Channel networks carved by subglacial water: Observations and reconstruction in the eastern Puget Lowland of Washington

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ABSTRACT

Channels of inferred subglacial fluvial origin ("channelways") are conspicuous across the recently deglaciated eastern Puget Lowland of Washington State. Exploiting the unusually well defined geometry of the former Puget-lobe ice sheet, we have reconstructed the subglacial hydraulic head in this region to better understand the location and orientation of these channels. We used established principles of channelized water flow under static ice because a three-dimensional linear-viscous ice model shows that sliding-induced pressure variations can be ignored at all but the smallest scales of individual landforms. Two types of predicted subglacial flow paths emerge. One drained continuously from the interior toward the ice-sheet margin; the other diverted water into a series of valleys in the flanking Cascade Range, forming ice-marginal lakes that discharged episodically and probably catastrophically. These reconstructed flow paths correlate very well with the independently mapped distribution of channelways, and also of ice-marginal lakes, throughout this area. Our study suggests that the formation and preservation of these landforms require both the concentration of subglacial water, notably near ablation-zone ice margins, and favorable preglacial bedrock topography. Where these elements occur together, landforms of subglacial fluvial origin should be recognizable across the now-exposed beds of other former ice sheets as well.

INTRODUCTION

Although the presence of water at the base of temperate ice sheets is well known, the geomorphic effects of that water are seldom examined in detail. Small-scale features, such as sichelwannen and potholes, and large-scale features, such as eskers and tunnel valleys, are acknowledged expressions of that water. Yet the most commonly recognized characteristics of a "glaciated landscape," particularly erosional landforms, generally are assumed to result from the action of sliding ice, not flowing water.

The Puget lobe of the former Cordilleran ice sheet (Fig. 1) provides an excellent opportunity to investigate the geomorphic products of subglacial water. Distinct valleys inferred to be water excavated ("channelways") abound over much of the area; their distribution can be compared with the drainage pattern of subglacial water, which can be reconstructed with considerable confidence. This reconstruction reflects existing knowledge of the geometry of the former ice sheet, which is unusually complete because of abundant ice-flow direction

indicators and distinct ice limits due to the high relief along both sides of the Puget Lowland (Thorson, 1980; Booth, 1986a).

Herein, we first review the primary characteristics of channelways and discuss their likely subglacial fluvial origin. We then reconstruct the pattern of subglacial water flow, first by considering the physics of channeled water flow under glaciers and ice sheets and then by applying this analysis to a region of the eastern Puget Lowland. A good correlation between predicted subglacial flow paths and observed channelways emerges, which reinforces our geologic interpretation that these features were eroded by subglacial water. This anal-
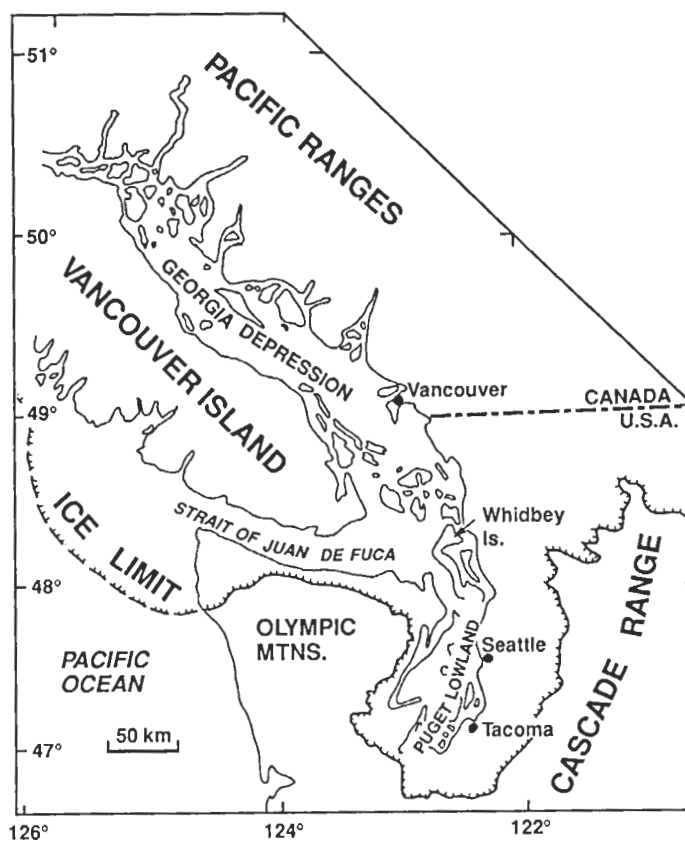

Figure 1. The Puget lobe of the Cordilleran ice sheet at the last-glacial maximum, ca. 15,000 yr B.P.

Figure 2. Index map of modern rivers and specific localities discussed in text, and area of detailed reconstruction of former subglacial drainage.

ysis also suggests the general conditions favoring local dominance of fluvial-erosional processes in previously glaciated environments.

SUBGLACIAL FLUVIAL LANDFORMS OF THE EASTERN PUGET LOWLAND

Setting

The Cordilleran ice sheet expanded into the Puget Lowland at least six times during the Pleistocene (Crandell and others, 1958; Easterbrook and others, 1967; Blunt and others, 1987). The most recent advance, the Vashon stade of the Fraser glaciation of Armstrong and others (1965), provides the best record 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 over a period of 10,000–15,000 yr (Clague, 1981). The ice sheet entered the Fraser Lowland in southern British Columbia some time after 18.3 ± 0.17 ka (GSC-2322; Armstrong and Clague, 1977). In the Seattle area, 160 km to the south, a preglacial age of 15.0 ± 0.4 ka (W-1227; Mullineaux and others, 1965) and a postglacial age of 13.65 ± 0.55 ka (L-346a; Rigg and Gould, 1957) bracket its final 100 km of advance and retreat.

Channeled Features

Along the eastern border of the central Puget Lowland (Figs. 2 and 3), numerous troughs and valleys incise the till-mantled bedrock landscape. Although some have been scoured by ice or eroded by subaerial recessional meltwater, many are inferred to have been eroded by subglacial meltwater. These "channelways" (Booth, 1990), so named to distinguish this inferred mode of subglacial origin, are narrow, sinuous to linear valleys occupied by lakes, bogs, or underfit streams. They have low gradients, typically small drainage areas, and steep sidewalls, which range from 10 m to over 100 m high. Poor exposures along the valley sides mainly show bedrock or till. Small-scale, fluvially eroded bedrock features, as described by Dahl (1965) or Sharpe and Shaw (1989), are sporadically observed. Eskers also run along the floor of at least two channelways in this region (Booth, 1989, 1990). These characteristics are shared with those of the tunnel valleys of central and northern Europe (Schou, 1949; Ehlers, 1981; Grube, 1983).

Channelways are common within ~20 km of the margin of the former Puget lobe (Fig. 3), but they are generally absent in the interior (for example, Booth, 1991a). They can be followed from about latitude 48°15′N south for more than 100 km (Fig. 1) to where bedrock is completely buried under till and recessional outwash deposits (Crandell, 1963; Frizzell and others, 1984).

Although channelways locally can be isolated, they more commonly form dendritic or anastomosing networks. They are particularly prominent across the west-trending bedrock spurs of the Cascade Range that lie transverse to the ice-flow direction. In these settings, they can behead the lateral spur and entirely dissect its lower part, isolating individual bedrock hills (Fig. 4). Between the Skykomish and Tolt Rivers, for example, these hills are nearly equidimensional in plan view, ~100 m high, and four or five times as wide (Fig. 5). The intervening channels typically range from 50 m to 150 m wide and are now floored with recessional and postglacial deposits.

