WATER FLOW THROUGH TEMPERATE GLACIERS

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Abstract. Understanding water movement through a glacier is fundamental to several critical issues in glaciology, including glacier dynamics, glacier-induced floods, and the prediction of runoff from glacierized drainage basins. To this end we have synthesized a conceptual model of water movement through a temperate glacier from the surface to the outlet stream. Processes that regulate the rate and distribution of water input at the glacier surface and that regulate water movement from the surface to the bed play important but commonly neglected roles in glacier hydrology. Where a glacier is covered by a layer of porous, permeable firn (the accumulation zone), the flux of water to the glacier interior varies slowly because the firn temporarily stores water and thereby smooths out variations in the supply rate. In the firn-free ablation zone, in contrast, the flux of water into the glacier depends directly on the rate of surface melt or rainfall and therefore varies greatly in time. Water moves from the surface to the bed through an upward branching arborescent network consisting of both steeply inclined conduits, formed by the enlargement of intergranular veins, and gently inclined conduits, spawned by water flow along the bottoms of near-surface fractures (crevasses). Englacial drainage conduits deliver water to the glacier bed at a limited number of points, probably a long distance downglacier of where water enters the glacier. Englacial conduits supplied from the accumulation zone are quasi steady state features that convey the slowly varying water flux delivered via the firn. Their size adjusts so that they are usually full of water and flow is pressurized. In contrast, water flow in englacial conduits supplied from the ablation area is pressurized only near times of peak daily flow or during rainstorms; flow is otherwise in an open-channel configuration. The subglacial drainage system typically consists of several elements that are distinct both morphologically and hydrologically. An upglacier branching, arborescent network of channels incised into the basal ice conveys water rapidly. Much of the water flux to the bed probably enters directly into the arborescent channel network, which covers only a small fraction of the glacier bed. More extensive spatially is a nonarborescent network, which commonly includes cavities (gaps between the glacier sole and bed), channels incised into the bed, and a layer of permeable sediment. The nonarborescent network conveys water slowly and is usually poorly connected to the arborescent system. The arborescent channel network largely collapses during winter but reforms in the spring as the first flush of meltwater to the bed destabilizes the cavities within the nonarborescent network. The volume of water stored by a glacier varies diurnally and seasonally. Small, temperate alpine glaciers seem to attain a maximum seasonal water storage of ~200 mm of water averaged over the area of the glacier bed, with daily fluctuations of as much as 20–30 mm. The likely storage capacity of subglacial cavities is insufficient to account for estimated stored water volumes, so most water storage may actually occur englacially. Stored water may also be released abruptly and catastrophically in the form of outburst floods.

1. INTRODUCTION

The movement of water through glaciers is important for scientific understanding and for immediate practical applications. Water in glaciers profoundly affects glacier movement by influencing the stress distribution at the glacier bed and thereby the rate at which the ice slides over the bed. This process is important for both alpine glaciers [e.g., *Iken and Bindschadler*, 1986] and polar ice streams [e.g., *Alley et al.*, 1987; *Echelmeyer and Harrison*, 1990; *Kamb*, 1991]. The episodic surging (orders-of-magnitude increase in speed) of some glaciers is evidently due to temporal changes in subglacial hydrology [*Kamb et al.*, 1985]. Glacial outburst floods, a common hazard in mountainous regions, result from the rapid release of large volumes of water stored either within a glacier or in a glacier-dammed lake [*Björnsson*, 1992; *Haeberli*, 1983; *Walder and Costa*, 1996]. Water from glaciers is becoming increasingly important for hydroelectric power generation [*Benson et al.*, 1986; *Lang and Dyer*, 1985]. Some hydropower projects in France [*Hantz and Lliboutry*, 1983], Norway [*Hooke et al.*, 1984], and Switzerland [*Bézinge*, 1981] have involved tapping water from directly beneath a glacier.

The purpose of this paper is to present a conceptual model of water flow through a glacier based on a syn-

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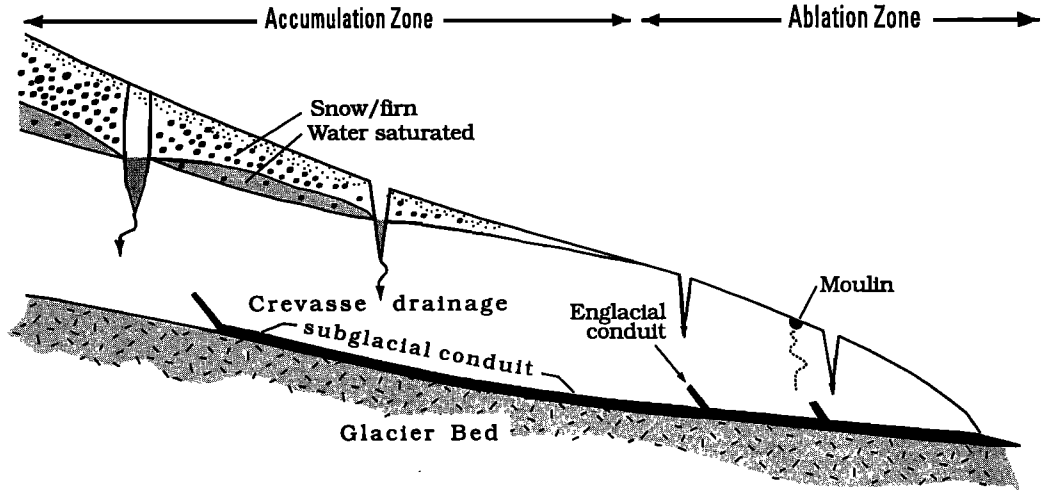

