reached the glacier bed.

This simple conclusion of basal meltwater movement results from the size of the Puget lobe, its high sliding velocity, and the consequent high flux of meltwater that would be required to drain through the subglacial aquifer. The assumptions are conservative, because any discontinuity or thinning of the aquifer, or substantially lower till permeability to impede the entry of subglacial meltwater into this aquifer, would further amplify the result. Conversely, the groundwater flow could drain the net production of meltwater only by increasing the presumed permeability of the outwash to values that nowhere have been reported in the region, or by reducing the total length of the ice sheet (L) by an order of magnitude or more, appropriate to much shorter valley glaciers. Equivalent conditions of subglacial water flowing at or near the overburden pressure also have been inferred (Alley *et al.*, 1987) and recently observed (Engelhardt *et al.*, 1990) at the base of Ice Stream B in Antarctica, a temperate-based glacier with remarkably similar parameters to those of the Puget lobe (total overbur-

den = 9 MPa, basal shear stress = 50 kPa, and a largely soft-sediment bed).

Effect of subglacial tunnels on pore-water pressure

Because most of the meltwater production cannot be drained through the substrate, pressurized subglacial drainage must occur. The flow is likely to be concentrated into channels at the glacier bed (Stenborg, 1968; Shreve, 1972; Walder and Hallet, 1979). The pressure in these channels must be lower than the overburden, however, to allow fresh ice to replace the ice melting off the tunnel walls. Thus the presence of tunnels reduces the pore pressure in subglacial sediments otherwise predicted by simple hydrostatics. This effect was investigated by Lliboutry (1983), with a major influence predicted from the influx of warm groundwater and the consequent increase in tunnel melting rate. This condition is judged highly unlikely beneath an extensive ice sheet.

Two other factors, however, are more relevant to pore-pressure reduction in tunnels. First is the reduction in pressure from potential energy losses from the flowing water, resulting in melting of the tunnel walls. This effect reduces pore pressures in the ice-sheet interior by only a few percent from that of the net overburden (Fig. 13a). This reduction becomes more significant within a few kilometers of the terminus, however, where the tunnel melt rate can exceed the closure rate and the water pressure drops to zero (Fig. 13b). This effect was further investigated by Hooke (1984), who developed a relationship between the depressurizing of tunnels and the water discharge, bed slope, and ice thickness. Where bed slopes are low, zero-pressure tunnels are predicted with his formulation beneath Puget-lobe ice thicknesses less than about 200 m.

A related factor in water-pressure reduction, the seasonal drainage of tunnels near the ice margin (Mathews, 1964; Lliboutry, 1983), can also result in a local, transient reduction of pore-water pressure that may reach farther upglacier than steady-state drainage. Propagating through the surrounding sediment, each pressure drop would raise the effective normal stress on the subglacial sediments (Booth, 1984b). Any effect will be localized, however, within a few kilometers of the ice margin, where viscous deformation of the relatively thin ice is too slow to close tunnels rapidly at the end of the melt season.

Thus slower flow rates under thinner ice allow the water pressure in tunnels to decline towards the terminus. The effective normal stress, the difference between the total ice overburden and the basal water pressure, will therefore *increase* towards the ice margin, because the water pressure drops downglacier more rapidly than the ice sheet thins (Fig. 13b; see also Boulton and Paul, 1976).

Geologic evidence of subglacial water pressures

The history of water-pressure fluctuations in sediment can be inferred only under certain conditions (Shaw and Gilbert, 1990). More commonly, only the *maximum* effective normal stress ever applied to the sediment is recorded, typically by the overconsolidation of clay. If the maximum thickness of ice is known, then the difference between total and effective normal stress represents the minimum water pressure in the sediment during loading.