The most extensive channelway network in this region can be traced south from the Skykomish River >50 km through the bedrock uplands near the eastern ice margin. Although originally continuous,
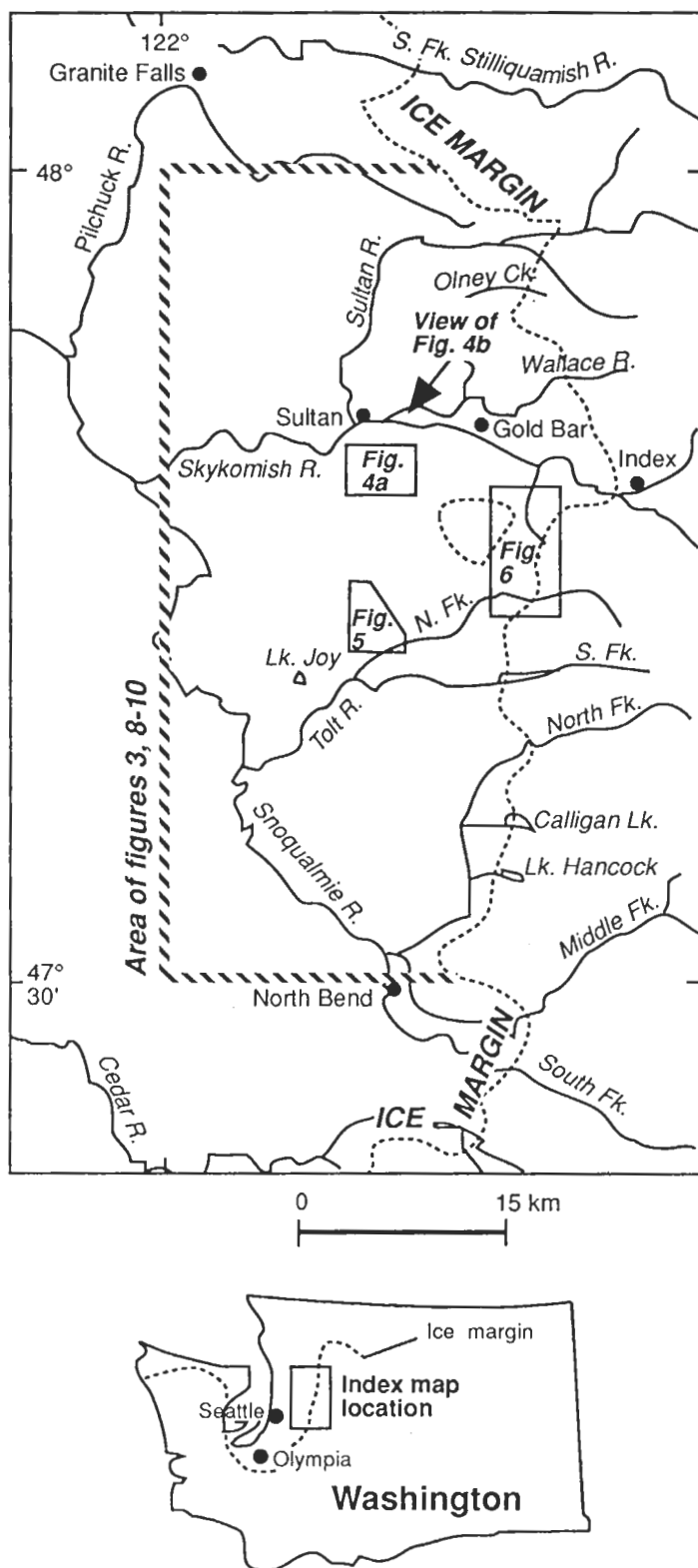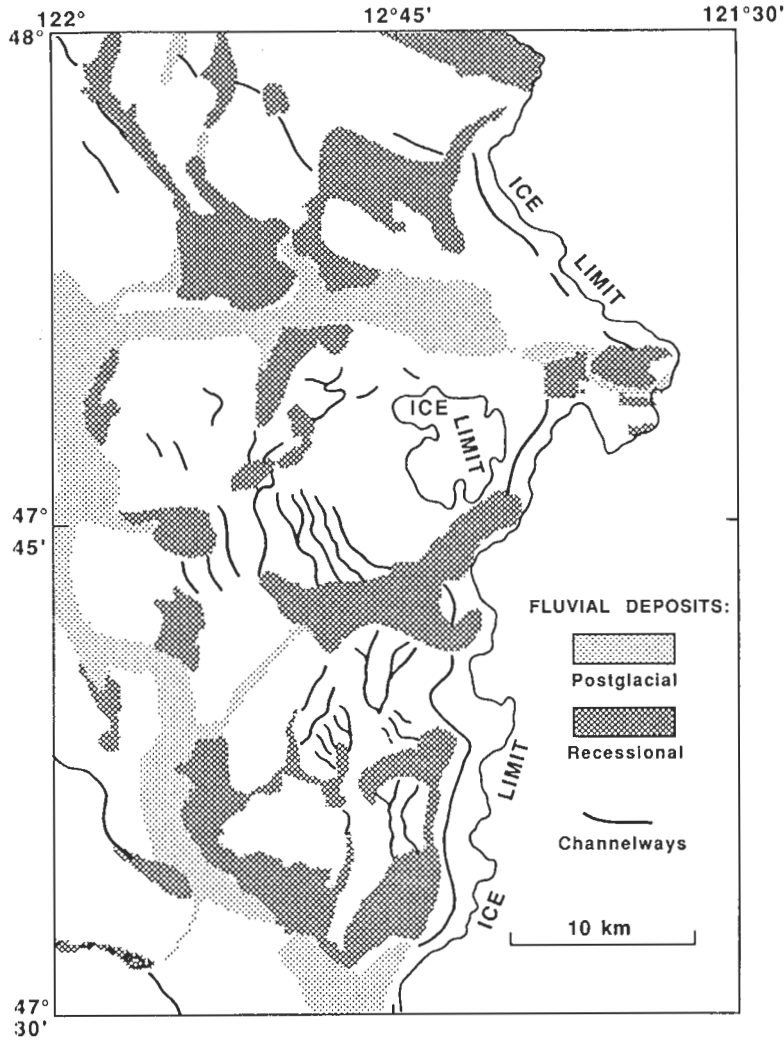
Figure 3. Observed channelways in the eastern Puget Lowland, mapped from Booth (1990). Stippled areas represent major recessional and Holocene alluvial deposits that obscure the former configuration of the subglacial bed.

it now has been broken into isolated segments by the Holocene incision of west-flowing alpine rivers. The northernmost segment consists of a broad trough, almost 1 km wide and 8 km long (Fig. 6), that crosses a major bedrock spur south of the Skykomish River and whose floor lies nearly 500 m below the Vashon ice-maximum limit. Along the eastern edge of this trough, a fault-line scarp separates rocks of greatly different erosional resistance (Tabor and others, 1982) and probably controlled preglacial drainage in this area. The present width and depth of the trough, however, dwarfs any likely erosion from subaerial streams, which is now limited to minor incision (Booth, 1990).

The more southerly segments of this network also lie several hundred meters below, and up to several kilometers within, the Vashon ice limit. The individual channel segments range from 200 to 500 m wide, but except where occupied by the North Fork of the Snoqualmie River (Fig. 2), they carry only trivial modern drainage.

Origin of Channelways

Few processes can plausibly be responsible for the formation and characteristics of the channelways. For example, bedrock structures or significant variation in erodibility on the scale of individual features is generally absent (for example, Fig. 4). Demonstrable fault traces only rarely correspond to channelway locations and are not likely to coincide with the numerous sinuous segments.

Channel excavation by sliding ice also is unlikely. Unlike fjord valleys, the relief of the bed topography is inadequate to channelize the ice flow from a large source area. Indeed, the valleys commonly lie oblique to the ice-flow direction, as shown by independent indicators such as striae. Even where generally aligned with the regional ice flow, most of the valleys are typically deep, narrow, sinuous, and steep walled. They would offer higher resistance to ice flow than the unincised (but also ice-covered) topography around them (Weertman, 1979) and therefore are unlikely to have been sites of preferential incision by ice erosion. Finally, the bedrock knobs in areas of best-developed channelways commonly are not strongly streamlined (see, for example, Fig. 5), a characteristic that intuitively precludes the dominance of erosion by ice (Linton, 1963).