Figure 1. Idealized longitudinal cross section of a temperate alpine glacier showing the important hydrological components. In the accumulation zone, water percolates downward through snow and firn to form a perched water layer on top of the nearly impermeable ice, and then flows from the perched water layer in crevasses (open fractures). In the ablation zone, once the seasonal snow has melted, water flows directly across the glacier surface into crevasses and moulins (nearly vertical shafts). Based on Figure 10.11 of *Röthlisberger and Lang* [1987]; copyright John Wiley and Sons Ltd.; reproduced with permission.

thesis of our current understanding. We have extended the scope of previous reviews [*Röthlisberger and Lang*, 1987; *Lawson*, 1993] by focusing on ways in which the various components of the drainage system interact. As part of the conclusions, we outline subjects that need further investigation. This paper emphasizes temperate alpine glaciers (glaciers at their melting point), but results from and implications for ice sheets are included where appropriate. We do not discuss the hydrological role of the seasonal snowpack, as there is a comparative wealth of literature on the subject [e.g., *Male and Gray*, 1981; *Bales and Harrington*, 1995] and because the effect of snow on glacier hydrology has recently been reviewed [*Fountain*, 1996].

2. HYDROLOGY OF THE FIRN AND NEAR-SURFACE ICE

At the end of the melt season the surface of a glacier consists of ice at lower elevations in the ablation zone, where yearly mass loss exceeds mass gain, and snow and firn at upper elevations in the accumulation zone, where yearly mass gain exceeds mass loss (Figure 1). Firn is a transitional material in the metamorphosis of seasonal snow to glacier ice. As we will discuss in section 2.1, the presence or absence of firn has important implications for subglacial water flow and for variations in glacial runoff.

2.1. Accumulation Zone

The accumulation zone typically covers ~50–80% of an alpine glacier in equilibrium with the local climate [*Meier and Post*, 1962]. The near surface of the firn is partially water saturated. The rate of water movement through unsaturated firn depends on the firn's permeability and the degree of saturation [*Ambach et al.*, 1981], similar to percolation through unsaturated snow [*Colbeck and Anderson*, 1982] and soil [e.g., *Domenico and Schwartz*, 1990]. The near-impermeable glacier ice beneath promotes the formation of a saturated water layer at the base of the firn. Such water layers are common in temperate glaciers [*Schneider*, 1994]. The depth to water generally increases with distance upglacier [*Ambach et al.*, 1978; *Fountain*, 1989], as can be expected from the general increase in snow accumulation with elevation. High in the accumulation zone, the water table may be as much as 40 m below the glacier surface [*Lang et al.*, 1977; *Schommer*, 1977; *Fountain*, 1989].

The hydrological characteristics of firn are fairly uniform between glaciers. Field tests of the hydraulic conductivity (permeability with respect to water) of the firn at five different glaciers [*Schommer*, 1978; *Behrens et al.*, 1979; *Oerter and Moser*, 1982; *Fountain*, 1989; *Schneider*, 1994] indicate a surprisingly narrow range of $1–5 \times 10^{-5}$ m/s. This may reflect a uniform firn structure resulting from a common rate of metamorphism of firn to ice. Firn samples from South Cascade Glacier, Washington State, had a porosity of 0.08–0.25 with an average of 0.15 [*Fountain*, 1989]. This average value is equal to the value that *Oerter and Moser* [1982] found to be most appropriate for their calculations of water flow through the firn. Within the water layer, ~40% of the void space is occupied by entrapped air [*Fountain*, 1989].

The depth to the water layer depends on the rate of water input, the hydrological characteristics of the firn, and the distance between crevasses, which drain the

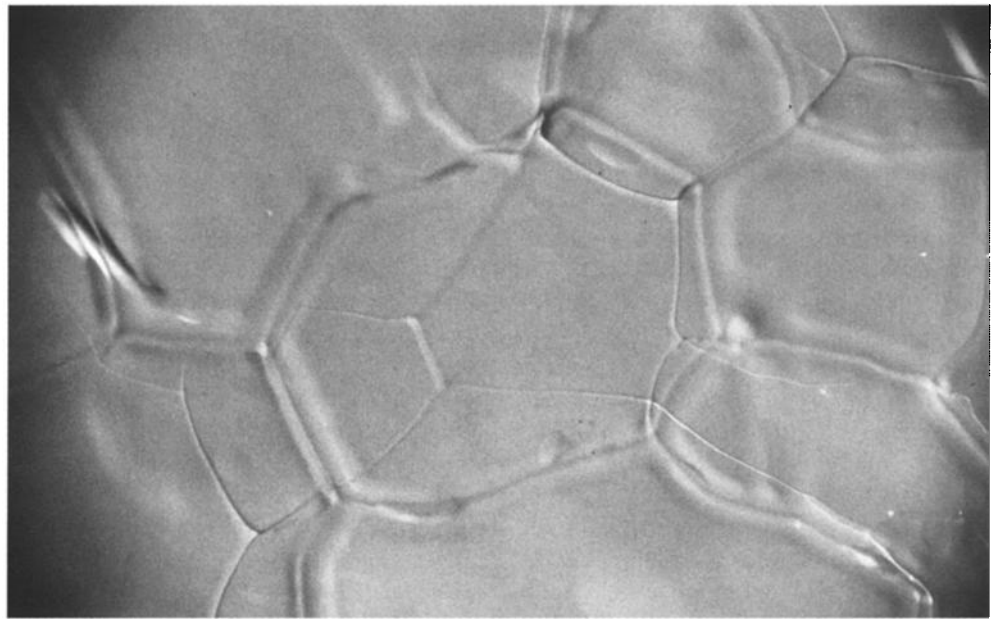

Figure 2. Photomicrograph showing the veins in glacier ice formed at junctions where three grains of ice are in contact. The grain at the center is ~1 mm across in its longest dimension. The photograph is courtesy of C. F. Raymond.