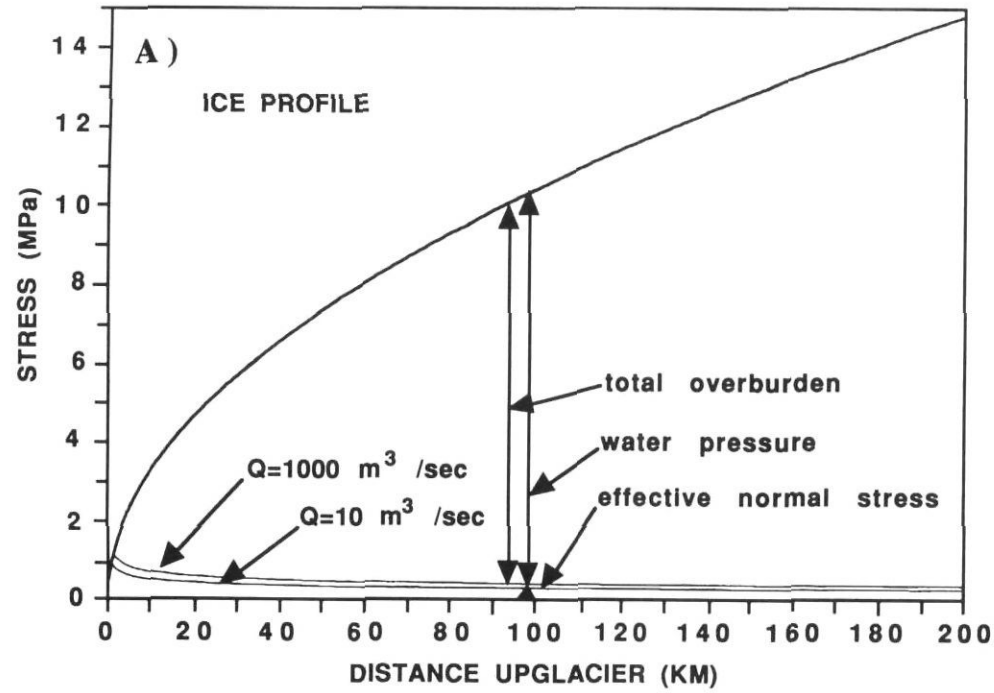
In the Puget Lowland, such data are locally available. Measured overconsolidation data in clay sediment, widely distributed as preglacial and proglacial lacustrine deposits, were compiled by Laprade (1982). All were from the Seattle area, where the ice at maximum applied a total overburden pressure of about 10 MPa. Maximum measured overconsolidation values, however, were 3 MPa with the great majority under 1.5 MPa. Thus *minimum* pore-water pressures were on the order of 80-90% of the maximum ice overburden and could have been even higher during most of the ice occupation.

Brown *et al.* (1987) also argue for high subglacial water pressure from the nature of observed shearing of subglacial sediment. Local deformation of till in the Puget Lowland is evident by numerous exposures of fractured, swirled, pinched, and boudinaged layers. Based on measured strength parameters in till, showing cohesion values near zero and internal friction angles between 30 and 40° (Linell and Shea, 1960), effective normal stresses of less than about 100 kPa would be necessary for sediment deformation under the 50-kPa basal shear stress applied by the ice. This in turn requires pore-water pressures of about 99% of the total ice overburden in the middle of the Puget lobe, for as long as sediment deformation occurred. On rheologic grounds, Alley (1989) also argues for very low effective normal stresses (about 10 kPa) wherever the glacier bed is underlain by soft sediment and subglacial water flow is not channelized. In the Puget Lowland, only near the ice margin would episodic emptying of subglacial tunnels, allowed by slow deformation of thin ice, have reduced water pressures and locally inhibited the sediment shearing so commonly observed.

DISCRIMINATION OF "INTERIOR" AND "MARGINAL" ZONES

An intuitive division has often been implied by the division of an ice sheet into "interior" and "marginal" zones (*e.g.*, Smalley and Unwin, 1968; Mickelson *et al.*, 1979; Mooers, 1990). The reconstructed thermal regime of the Laurentide Ice Sheet has motivated much of this discussion, because the processes beneath a frozen toe are likely to differ greatly from those beneath the temperate basal ice of the interior. Yet even beneath an ice sheet that was temperate throughout, such as the Puget lobe, the discrimination is warranted and useful.

Within the interior zone of the Puget lobe, and probably most other temperate ice sheets with a high mass flux associated with a maritime climate, water derived from basal melting alone is ample to produce subglacial water pressures approaching the ice pressure. Irrespective of the spatial variability in sliding velocity, substrate permeability, and tunnel location, very high basal pore pressures should be pervasive. Thus both effective normal stress and till strengths should be uniformly low, because the slow drainage of basal meltwater swamps all potential sources of variability, leaving the bulk of the ice-covered terrain under nearly uniform conditions. Even with fortuitously improved subglacial drainage, the variability of measured till properties should be much lower than the full range of overburden loads applied by the overlying ice (*e.g.*, van Gelder *et al.*, 1990).