Channelways resemble features attributed primarily to subaerial fluvial erosion in the northeastern United States and termed "overflow channels" or "sluiceways" (Coates and Kirkland, 1974); however, several of their characteristics are inconsistent with subaerial formation (for example, Mannerfelt, 1949; Peel, 1956; Sissons, 1958a, 1958b, 1960, 1963; Derbyshire, 1958, 1961, 1962; Price, 1963; Clapperton,

Figure 4. a. Map view of the geology and topography of multiple channelways (shaded) that dissect a bedrock spur extending west from the Cascade Range (50-m contour interval). Tv = Tertiary volcanics; pTm = Tertiary melange (Tabor and others, 1982). Note that landform development is relatively insensitive to bedrock type. Ice flow is from the northwest.

b. Aerial photograph of the region of Figure 4a. Image spans about 8 km and is looking southwest, obliquely downglacier, with ice flow from the right (northwest) side of the image. The ice limit lies roughly two-thirds up the main ridge, near the left (eastern) edge of the photograph on the skyline. At lower elevations, subglacial meltwater channels have deeply dissected this spur, leaving a characteristic landscape of isolated, nonstreamlined hills.
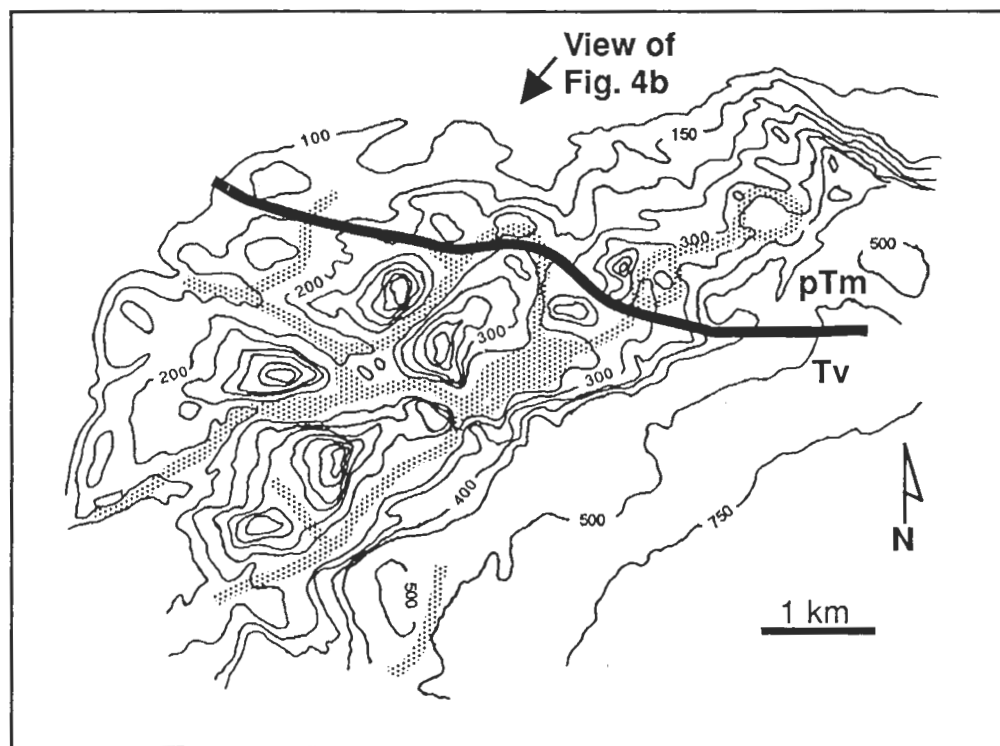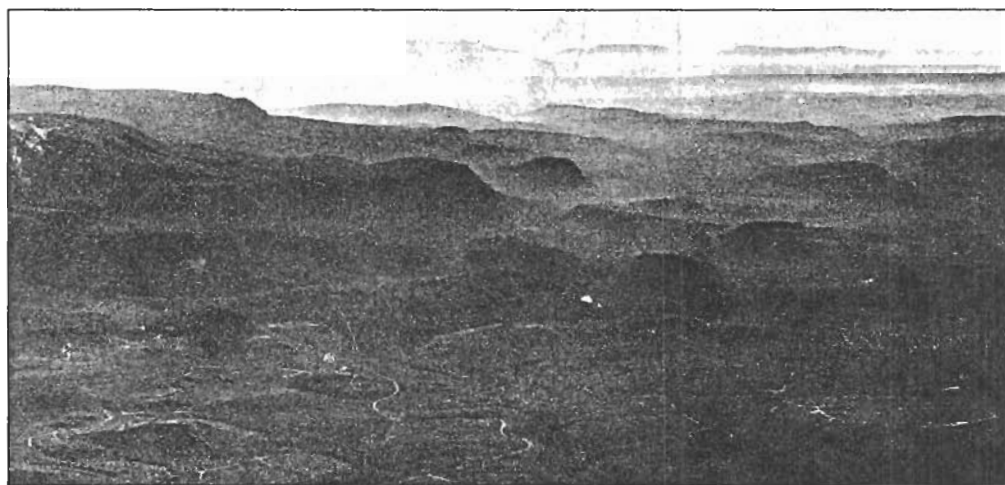
1968, 1971; Pasierbski, 1979; Young, 1980; Ehlers, 1981). Among these characteristics are (1) "humped" or "up-and-down" longitudinal profiles; (2) abrupt termini without deltas or fans; (3) multiple intakes and outflows; (4) presence of till, eskers, or other ice-contact deposits; and (5) perched topographic position without significant drainage area.

Marginal and proglacial drainage has been proposed to explain these characteristics here and elsewhere (for example, Kendall, 1902; Mackin, 1941). Yet ice retreat in this region was so rapid (Thorson, 1980; Booth, 1987) that any water diverted subaerially through these valleys would not have been present long. Catastrophic proglacial

discharges from ice-dammed lakes probably occurred but was quite localized (see below).

Subglacial water is the most plausible agent of channelway formation. It is ubiquitous under modern temperate glaciers (for example, Mathews, 1964; Stenborg, 1968; Engelhardt, 1978) and beneath the Puget lobe in particular (Brown and others, 1987; Booth, 1991b); its erosive potential is also well documented (for example, Chernova, 1981; Gustavson and Boothroyd, 1987). The largest channels found in this area (for example, Fig. 6) invite comparison with the "through valleys" of New England (Coates and Kirkland, 1974), argued by Shaw and Kvill (1984) to owe their formation to vast subglacial melt-
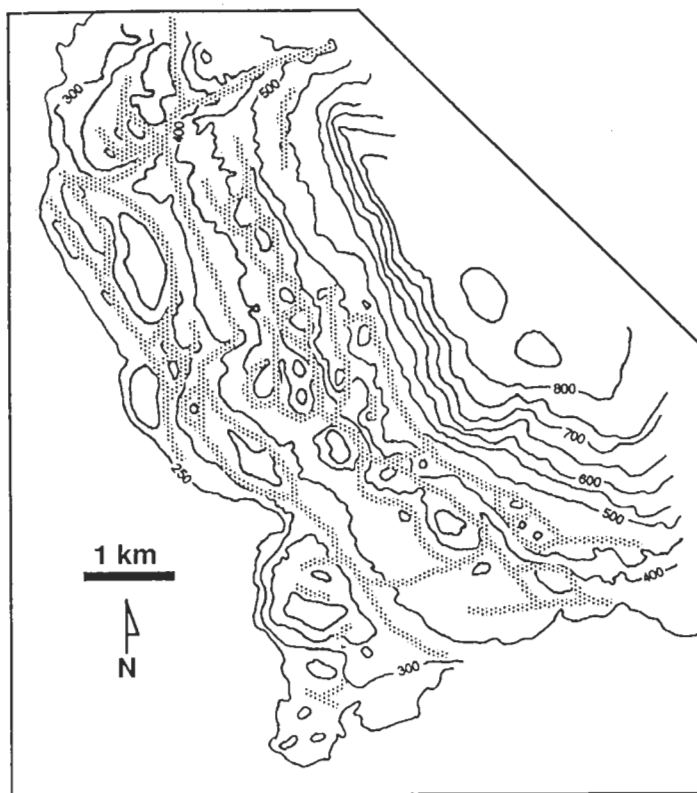
Figure 5. Multiple channelways (shaded), trending northwest-southeast and isolating numerous 0.5-km bedrock hills over most of the map area. Ice flow is from the northwest.

water discharges. These inferences, however, can be augmented considerably by analyzing subglacial water flow to demonstrate that the observed location and orientation of channelways in fact correspond closely to the reconstructed drainage pattern beneath the former ice sheet.