water from the firn [*Lang et al.*, 1977; *Schommer*, 1977; *Fountain*, 1989]. Over small areas (length scales of $\sim 10^2$ m) of a glacier both surface melt rate and firn permeability are relatively uniform, and the depth to water is controlled primarily by crevasse spacing. Water input to the firn varies both seasonally and daily. Seasonal variations, ranging from no water input in winter to perhaps as much as several tens of millimeters per day in summer, cause the thickness of the water layer to vary up to several meters [*Lang et al.*, 1977; *Schommer*, 1977, 1978; *Oerter and Moser*, 1982]. Typically, the water table responds to melt and precipitation within a day or two [*Oerter and Moser*, 1982; *Schommer*, 1977; *Schneider*, 1994], although diurnal variations have been occasionally observed [e.g., *Fountain*, 1989].

Nye and Frank [1973] suggested that significant quantities of meltwater may drain from the glacier surface to the bed through intergranular veins in the ice. However, observed veins are quite small (Figure 2) [see also *Raymond and Harrison*, 1975], and water passage may often be blocked by air bubbles [*Lliboutry*, 1971]; furthermore, the permeability of the ice may actually be lower near the ice surface than within the body of the glacier [*Lliboutry*, 1996]. Thus intergranular drainage is probably negligible, and water drains from the firn into crevasses that penetrate into the body of the glacier (Figure 3). From a hydrological perspective the firn is a perched, unconfined aquifer that drains into otherwise impermeable ice underneath via crevasses.

One important difference between a firn aquifer and a typical groundwater aquifer is that the thickness and extent of the firn continually change, whereas a groundwater aquifer is relatively constant over time. Permeable firn is lost as metamorphic processes transform firn to ice, closing the passages between the void spaces and rendering the matrix impermeable to water flow [*Shumskii*, 1964; *Kawashima et al.*, 1993]. At the same time, more firn is added as the seasonal snow ages and snow

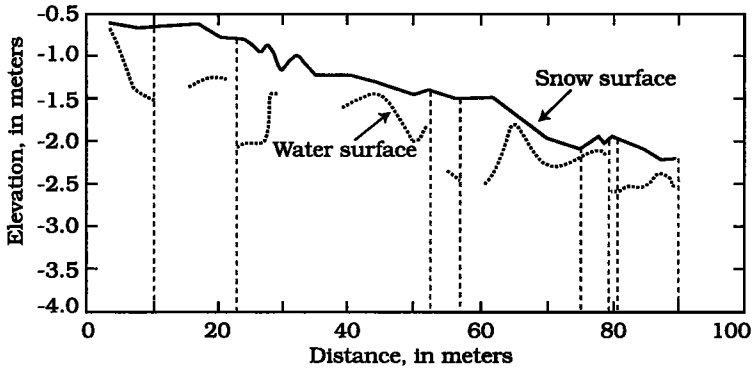

Figure 3. Profile of the depth to the firn water table on South Cascade Glacier on August 26, 1986 [after *Fountain*, 1989]. The solid curve is the snow surface, the dotted curve is the top of the perched water layer, and the dashed vertical lines represent crevasses. The datum is arbitrary. Reprinted from *Annals of Glaciology* with permission of the International Glaciological Society.

grains sinter together. This process raises the base level of the water layer relative to its former position [*Fountain*, 1989]. Firn thicknesses change from year to year depending on the residual snow thickness at the end of the summer.

The primary hydrological effects of the firn on glacier hydrology are to temporarily store water, to delay its passage to the interior of the glacier, and to smooth out diurnal variations in meltwater input. Water storage in the firn water layer delays the onset of spring runoff from glaciers and delays the cessation of flow in the autumn after surface melting has ended. For typical values of firn porosity and water saturation the water content of a perched layer 1 m thick is equivalent to that of a layer of water ~0.09 m thick. *Fountain* [1989] showed that at South Cascade Glacier the volume of water stored in the firn is equivalent to ~12% of the maximum volume of water stored seasonally by the glacier [*Tangborn et al.*, 1975]. In comparison, water storage in the firn at Storglaciären, Sweden [*Schneider*, 1994], accounted for 44% of the maximum seasonal water storage estimated by *Östling and Hooke* [1986]. Transit time through the firn depends on the speed of a wetting front in unsaturated firn, ~0.25 m/h [*Schneider*, 1994], and the response time of the saturated layer at the base. For example, if the water table is 10 m below the firn surface, the transit time to the water table is ~40 hours (longer if a seasonal snow layer is present). Transit time through the saturated water layer to crevasses depends on the distance between the crevasses and on the slope of the water surface. Considering both percolation to the firn water table and flow in the water table before exiting into a crevasse, a parcel of water commonly takes ~10–160 hours. In comparison, transit times in the body of the glacier are commonly no more than a few hours for moderate-sized temperate glaciers [e.g., *Hock and Hooke*, 1993; *Fountain*, 1993; *Nienow*, 1994]. Because the crevasses are not uniformly spaced and the thickness of the firn increases with elevation, the transit time through the firn to the interior is spatially variable with the net effect of smoothing diurnal variations in meltwater input to the glacier. We believe that water passage through the snow and firn of the accumulation zone is the source of the slowly varying component (base flow) of glacial runoff.