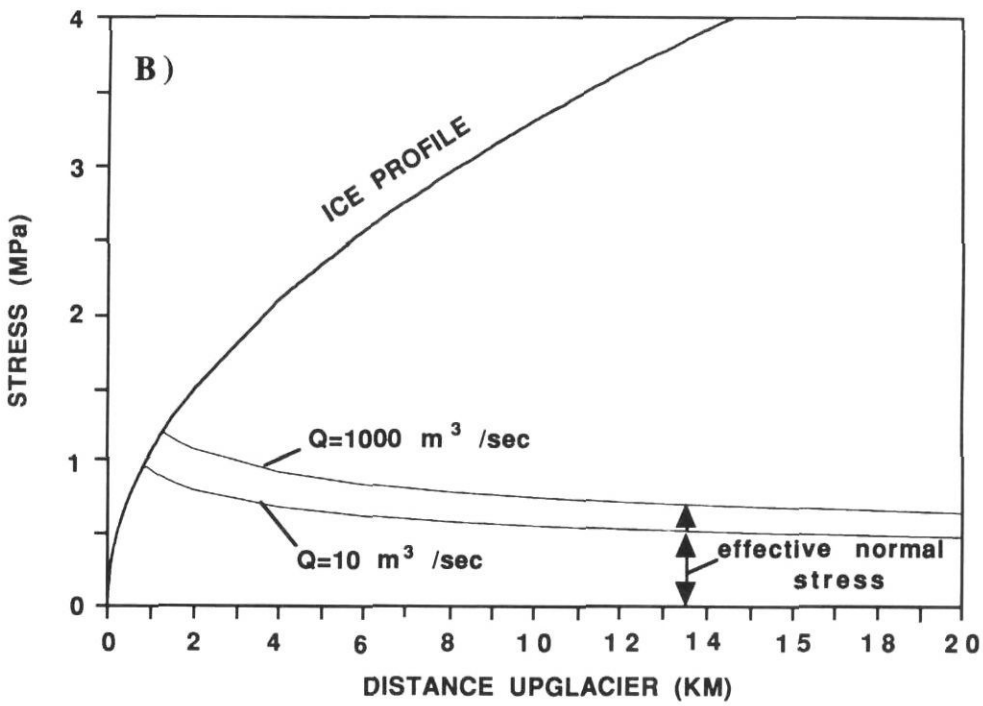


FIGURE 13. A. Comparison of effective and total overburden stresses under the centerline profile of the Puget Lobe, based on Lliboutry (1983) and available data on basal meltwater production and subglacial sediment properties. Over the vast majority of the ice lobe's interior, effective stresses are uniformly low with very little sensitivity either to position or to subglacial tunnel discharges, graphed for two sample discharges that bracket likely tunnel discharges beneath the ice lobe. B. Expanded scale of the marginal zone of the ice lobe in Figure 13a. The effective stress rises to equal the overburden only within 2 km of the margin, achieving a maximum value of about 1 MPa. Seasonal drainage of tunnels, if active farther upglacier, could locally and episodically raise the effective stress by perhaps a factor of 3. Retreat of an active ice margin would eventually subject the entire interior zone to the same magnitude of overconsolidation.

A. Comparison entre les contraintes réelles et totales sous le lobe de Puget, fondée sur Lliboutry (1983) et les données disponibles sur l'apport en eau de fonte et les propriétés des sédiments sous-glaciaires. Sur la plus grande partie de l'intérieur du lobe, les contraintes réelles sont uniformément faibles et peu sensibles à l'emplacement ou aux débits des tunnels sous-glaciaires calculés pour deux types de débits vraisemblablement représentatifs des débits des tunnels situés sous le lobe. B. Zone marginale du lobe glaciaire à plus grande échelle que dans A. La contrainte réelle augmente pour devenir égale à la couverture de glace à seulement 2 km de la marge et atteindre une valeur maximale de 1 MPa. Le drainage saisonnier des tunnels, s'il se fait plus en amont, pourrait localement et à l'occasion augmenter la contrainte réelle par peut-être un facteur de 3. Le recul d'une marge glaciaire active pourrait en fin de compte soumettre toute la zone interne à une surconsolidation des sédiments de même amplitude.

Farther downglacier, pore pressures decrease more rapidly than the ice thins. This relaxation, however, is not uniform — the intrinsic spatial variability of the substrate and the subglacial water system becomes progressively more significant. Areas of the bed near to tunnels, or underlain by thicker or more permeable sediment, will experience comparatively greater pore-pressure declines, and thus greater increases in effective normal stresses, than other areas under equivalent thicknesses of ice but with less favorable conditions for drainage. Within a few kilometers of the margin, pressure in subglacial tunnels may drop to atmospheric levels for a substantial fraction of each year, allowing overconsolidation of the bed that is adjacent to the empty tunnel by the full ice overburden load.