GENERAL PRINCIPLES OF SUBGLACIAL WATER FLOW

Static Ice Analysis

Water flow within or beneath a glacier can be described mathematically in terms analogous to those developed for ground water. Water moves from areas of high total hydraulic head to areas of low head (Shreve, 1972; Rothlisberger, 1972; Nye, 1976). Under static ice the total head (H) at a point is well approximated by the sum of the elevation head (z), the height above a chosen datum such as sea level, and the pressure head, which for a steady-state water-filled tunnel is nearly equal to the pressure exerted by the overlying ice of thickness h (Shreve, 1972), typically expressed as the height of a column of water that would exert the same pressure. Thus, in units of length,

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

where ρ_w = water density (= 1,000 kg · m^{-3}), and ρ_i = ice density (≈ 900 kg · m^{-3}). Although tunnels may be only partially filled with water (and therefore unpressurized) where they plunge steeply down-

glacier (Hooke, 1984) or where they receive ground water at about +1 °C or more (Lliboutry, 1983, equation 20), neither condition is likely beneath an extensive ice sheet.

The two-dimensional pattern of subglacial water flow is readily obtained by differentiating equation 1 in the downglacier direction (by convention here, the direction of increasing x):

$$\frac{\partial H}{\partial x} = \frac{\partial z_b}{\partial x} + \left(\frac{\rho_i}{\rho_w}\right)\frac{\partial(z_s - z_b)}{\partial x} \qquad (2)$$

where z_b = bed elevation and the ice-surface elevation (z_s) = $z_b + h$. Water will flow downglacier as long as total head H decreases downglacier (that is, $\frac{\partial H}{\partial x} < 0$), which yields

$$\frac{\partial z_s}{\partial x} < -0.11 \left(\frac{\partial z_b}{\partial x}\right). \qquad (3)$$

Thus water flow is influenced by both the ice surface and bed topography, but it tends to follow the ice-slope direction, flowing up as well
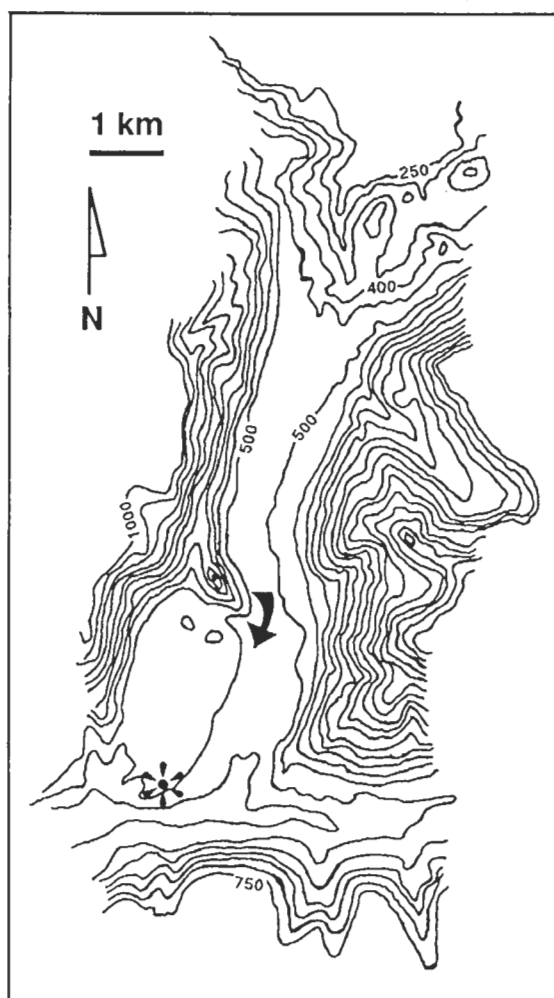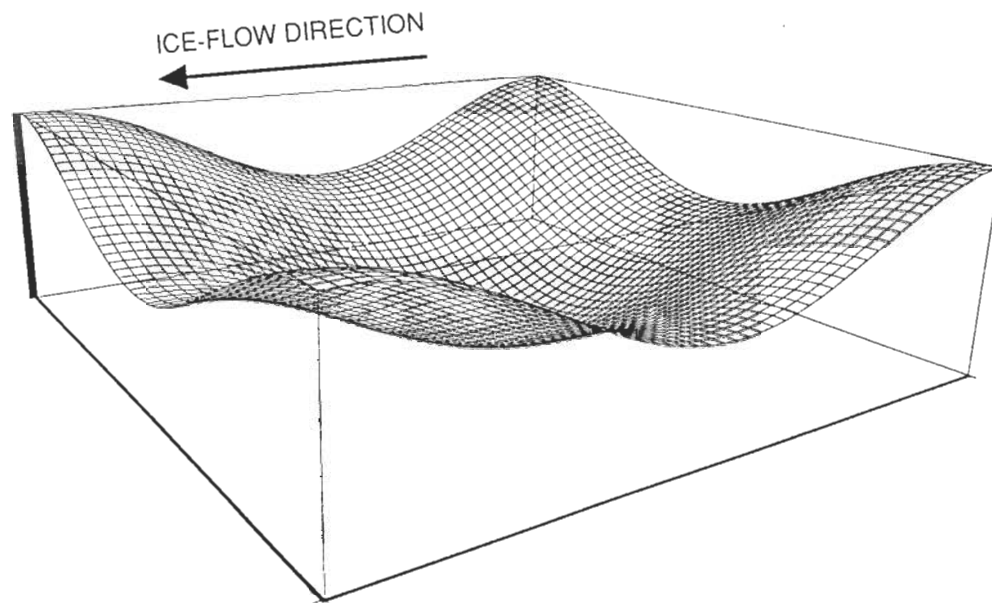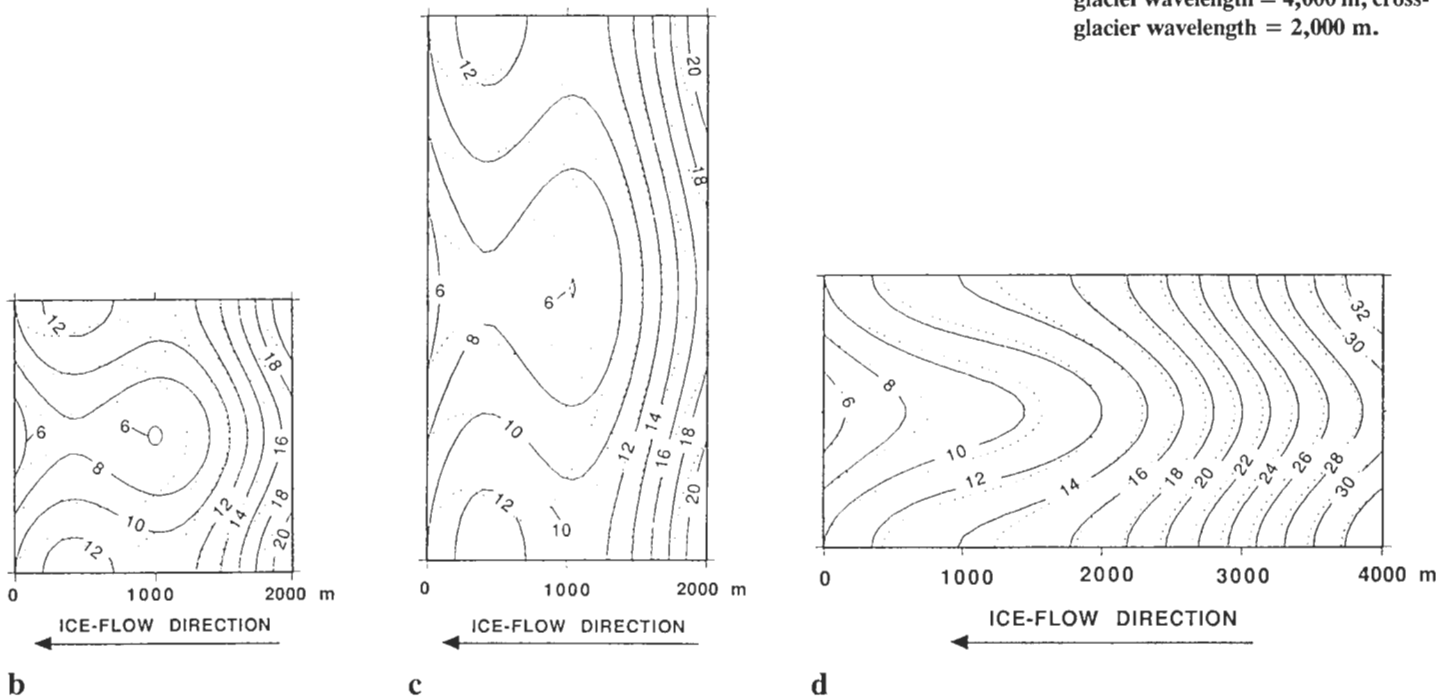

Figure 6. Northern segment of the eastern submarginal channelway system. North of this channel, ice flow was from the west-northwest. Arrow shows approximate location of subglacial spillway for glacial Lake Skykomish; exposure of 1-m to 3-m boulders, deposited by multiple glacial-lake discharges, is located by asterisk.