2.2. Ablation Zone

In the ablation zone the seasonal snowpack retains meltwater and thus retards runoff during the early part of the melt season [*Fountain*, 1996]. After the seasonal snow has melted, revealing glacier ice, channels develop on the glacier surface that drain meltwater directly into crevasses and moulins (naturally occurring vertical tunnels) [*Stenborg*, 1973]. In the absence of the seasonal snowpack the delay for rainwater and meltwater to enter the body of the glacier is brief, for example, no more than 40 min at Haut Glacier d'Arolla, Switzerland (M. J. Sharp, written communication, 1996). The presence of pools of water and surface streams in the ablation zone indicates the relative impermeability of the ice. The near-surface ice is not completely impermeable, however. Water may be transported along grain boundaries in veins, which are enlarged by solar radiation. This process is limited to the uppermost few tens of centimeters in the ice owing to the limited penetration of shortwave solar radiation [*Brandt and Warren*, 1993]. The permeability of the near-surface ice may account for small (several centimeters) fluctuations of water levels in boreholes that do not connect to a subglacial hydraulic system [*Hodge*, 1979; *Fountain*, 1994]. However, the water flux through this near-surface layer is almost certainly negligible compared with the flow in supraglacial streams.

Where the ice is moving, melted surface ice is replenished by ice emerging from the interior of the glacier [*Meier and Tangborn*, 1965], and a deeply weathered crust, from the effects of solar radiation, does not develop. In contrast, in regions of "dead ice," where the ice is not replenished, the near-surface ice can become weathered and quite permeable. Observations of water level fluctuations in boreholes in dead ice indicate a saturated water layer several meters thick [*Larson*, 1977, 1978]. On the basis of pump tests the hydraulic transmissivity T of the permeable surface layer is $\sim 8 \times 10^{-5}$ m^2/s [*Larson*, 1978]. If the perched water table thickness is $b = 2$ m, then the hydraulic conductivity $\kappa = T/b$ of the near-surface ice is $\sim 4 \times 10^{-5}$ m/s. This value is within the range given for firn, but the correspondence is probably coincidental.

In summary, the near-surface processes in the snow-free ablation zone introduce little delay in the routing of water into the body of the glacier. Moreover, the water flux into the glacier is greater in the ablation zone, compared with the accumulation zone, because the melt rate is greater owing to both the lower albedo of ice as compared with snow and the warmer air temperatures at the lower elevations. Consequently, both the mean daily flux of meltwater and the variability in the flux of meltwater are greater in the ablation zone than in the accumulation zone.

3. WATER MOVEMENT THROUGH THE BODY OF A GLACIER (ENGLACIAL HYDROLOGY)

For temperate glaciers, nearly all rain and surface meltwater enters the body of the glacier through crevasses and moulins [e.g., *Stenborg*, 1973]. As was discussed in section 2.2, the flux through the veins in the ice is probably negligible. Crevasses are the most important avenue for water because they are more numerous than moulins and are found over the entire glacier, whereas moulins are generally restricted to the ablation zone. Water-filled crevasses are not common, indicating that they efficiently route water into the body of the glacier. This conclusion is supported by *Stenborg*'s [1973] work

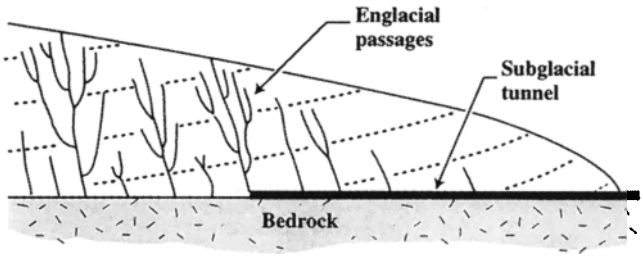

Figure 4. Fluid equipotentials (dotted curves) and a hypothetical network of arborescent englacial channels [after *Shreve*, 1985]. Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright @ 1985 Geological Society of America.

showing that moulins develop from crevasses. Neither the nature of hydraulic links between crevasses and the body of the glacier nor the formation of such links is well understood. We attempt to address these topics below.

Water flows englacially (through the body of a glacier) via ice-walled conduits. The mechanics of steady flow in englacial conduits have been described theoretically by *Röthlisberger* [1972] and *Shreve* [1972]. As with conduits at the glacier bed, discussed in section 4, englacial conduits exist if the tendency for closure, from the inward creep of ice, is balanced by the melt enlargement resulting from the energy dissipated by flowing water. *Shreve* [1972] argued that englacial conduits should form an upward branching arborescent network, with the mean flow direction oriented steeply downglacier, as determined by the gradient of the total potential (gravity and ice pressure) driving the flow (Figure 4). Empirical results based on the dispersion and travel time of numerous tracer injections in crevasses support the arborescent-network hypothesis [*Fountain*, 1993].

There are few data bearing on the distribution and geometry of englacial conduits or on englacial water pressures and flow rates. Most of our information comes from boreholes drilled to the glacier bottom using a jet of hot water [*Taylor*, 1984]. About half of all such boreholes drain before the glacier bed is reached, indicating that many boreholes intersect englacial passages [*Engelhardt*, 1978; *Hantz and Lliboutry*, 1983; *Fountain*, 1994; *Hooke and Pohjola*, 1994]. Measurements of water level, water quality, and flow direction in boreholes [e.g., *Sharp et al.*, 1993b; *Meier et al.*, 1994] and measurements of tracers injected into boreholes [e.g., *Hooke et al.*, 1988; *Hock and Hooke*, 1993] strongly suggest the presence of englacial conduits.