This transition between near-uniform conditions upglacier and spatially and temporally variable conditions downglacier provides a physical basis for the discrimination of "interior" and "marginal" zones. For the Puget lobe, the calculated rate of

pore- and tunnel-pressure dissipation (Fig. 13), together with typical measured overconsolidation of 1-3 MPa (Laprade, 1982), suggest that the marginal zone is less than 10 km wide. This zone includes less than 3% of the total length of the Puget lobe; it is the only region where basal conditions were significantly heterogeneous, and unpredictably so. Upglacier, conditions are much more uniform. Across the Puget Lowland, and presumably other glaciated areas as well, the consequences of this uniformity include the low strength of subglacial till, its homogeneity over hundreds of square kilometers (see also Karrow, 1976), and the potential for extensive ice-bed separation with abundant interbedded water-worked sediment in the basal deposits (see, for example, Easterbrook, 1968; Booth, 1990; also Shaw, 1979; Eyles *et al.*, 1982).

The non-uniform conditions inferred for the marginal zone should migrate along with the backwasting of an active ice front. They may therefore affect every part of a till-covered region. Because effective stresses are higher near the ice margin, this state will be superimposed over sediments that were deposited, under a very different stress regime, in the ice-sheet interior. What is exposed today, throughout the Puget Lowland and perhaps in most glaciated regions, are sediments with properties inherited from this last-stage retreat. They do not necessarily reflect conditions during the vast majority of ice occupation.

CONCLUSIONS

The geologic record of glaciation reflects a variety of physical processes by the agents of ice and water. Detailed observations can provide the chronology, the location, and perhaps some constraints on the processes by which these agents accomplished their work. Yet to interpret more fully their action, their relative importance, and their effect on the landscape, a physical reconstruction of the ice-water system is necessary.

The Puget lobe of the Cordilleran Ice Sheet offers a sufficient geologic record to accomplish this reconstruction. Its extent is deduced from ice limits and flow-direction indicators; by analogy to modern maritime glaciers, its mass balance can be estimated as well. Both ice and water fluxes were vigorous, with sliding velocities in excess of 500 m/a and water discharges of nearly 1×10^{11} m³/a (about 3000 m³/sec).

The flow of this subglacial water produced distinctive landforms, particularly near the lateral margins of the ice lobe's ablation zone. Where analyzed in detail, a roughly dendritic channel pattern, etched into both till and bedrock, drains the interior part of these near-marginal areas. Flow was downglacier and margin-ward, following the regional slope of the ice-sheet surface, and flowed up and down the bed topography as dictated by the gradients in hydraulic head at the base of the ice. Near to the eastern ice margin, a single channel of much larger size collected both interior meltwater and nonglacial runoff from the adjacent mountains. It drained subglacially and episodically, with ice-dammed lakes impounded in the adjacent mountain valleys releasing their water to this submarginal channel as episodic jokulhlaups.

The rate of production of basal meltwater, coupled with the longitudinal extent of the Puget lobe, generated very high water pressure at the base of the ice sheet, sufficient to establish near-hydrostatic conditions over most of the interior zone. Till strength was very low, suggesting extensive ice-bed separation under an essentially floating glacier.

Close to the ice margin, the release of water pressure via subglacial tunnels was sufficient to raise effective normal stresses to significant fractions of the total ice overburden. The recorded overconsolidation of sediments, typically about one-tenth of the ice-maximum overburden, reflects the marginal conditions of moderately thinner ice but much lower water pressures. This correspondence emphasizes the importance of upglacier migration of the marginal zone during ice retreat, when the conditions unique to the outermost few kilometers of an ice sheet can be superimposed over sediments with a much different depositional and deformational history.

The Puget lobe is somewhat unusual amongst the vanished Pleistocene ice sheets, in that fortuitous regional topography has recorded its shape and extent in considerable detail. Yet the physical behavior of the ice and water here is no different than beneath any other temperate ice sheet. Thus many of these results, particularly the pattern of subglacial meltwater flow and the variability of subglacial pore-water pressures, should be readily transferrable to other glaciers. Indeed, these approaches were developed from, and have been applied to, modern glaciers for decades. What Pleistocene glaciers, in contrast to their modern successors, lack in measurable parameters of the ice they gain in providing unrestricted access to their beds. An analysis of their physical processes, such as offered here for the Puget lobe, provides some of the tools to use more fully that information.

ACKNOWLEDGMENTS

The development of these ideas have benefitted greatly from discussions with Bernard Hallet, Charles Raymond, Steven Porter, Richard Waitt, Jr., and Rowland Tabor. Most of the original field work occurred while mapping bedrock and surficial deposits for the U.S. Geological Survey, and later for King County, throughout the south-çentral and eastern Puget Lowland. Bernard Hallet provided countless suggestions on earlier drafts of this manuscript, and his assistance is gratefully recorded here. Dr. J. Begét and L. Jackson have reviewed the paper and provided useful suggestions.

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