Figure 7. a. Perspective diagram of bed topography consisting of symmetrical sinusoidal hills. **b–d.** Contours of total hydraulic head beneath sliding ice (solid lines) over idealized hills centered at the corners of each figure. For comparison, the hydraulic head under static ice is also shown (dashed contours). Total topographic relief = 100 m. The pattern of hydraulic head mimics the bed topography, but maxima are shifted upglacier of topographic peaks. Contours are expressed in units of total hydraulic head (m) above an arbitrary datum. (b) Head over symmetrical hills with a 2,000-m wavelength. (c) Head over transverse hills; downglacier wavelength = 2,000 m; cross-glacier wavelength = 4,000 m. (d) Head contours over longitudinal hills; downglacier wavelength = 4,000 m; crossglacier wavelength = 2,000 m.

Pressure-Head Modification by Sliding Ice

Relative Importance. Basal ice motion over a nonplanar bed surface increases the pressure head on stoss slopes and reduces it on lee slopes (Weertman, 1957; Lliboutry, 1968; Nye, 1969). This effect has been discussed qualitatively by Shreve (1972) and Hooke (1989), but the rather minimal consequences of this dynamic component can be demonstrated more quantitatively (see also Shreve, 1985) even for basal ice velocities as high as several hundred meters per year (for

as down the bed topography with only moderate deflection (Shreve, 1972, 1985).

example, Booth, 1986a, for the Puget lobe). The linear glacier sliding theory of Nye (1969) and Kamb (1970) is used, because its inherent inaccuracy is not critical here and is well compensated for by its simplicity.

Using this theory, the pressure distribution over any bed with topographic irregularities that are small relative to the ice thickness can be analyzed. Consider "egg-carton" bed topography (Fig. 7a), consisting of simple sinusoidal variations in elevation both parallel and perpendicular to the ice-flow direction with a horizontal scale of 1 km or more. On this scale, regelation can be neglected, and so the linear solution to the equations of motion (Kamb, 1970, equation 33) yields a dynamic pressure component (p) of:

$$p = -2\eta \, U \, A_x \, h_x^2 \, \sin(h_x \, x) \qquad (4)$$

where η is the effective ice viscosity, U is the sliding velocity, and A_x is the amplitude and h_x is the wave number ($= 2\pi/\lambda_x$ where λ_x is the wavelength) of the topography in the downglacier direction. Only x terms appear in this pressure solution; cross-glacier (y) variation in the pressure head is imposed only by topography, not by sliding.

To establish the relative importance of the various components of the total hydraulic head, consider a simplified representation of the central Puget-lobe ice sheet in which the ice surface is of uniform slope S with elevation $z_s = z_0 - Sx$, and the bed is horizontal except for sinusoidal relief. In the x direction, the total head is given by

$$H(x) = \frac{\rho_i}{\rho_w}(z_0 - Sx) + \left(1 - \frac{\rho_i}{\rho_w}\right) A_x \cos(h_x \, x)$$
$$- (2\eta \, U \, A_x \, h_x^2 \, / \, \rho_w \, g) \sin(h_x \, x) \qquad (5)$$

where total head is determined by the elevation of the ice surface (first term), the bed elevation (second term), and the dynamic pressure (third term, with g = acceleration due to gravity, 9.8 m · s^{-2}). By means of the formula for the cosine of the sum of two angles, Equation 5 can be rearranged:

$$H(x) = \frac{\rho_i}{\rho_w}(z_0 - Sx) + C\left(1 - \frac{\rho_i}{\rho_w}\right) A_x \cos[h_x(x + x_m)] \qquad (6)$$

in which

$$C = [1 + \tan^2(h_x \, x_m)]^{0.5} \qquad (7)$$

and

$$\tan(h_x \, x_m) = \frac{2\eta \, U \, h_x^2}{\left(1 - \dfrac{\rho_i}{\rho_w}\right)\rho_w \, g} \qquad (8)$$

Thus for such topography, the effect of basal sliding is to shift the sinusoidal part of the total head upglacier and increase its amplitude. Although the amount of shift (x_m) cannot exceed one-fourth of the wavelength, the increase in amplitude can vary widely. For the simplified Puget lobe, the ice thickness z_0 is 1,000 m, the surface slope S is 0.006, the amplitude of the bed topography in both x and y directions (A_x and A_y) is 25 m (100 m of total relief), and the wavelength of sinusoidal hills is 2,000 m ($h_x = 0.003$ m^{-1}). Over one wavelength, the static head changes by 12 m due to the ice-surface slope and varies with an amplitude of 5 m due to the sinusoidal bed topography. Using a sliding velocity U of 500 m · a^{-1} and taking the effective ice viscosity η to be 100 kPa · a (1 bar-year; Nye, 1970; Kamb, 1970), the shift x_m is about 250 m, the amplitude of the sinusoidal (dynamic) variation in head is 2.2 m, and thus the increase in the amplitude due to sliding is 20% or less of the total head.

Thus at the scale of a few kilometers, the dynamic component of the pressure head makes a small but not entirely insignificant contribution to the total head. As the size of the analyzed area increases, however, the overall pattern of water flow becomes more and more closely approximated by changes in the static components alone. As a result, large-scale reconstruction of Puget-lobe bed topography and water flow (see below) can disregard the effects of basal sliding altogether. It is nevertheless instructive to consider dynamic effects in the vicinity of individual landforms. These effects are likely to play a

significant role, albeit localized, in both the routing of subglacial water around individual hills and the excavation of the intervening channels.

Local Influence of Elongated Bed Topography. Although unimportant at a scale much beyond a few kilometers, changing the elongation and orientation of topographic features at the bed does affect the local patterns of subglacial water flow under sliding ice, particularly where the local relief is high. Examples of this effect are shown in Figures 7b–7d, which display the contours of equal head resulting from equidimensional hills and troughs (Fig. 7b); "longitudinal topography," with elongated ridges parallel to the ice flow (Fig. 7c); and "transverse topography," with ridges perpendicular to the ice flow (Fig. 7d). As expected from equation 6, the maxima in total head are shifted slightly upglacier relative to the static (that is, nonsliding) condition in each case; drainage is via the longitudinal channels between topographic highs, with the steepest gradient in total head located just downglacier of the channel-flanking knobs.

Differences are also apparent in the magnitude of total-head gradient, which in turn should reflect different rates of water erosion. In particular, bed topographies with shorter downglacier wavelengths yield steeper gradients along channels (compare Figs. 7c and 7d), because the dynamic component of head varies inversely with only the longitudinal wavelength squared (that is, the term h_x^2, proportional to [landform length]$^{-2}$, in equations 4 and 5). As a consequence, any subglacial fluvial erosion along the channels between transverse hills should proceed more rapidly, preferentially deepening those channels, than along the channels of similarly sized features oriented longitudinally.

RECONSTRUCTION OF PUGET-LOBE SUBGLACIAL WATER FLOW

To compare the distribution of observed channelways with the reconstructed pattern of subglacial water flow, we chose a region in the eastern Puget Lowland (Fig. 2) where channelways are particularly well developed and bedrock is typically mantled by only thin (0–3 m) glacial deposits. Two reconstructed elements are required: the bed and the ice surface. The first is shown in a computer-generated topographic map of this region (Fig. 8). Elevations were digitized from USGS 15′ and 7.5′ topographic maps by tracing individual contours. The resulting data points (about 200,000 in total) were then condensed by spatial averaging into a square grid. A 300-m spacing was chosen, yielding a 120- by 200-point grid, to minimize the number of subsequent calculations while maintaining an accurate representation of the overall pattern of water flow. This scale also obliterates all but the largest channelways from the topographic reconstruction, reducing the analytical circularity of having the presence of existing valleys predetermine the location of predicted flow paths.