A number of direct measurements of englacial passages exist. *Hodge* [1976] found englacial voids with typical vertical extents of ~0.1 m, and *Raymond and Harrison* [1975] found small, arborescent, millimeter scale passages, but whether these voids or passages were part of an active hydraulic system was unclear. (Void is used here to mean a water-filled pocket in the ice, which may or may not be part of the englacial hydraulic system. Isolated voids are known to exist.) Englacial conduits have been observed where they debouch into subglacial tunnels (Figure 5) and where they intersect boreholes (Figure 6). Video cameras lowered into boreholes by *Pohjola* [1994] and by *Harper and Humphrey* [1995] revealed multiple englacial voids through nearly the entire ice thickness. Voids that intersected opposite sides of the borehole wall were interpreted as englacial conduits; typically, one or two such features were encountered in

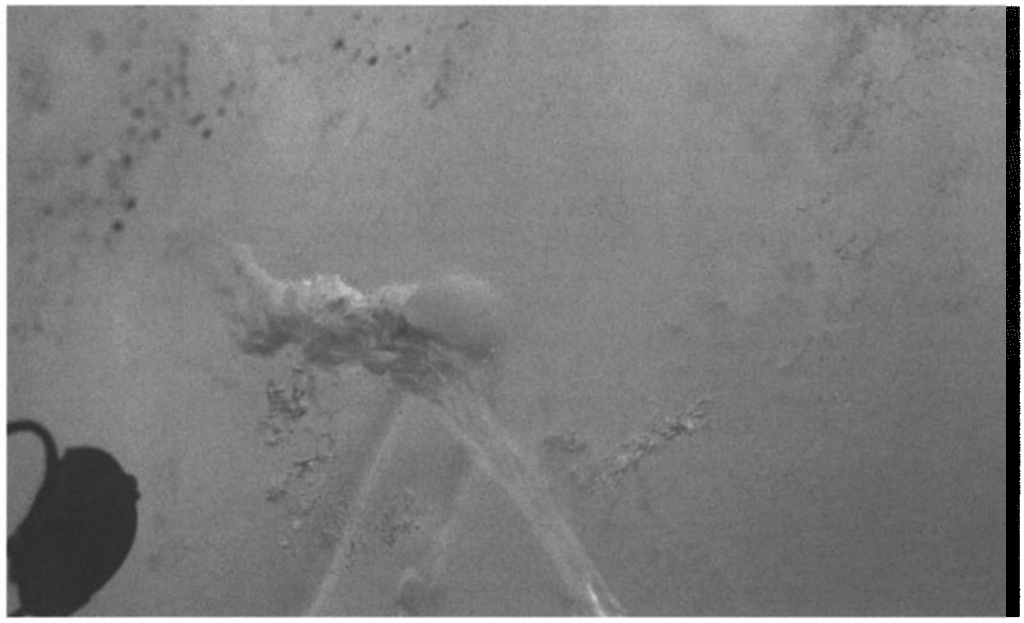

Figure 5. Water pouring from an englacial conduit at the base of South Cascade Glacier. The englacial conduit intersects a subglacial tunnel accessible from the glacier margin. Sunglasses are shown for scale.

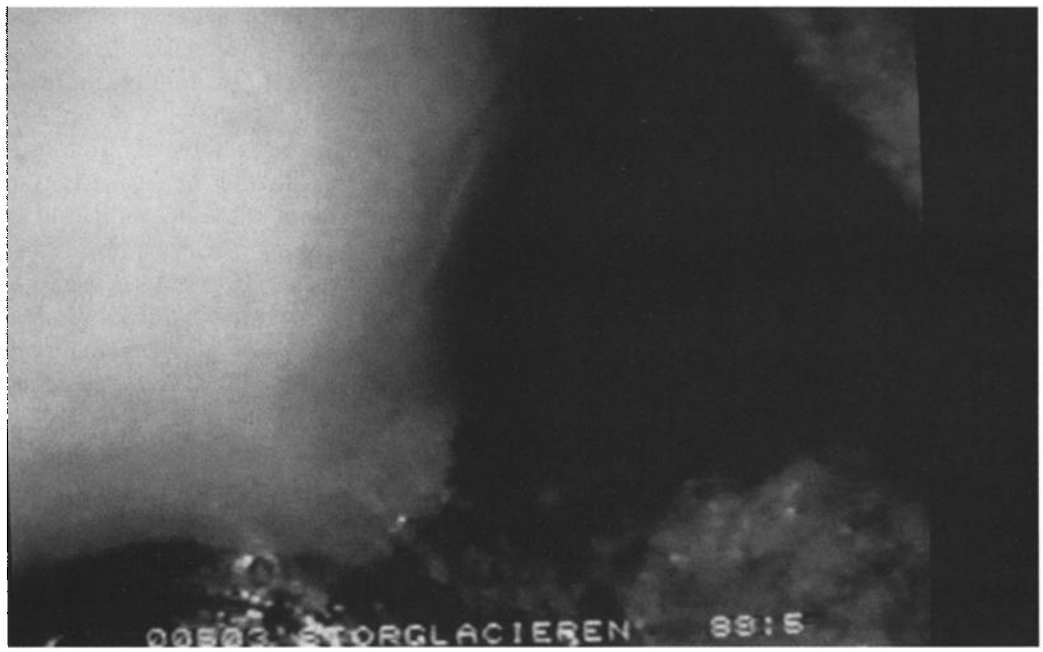

Figure 6. An englacial conduit (black triangular void at right) intersecting a vertical borehole. Note the ripples on the water surface in the borehole caused by water flowing from the conduit, which has a diameter of ~0.1 m. This video image from Storglaciären, Sweden, is courtesy of V. Pohjola.

each borehole, with diameters typically ~0.1 m. *Pohjola* [1994] determined that water was flowing in a few englacial conduits and estimated a flow speed in one of ~0.01–0.1 m/s, the same range estimated by *Hooke et al.* [1988] using dye tracers. Most conduits seen by borehole video seemed to be nearly horizontal, although Harper and Humphrey mention one plunging at an angle of ~65°. Despite obvious sampling problems, which include small sample size, biased sampling of conduit orientations from vertically oriented boreholes, and a borehole location largely restricted to the ablation zone, several conclusions can be drawn. First, glaciologists need to consider how borehole water levels and tracer injections, typically used to investigate subglacial hydraulic conditions, may be affected by englacial hydraulic connections [*Sharp et al.*, 1993b]. Second, the near-horizontal slope of many conduits needs to be reconciled with the relatively steep plunge expected from theoretical considerations [*Shreve*, 1972].

3.1. Origin of Englacial Passages

We suggest that the near-horizontal englacial conduits encountered in boreholes may originate from the action of water flowing along the bottom of crevasses. Some support for this notion comes from the borehole-video observations of *Pohjola* [1994], who found that englacial passages were usually in close proximity to bands of blue (i.e., bubble free) ice and who suggested that these bands originated from water freezing in crevasses. *Harper and Humphrey* [1995] also noticed that englacial passages and blue-ice bands tended to occur together on the walls of boreholes.

We conjecture that water flowing along the base of a crevasse either enters relatively steeply sloping, enlarged veins that form a network of arborescent passages [*Shreve*, 1972] or enters the microfractures created during crevasse formation that connect to other crevasses. Water flowing along the base of the crevasse will melt the walls and tend to widen and deepen the crevasse (Figure 7a). Melting occurs where water is in contact with the ice, such that the crevasse surface in contact with the flowing water the longest, the crevasse bottom, will melt the most. This scenario should also apply to water flowing along the glacier margin (Figure 7b). Typically, water from snowmelt and rainfall on adjacent slopes flows under the edge of a glacier, exploiting gaps between the thinner ice along the glacier margin and the bedrock. Water quickly descends until the gaps no longer connect and the water forms a stream at the ice-bedrock interface. The stream melts the ice, causing the channel to descend along the sloping bedrock.

The rate of downcutting by a crevasse-bottom stream may be crudely estimated if we assume that the cross section of the crevasse bottom maintains a semicircular geometry of radius R (Figure 8). Thus downcutting proceeds without widening. The local melt rate normal to the channel wall is then $\dot{d}\cos\theta$, where $\dot{d}$ is the melt rate at the bottom, that is, the downcutting rate. The average melt rate $\langle\dot{m}\rangle$ normal to the channel surface is then given by

$$\langle\dot{m}\rangle = \frac{1}{\pi}\int_{-\pi/2}^{\pi/2} \dot{d}\cos\theta\, d\theta = \left(\frac{2}{\pi}\right)\dot{d} \qquad (1)$$

Further assuming that all energy dissipated by the flowing water actually causes melting, we find

$$\langle \dot{m} \rangle = \frac{Q\rho_w g S}{\pi \rho_i h_{iw} R} \tag{2}$$

where Q is the water flux, ρ_w is the density of water, ρ_i is the density of ice, g is the acceleration due to gravity, S is the slope of the crevasse bottom, and h_{iw} is the latent heat of melting. The hydraulics are reasonably described by the empirical Manning equation, which we write as

$$Q = \frac{\pi}{2\tilde{n}} R^{8/3} S^{1/2} \tag{3}$$

where $\tilde{n}$ is the roughness. Combining (1) through (3), we can relate $\dot{d}$ to Q and S by

$$\dot{d} = \frac{1}{2}\left(\frac{\pi}{2\tilde{n}}\right)^{3/8}\left(\frac{\rho_w}{\rho_i}\right)\left(\frac{g}{h_{iw}}\right) S^{19/16} Q^{5/8} \tag{4}$$

Using $\rho_w = 10^3$ kg/m^3, $\rho_i = 9 \times 10^2$ kg/m^3, $g = 9.8$ m/s^2, $h_{iw} = 3.35 \times 10^5$ J/kg, and $\tilde{n} = 0.01$ s/m$^{1/3}$ (appropriate for a smooth conduit), we have calculated values of $\dot{d}$ for the case $S = 0.1$ (Table 1). This value of S is appropriate for crevasses trending longitudinally along the glacier and therefore represents a plausible upper bound.

The values in Table 1 indicate that even very modest flow rates can cause crevasse-bottom streams to cut down at a rate of a few to a few tens of meters per year. (Note that to give downcutting rates relative to the glacier surface, tabulated rates must be corrected for the vertical component of ice velocity.) These rates are many orders of magnitude greater than the rate at which a water-filled crevasse deepens owing purely to the weight of the water and the fracture-mechanical properties of the ice [*Weertman*, 1971]. Furthermore, the downcutting rate increases with Q, suggesting a possible mechanism for stream capture; in regions of complex, crosscutting crevasses those hosting relatively large water fluxes may cut down through others hosting small fluxes.