Where bedrock is absent, notably in the valleys of the Skykomish and Snoqualmie Rivers (Fig. 2), the exact topography of the glacier bed is indeterminant. However, these two valleys contain abundant recessional deposits, and so they must also have existed subglacially. Therefore, no correction is made for postglacial modification of the bed in these valleys or elsewhere in the map area, as the modification does not affect either the overall drainage pattern or the areas of best-developed and best-preserved channelways. Postglacial isostatic rebound (Thorson, 1989) also is ignored, as the gradient change ($\sim 10^{-3}$, at maximum) negligibly affects the reconstructed water-flow patterns.

The Vashon ice-sheet surface at its maximum stand is equivalently reconstructed (Fig. 9), based on geologic evidence that con-
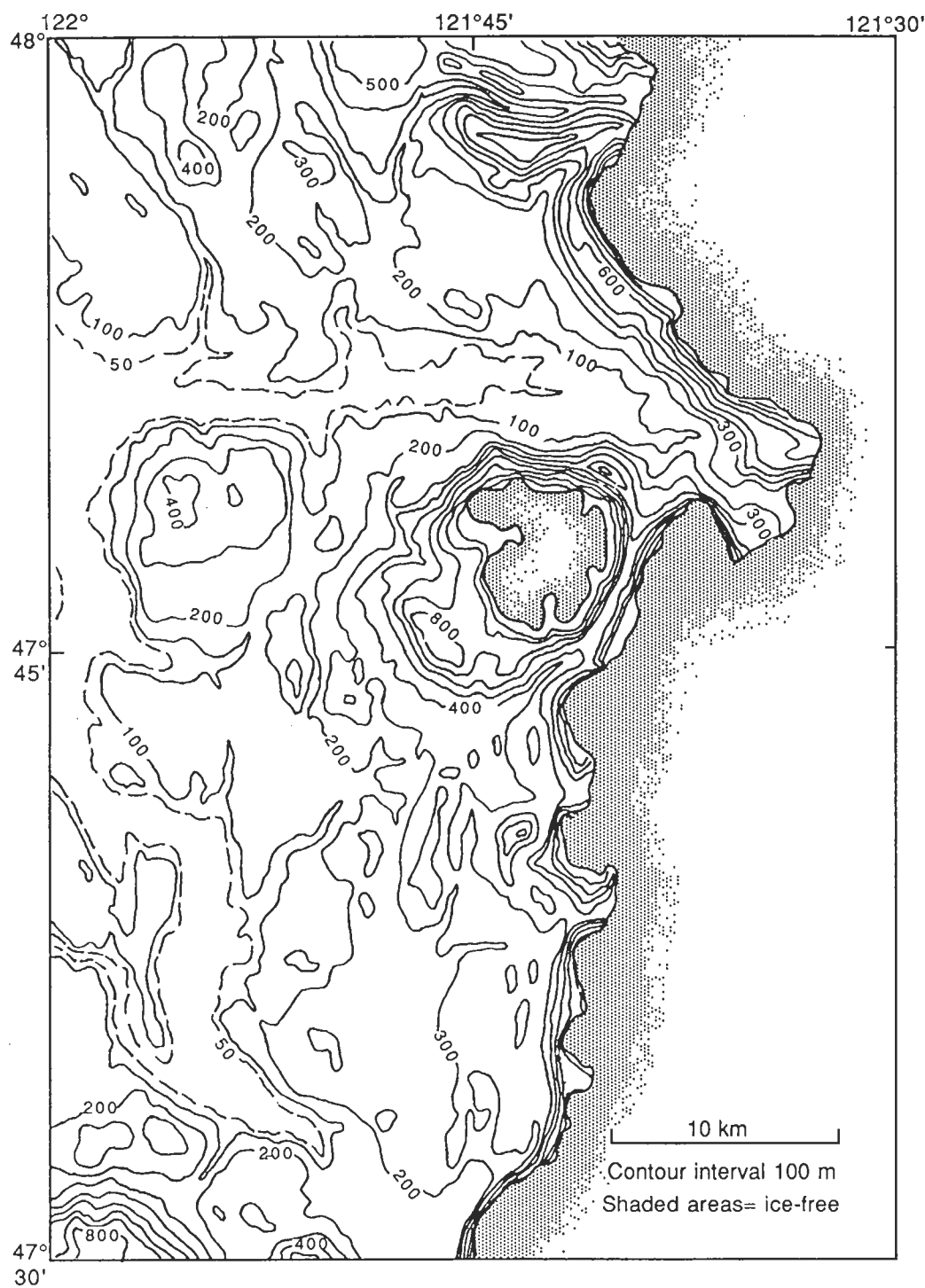
Figure 8. Ground topography of the Skykomish-Snoqualmie region in the eastern Puget Lowland. Contour interval = 100 m (dashed contour = 50 m); ice-free areas are shaded. The Vashon-age ice limit, drawn along the eastern boundary, is approximated at the mouth of each alpine valley. Digitized from Monroe 15', Mount Si 15', Lake Joy 7.5', Carnation 7.5', Fall City 7.5', and Snoqualmie 7.5' USGS topographic maps.

strains the ice limit along the eastern margin (Booth, 1990), the assumed parallelism between ice-flow direction indicators and ice-surface gradient (Weertman, 1964), and the full Vashon-age Puget-lobe reconstruction by Thorson (1980). The contour interval, 20 m, reflects typical imprecision in ice-limit data collected in the field (see Booth, 1986a, for further discussion). The latest (Vashon) ice advance is modeled here not only because its effects are best displayed, but also because the only other (older) glacial advance recognized in this area

had an ice limit that lay almost uniformly 50 m above that of the Vashon (Booth, 1990), altering neither the ice-surface gradient nor the pattern of total subglacial head.

Using these data and Equation 1, the total two-dimensional head field is calculated (Fig. 10). Subglacial flow paths are traced down the reconstructed head gradient from the north and west map boundaries wherever valleys, defined by the distribution of total head, enter the map area. A few additional flow paths are initiated within the map area
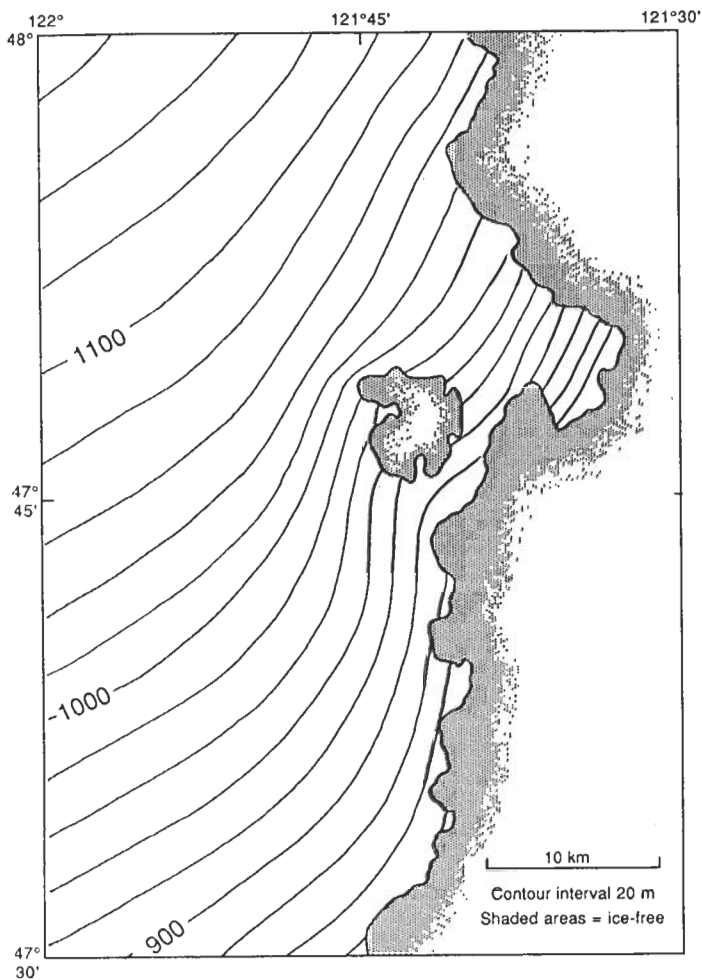
Figure 9. Ice-surface topography at Vashon ice-maximum stage. Contour interval = 20 m; ice-free areas are shaded. Values constrained at western and southern borders by Thorson (1980) and on eastern edge by Booth (1990).

at the upstream end of observed channelways, but their subsequent paths are based only on the reconstructed head gradient.