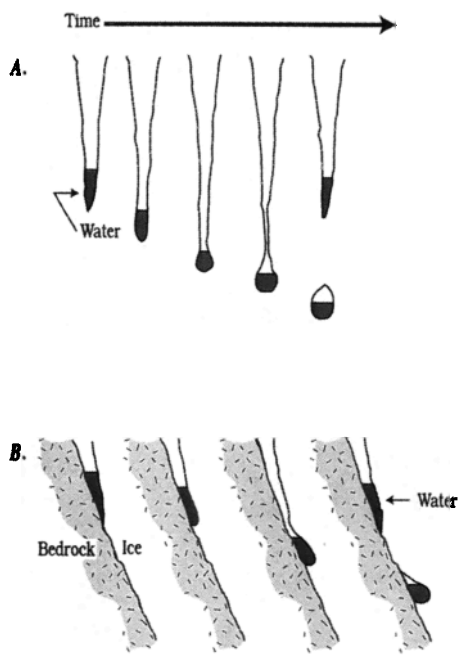

Figure 7. Qualitative depiction of the evolution of an englacial conduit from its origin (a) as a crevasse-bottom stream or (b) as a glacier-margin stream. The stream cuts down because melting occurs only where water is in contact with ice.

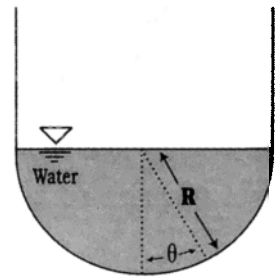

Figure 8. Idealized cross-sectional geometry of a stream flowing at the base of a crevasse. The rate at which the stream melts its way downward into the ice is calculated assuming that the crevasse "tip" is semicircular and that the stream cuts vertically downward.

We envisage the following scenario for the evolution of a crevasse-bottom stream: As the stream cuts down into the glacier, the creep of ice tends to pinch off the crevasse above the flowing water. Eventually, the channel becomes isolated from the crevasse except for rare, near-vertical passages that convey water from the crevasse to the channel (Figure 7a). The stream ceases to descend once the rate of closure equals the rate of melting at the bottom of the channel (downcutting) because it becomes fully enclosed in ice and melting occurs equally on all walls.

To estimate the ultimate depth to which a crevasse-bottom stream can cut down, we idealize the "mature" channel as having a circular cross section of radius R and assume that the melt rate is exactly balanced by the rate of ice creep. The channel will be flowing full but with zero pressure head (atmospheric pressure), like the "gradient conduit" discussed by *Röthlisberger* [1972]. Using *Nye*'s [1953] result for conduit closure, the balance between melting and closure is given by

$$\frac{\rho_w g Q S}{2\pi \rho_i h_{iw} R} - u_e = R\left(\frac{p_i}{nA}\right)^n \tag{5}$$

where n and A are flow-law constants [*Nye*, 1953], u_e is the ice emergence velocity, i.e., the vertical component of ice velocity, measured positive upwards, and p_i is the ice pressure at the depth, D_{MAX} at which the channel ceases to downcut. We take $p_i = \rho_i g D_{MAX}$ and rearrange (5) to write

$$D_{MAX} = \frac{nA}{\rho_i g}\left[\frac{\rho_w g Q S}{2\pi \rho_i h_{iw} R^2} - \frac{u_e}{R}\right]^{1/n} \tag{6}$$

Using (3) to eliminate R,

TABLE 1. Calculated Values of the Rate of Downcutting and Equilibrium Depth for Englacial Conduits for Several Values of Vertical Ice Velocity u_e

Water Flux, 10^{-3} m^3/s	*Rate of Channel Incision, m/yr*	*Equilibrium Channel Depth D_{MAX}, m*			
		$u_e = -1$ *m/yr*	$u_e = 0$	$u_e = 1$ *m/yr*	$u_e = 5$ *m/yr*
1	2.9	268	210	...	...
5	8.1	268	240	204	...
10	12	275	255	231	...
50	34	300	291	283	238
100	52	317	308	304	276

The value u_e is positive in the ablation zone. The tabulated downcutting rate does not take into account ice creep and therefore applies strictly only near the glacier surface.

$$D_{\mathrm{MAX}} = \left(\frac{nA}{\rho_i g}\right)\left[\left(\frac{\rho_w g}{2\pi\rho_i h_{tw}}\right)\left(\frac{\pi}{2\tilde{n}}\right)^{3/4} Q^{1/4} S^{11/8} - \left(\frac{\pi}{2\tilde{n}}\right)^{3/8} u_e S^{3/16} Q^{-3/8}\right]^{1/n} \quad (7)$$

The "equilibrium" depth D_{MAX} is weakly dependent on Q, although, of course, the time required for the channel to cut to this depth decreases as Q increases. For very small discharges the apparent value of D_{MAX} is negative. Physically, this means that a very small crevasse-bottom stream cannot cut down and simply remain at the base of the crevasse, the depth of which is governed by the rheological properties of ice [*Paterson*, 1994]. Calculated values of D_{MAX} given in Table 1 are quite large, perhaps reflecting an overestimate of A (which decreases as the water content of the ice increases [*Lliboutry*, 1983], underestimate of $\tilde{n}$, overestimate of S, or some effect of a possibly noncircular channel cross section [cf. *Hooke*, 1984; *Hooke and Pohjola*, 1994].

In arriving at the expression in (7) for D_{MAX} we assumed steady flow in the channel. However, for an alpine glacier a constant or slowly varying discharge cannot exist unless there is a storage mechanism that can maintain a supply of water. We conjecture that channels supplied with water from the ablation zone may be able to descend deeper than those in the accumulation zone because of the large daily variability in water flux. For channels supplied from the accumulation zone they come closer to the ideal case (equation (7)) because the firn filters out daily water fluctuations and provides a water storage reservoir, reducing longer-term variations. In neither case, however, do the channels reach a true equilibrium position because of the variations in water supply.