Based on the scaling results of the previous section, this reconstruction considers only the static components of subglacial head and neglects the contribution of sliding. It also neglects perturbations in the ice surface caused by flow over undulations in the bed (Robin, 1967). For the Puget lobe (slopes <0.01 and sliding velocities 100–500 m · a^{-1}), rises in the ice surface are virtually in phase with those of the bed and are reduced in amplitude (Hutter, 1983, p. 203, 224–225, Figs. 4.8–4.9). This perturbation therefore increases the magnitude of local highs in the head surface but not their overall pattern.

From the reconstruction, two distinct categories of flow paths are revealed. The first forms a crudely dendritic network that trends southeast, reflecting the overall control of the ice-surface slope. Water in this network follows routes of monotonically decreasing head; branches diverge where the direction of the reconstructed head gradient is ambiguous and the underlying bed material has been modified since deglaciation (that is, recessional or postglacial alluvium). These divergent paths may be artifacts of the reconstruction. Yet anastomosing channelways are observed in the field and may, in part, reflect

minor shifts in the total-head distribution over time or genuinely divergent, simultaneously occupied flow paths.

The second category of reconstructed flow paths follows a single submarginal valley near the eastern boundary of the ice sheet. Total head, however, does not decrease monotonically along this channel. Instead, local saddles along this path block the flow and would impound water behind them into the alpine valleys immediately upglacier (that is, north) of each saddle (Sultan, Skykomish, North Fork Tolt, South Fork Tolt, and North Fork Snoqualmie, approximately located in Fig. 10). In these valleys, thick accumulations of lacustrine sediments, derived in large part from the ice-sheet interior (Booth, 1986b, 1990), confirm the existence of these inferred lakes. As the lake levels rose, water would have eventually overtopped the controlling subglacial spillway, leading to catastrophic and repeated lake drainage (Nye, 1976; see Stone, 1963, and Post and Mayo, 1971, for modern examples). Thus the submarginal flow path of Figure 10 almost certainly was episodic in nature.

These five ice-marginal lakes lie in series, each draining via the submarginal channel into the next to the south. Even partial drainage of the largest lake, which occupied the modern Skykomish River valley with an estimated maximum volume of 120 km³, 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 West, 1988; see Porter, 1977, for justification of their applicability) and subaerial drainage areas, would have been only a few decades (Table 1). Thus even with the relatively rapid advance and retreat of the Puget lobe, multiple subglacial discharges along the Cascade rangefront probably were inevitable during ice occupation.

Other subglacial hydraulic saddles lie east of the main submarginal channel, but they are not evident at the scale of Figure 10 (see, however, Booth, 1986b). These saddles would have similarly impounded lakes in the adjacent valleys of the Cascade Range, but the lakes would have filled only from local drainage of surface meltwater off the ice sheet and from the upper alpine basins themselves.

COMPARISON OF OBSERVED AND RECONSTRUCTED DRAINAGE FEATURES

Distribution of Channelways

The distribution of field-identified channelways correlates very well with the predicted subglacial flow paths with only minor exceptions (Fig. 11). All of the major channelways (including the major valleys of the Skykomish and Snoqualmie Rivers) coincide or lie very close to reconstructed flow paths. At a finer scale, many small channelways initiate on uplands, particularly near the crest of topographic divides where the subaerial drainage area is very small but where subglacial fluvial incision is most likely because subglacial stream power (calculated from the product of discharge and head gradient) increases downstream most rapidly (Shreve, 1972, 1985). Once initiated at these favored sites, the downstream trends of these observed channelways continue to follow the reconstructed head gradient almost exactly.

The field data, however, also suggest that multiple, parallel channels are more common than predicted by our reconstruction. In a few localities, this condition likely reflects the coarseness of the subglacial hydraulic reconstruction. More typically, however, this imprecision in the reconstruction is unavoidable because of either the uncertainties imposed by postglacial deposition or the unmodeled possibility of
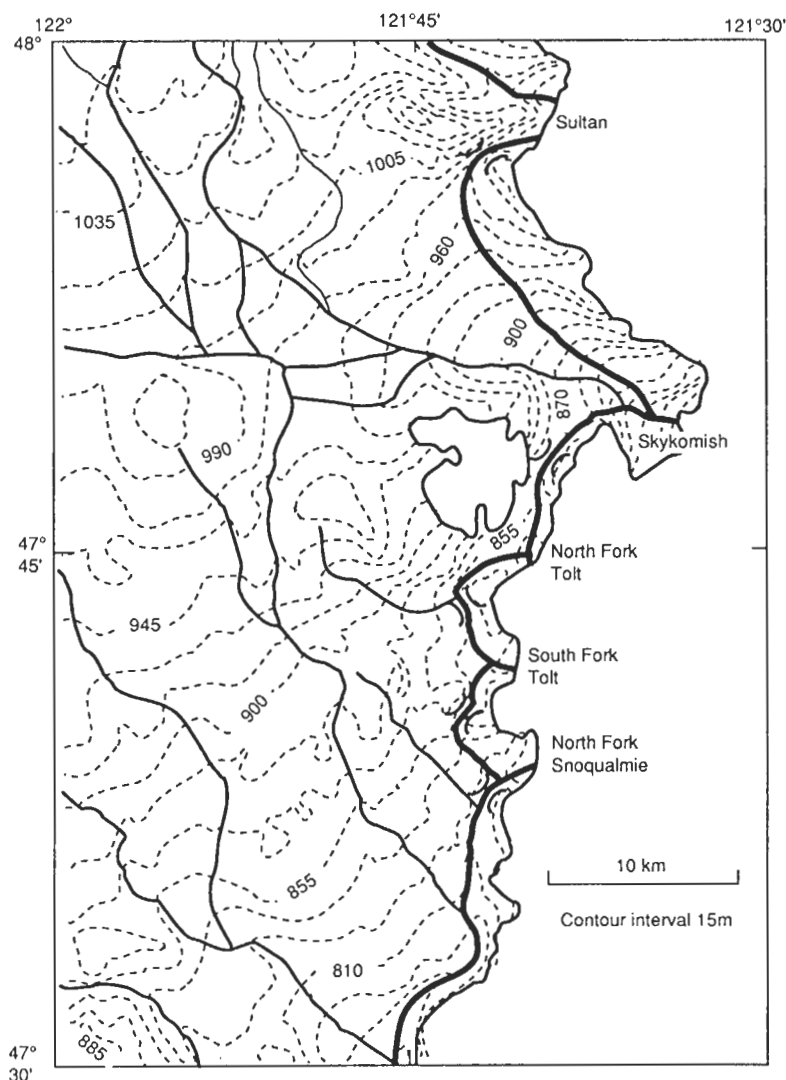
Figure 10. Contours of subglacial hydraulic head at ice-maximum stage (dashed; contour interval = 15 m) and corresponding subglacial water-flow paths. The thin lines trace paths of steady drainage; the thick line follows a drainage route along which the hydraulic head generally decreases from north to south but is interrupted by high-head saddles, indicated by) (. Marginal lakes impounded behind these saddles formed in each of the labeled valleys of the Cascade Range and drained episodically (see also Table 1).

sequential occupation of individual drainage channels due to minor changes in ice-surface gradients.

Conversely, all of the major (and almost all of the secondary) reconstructed flow paths are expressed by channelways. Even where covered by later deposits (and so mainly unmapped on Fig. 11), the valley form of partly buried channelways is recognizable along most of the corresponding routes of predicted subglacial drainage.

Submarginal Drainage and Ice-Marginal Lakes

The segmented channel near the eastern ice margin is an exceptional example of subglacial water erosion. On its west side (toward the center of the ice lobe) the channel is incised over 150 m into the glaciated upland surface; on its east side the channel lies at the base of an eroded 400-m escarpment along the margin of the Cascade Range. Early reconstruction of the Puget lobe (Mackin, 1941) suggested that an immense quantity of subaerial meltwater filled this channel during ice retreat, but our reconstruction of subglacial water flow, notably the episodic drainage of ice-marginal lakes, suggests that both the duration of fluvial erosion and the flux of water were probably much greater while the channel was active beneath the ice sheet.

The magnitude of subglacial discharges along this channel can be crudely estimated from empirical relationships between lake volume and jökulhlaup discharge, albeit extrapolated about an order of magnitude beyond modern data. Maximum discharges from the impounded ice-marginal lakes are predicted by the formula of Clague and Mathews (1973; with modification by Beget, 1986):

$$Q_{max} = 0.0065 V^{0.69} \qquad (9)$$

TABLE 1. LAKE PARAMETERS

Lake	Maximum volume (km³)	Recharge time (yr)	Maximum discharge (m³ · s⁻¹)
Sultan	31	47	1×10^5
Skykomish	120	34	3×10^5
North Fork Tolt	3.3	10	2×10^4
South Fork Tolt	4.8	27	3×10^4
North Fork Snoqualmie	13	29	6×10^4

Note: lakes are located along the submarginal channelway system of Figure 10. Maximum volumes determined at ice-maximum stage by planimetry of modern topographic maps, recharge time based on modern measured runoff rates (U S West, 1988) and subaerial drainage basins, and discharge calculated from empirical data on drainage of modern ice-dammed lakes (Clague and Mathews, 1973, as modified by Beget, 1986).

Figure 11. Comparison of observed channelways and postglacial channelized alluvium (Fig. 3) with predicted flow paths (Fig. 10). The correspondence is very good, except that multiple flow paths cannot be discriminated where recessional or postglacial deposits have obscured the original bed topography, and closely spaced parallel channelways are not well predicted by the distribution of total head due to the coarseness of the reconstruction.
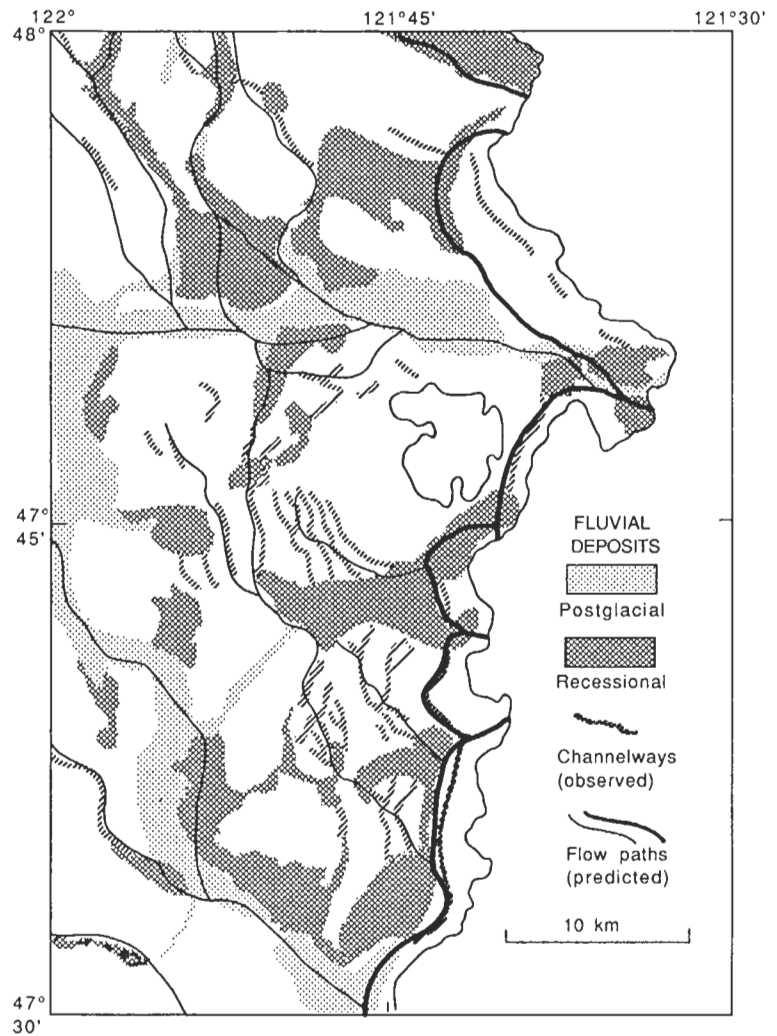
where Q_{max} = maximum jökulhlaup discharge in m^3s^{-1} and V = initial lake volume in m^3. Using a release volume of 10^{11} m^3 (see Table 1), the discharge is of order 10^5 m^3s^{-1}. For comparison, this discharge is one to two orders of magnitude less than predictions of Lake Missoula floods (Clarke and others, 1984; Beget, 1986; O'Conner and Baker, 1992), but it is five to ten times greater than computed paleodischarges in the Rocky Mountain Trench (Clague, 1975) and fully one to two orders of magnitude greater than reported from modern Alaskan ice-dammed lakes (Stone, 1963).

Deposits associated with the Puget-lobe jökulhlaups would be largely obscured by later (and finer) recessional-outwash deposits that occupied the same channel. A record of at least one and probably two discrete episodes, however, is preserved in 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. 6; Booth, 1991b).

THE REGIONAL DISTRIBUTION OF SUBGLACIAL FLUVIAL LANDFORMS

The eastern Puget Lowland provides abundant field evidence of the spatial variability and local significance of subglacial fluvial erosion. The characteristic landforms of this process—channelways,

truncated spurs, and isolated nonstreamlined hills—are common in this near-marginal region but are much sparser farther upglacier and toward the ice-sheet interior, even where bed topography and surficial deposits are otherwise favorable for their formation and preservation.

Part of this variability is due to differences in the fluxes of ice and water over the glacier bed. Longitudinally, of course, the action of sliding ice must increase from near-zero at the upglacier ice divide, reach its maximum near the equilibrium line, and vanish altogether at the glacier terminus. In contrast, the effects of subglacial water likely increase monotonically downglacier (Fig. 12). Lateral differences are also pronounced. Ice flux does not vary dramatically across broad lowland topography, but water flow typically is channeled and hence is concentrated over only a small fraction of the total glacier width (for example, Sharp, 1947; Rothlisberger, 1972; Shreve, 1972; Goldthwait, 1974; Vivian, 1975; Walder and Hallet, 1979; Brown and others, 1987).

Part of this variability also results from the influence of pre-existing bed topography. Preglacial topography, including the relict channelways of previous glaciations, will unavoidably channelize subglacial water flow. Other landforms provide the best opportunities to recognize unequivocally these water-formed channels. In this region, individual bedrock ridges lying transverse to the ice-flow direction, likely relics of preglacial topography (Newell, 1970; Rudberg, 1973; Gordon, 1981), provide such a setting. Even through multiple glaci-
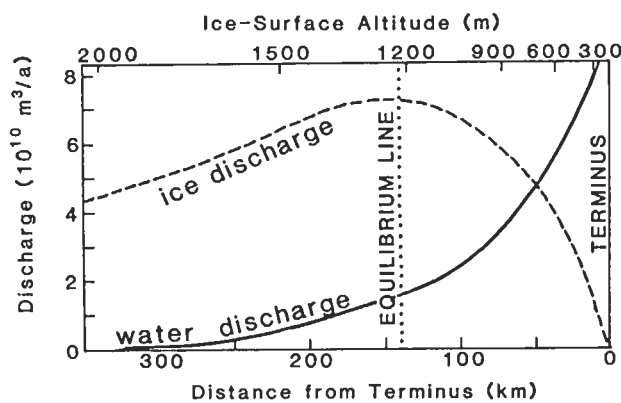
Figure 12. Total flux of ice and water across the entire Puget lobe as functions of longitudinal position. All values spatially averaged (that is, no channelization of water assumed). Equilibrium-line position and mass flux data are from Booth (1986a).

ations, these transverse landforms have persisted, and their local relief likely has been enhanced, as a result of lateral and lee-side dissection by subglacial water in the surrounding channels.

Geomorphic expression of subglacial fluvial erosion thus depends on abundant subglacial water and favorable preglacial topography, together with minimal postglacial modification. Where such water has been concentrated in regions beneath former ice sheets, notably near ablation-zone ice margins across projecting bedrock spurs, we expect the landforms of subglacial fluvial erosion to be recognizable and, in places, to dominate those of direct erosion by ice.

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