3.2. Synopsis and Implications

Our view of the englacial drainage system is shown in Figure 9. Subhorizontal channels, spawned by the water flow in crevasse bottoms, are connected by either steeply plunging passages, formed by the enlargement of intergranular passages [*Shreve*, 1972], or microfractures between crevasses. Marginal channels also form under the edge of the glacier where water collects from the valley walls; these channels may eventually connect to channels spawned from crevasses. We infer that surface water reaching the bed is generally shifted downglacier so that it extends the influence of the firn over a larger subglacial area than it covers at the surface. This view is supported by the observations of *Iken and Bindschadler* [1986] at Findelengletscher, Switzerland, where glacier-surface streams, which exhibited marked diurnal variations, flowed into crevasses near boreholes drilled to the bottom of the glacier. Subglacial water pressure variations, as indicated by fluctuations in water level, did not correlate with the diurnal streamflow variations but rather correlated with the slower variation of meltwater input from a snow cover upglacier. Lateral shifts in surface to bed water routing are supported by the results of an experiment whereby a tracer injected at the bottom of a borehole drained to one outlet stream while a tracer injected into a crevasse adjacent to the borehole appeared in a different outlet stream [*Fountain*, 1993].

The englacial hydraulic system supplied from the ablation zone should develop more quickly and to a much greater extent than that supplied from the accumulation zone. As the snow line moves up the glacier

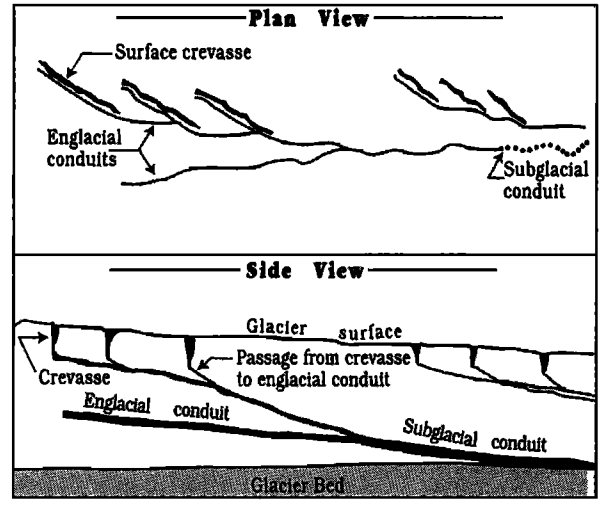

Figure 9. Geometry of hypothetical englacial drainage system, comprising both gently plunging conduits spawned by crevasse-bottom streams and steeply plunging "Shrevian" conduits formed where water exploits veins in the ice.

during the summer, ice is exposed in the ablation zone. The combination of lower albedo for ice, compared with snow, and warmer air temperatures lower on the glacier increases the water flux into the ablation zone compared to the accumulation zone. We therefore expect englacial conduits receiving water from the ablation zone to be more developed compared to the conduits receiving water from the accumulation zone. The common occurrence of moulins in the ablation zone rather than in the accumulation zone probably reflects the difference in drainage development between the two zones. The englacial drainage system must be highly dynamic, with channels being continuously reoriented by differential shear as ice is advected downstream or severed in icefalls. Channel segments must frequently close off because their water supply is lost to other channels by drainage capture or because their connection to the glacier surface is interrupted neither by refreezing during the winter or by ice creep.

Table 1 indicates that for ice thicknesses of 200 m or less, descending englacial channels may reach the bottom of the glacier to become subglacial conduits. This process provides a mechanism to route water from the glacier surface to the bed and a process by which new subglacial conduits are formed. We do not expect the englacial conduits to descend much below 300 m except in unusual circumstances; therefore subglacial conduits found below 300 m are probably formed from some other mechanism.

Our conception of the englacial drainage system has significant implications for overdeepened regions of alpine glaciers. An overdeepening is a topographic feature that would form a closed depression, and likely host a lake, if the glacier were removed [e.g., *Hooke*, 1991]. Three borehole studies at Glacier d'Argentière, France [*Hantz and Lliboutry*, 1983], Storglaciären [*Hooke and Pohjola*, 1994], and South Cascade Glacier [*Hodge*, 1976, 1979; *Fountain*, 1994] have involved drilling to the glacier bed in overdeepened areas. In all three cases the drilling was done in the ablation zone, and certain qualitative features were common:

1. Many boreholes encountered englacial conduits.
2. Once boreholes reached the bed, water levels remained close to the ice overburden pressure and did not exhibit diurnal fluctuations; thus there was no indication of low-pressure conduits within the overdeepening.
3. Low-pressure conduits seemed to exist near the valley sides.

These observations seem to contradict our conclusion that englacial conduits in the ablation zone should normally be efficient at conducting water to the bed. A resolution of this apparent contradiction is possible, however, if we consider the peculiar effect of the overdeepening on the englacial conduits.

Figure 10 shows a section of a subhorizontal conduit, one that has developed from a crevasse-bottom stream, that has cut down to the lip of an overdeepening. The conduit is now "pinned" at the downstream end and will evolve into a conduit approximately paralleling the ice surface. In that configuration the conduit is full of water, so melting occurs equally on all surfaces, and downcutting ceases. This conclusion holds for a single continuous conduit or for a network of conduits. We therefore suggest that in overdeepened areas the movement of water from the glacier surface to the bed will be restricted, as will the development of conduits at the glacier bed. Subglacial conduits can be developed from water percolating under the glacier margins, but such conduits will also be pinned by the lip of the overdeepening. We expect that subglacial conduits will be most commonly observed to pass around overdeepenings near the valley walls, as was suggested by *Lliboutry* [1983].

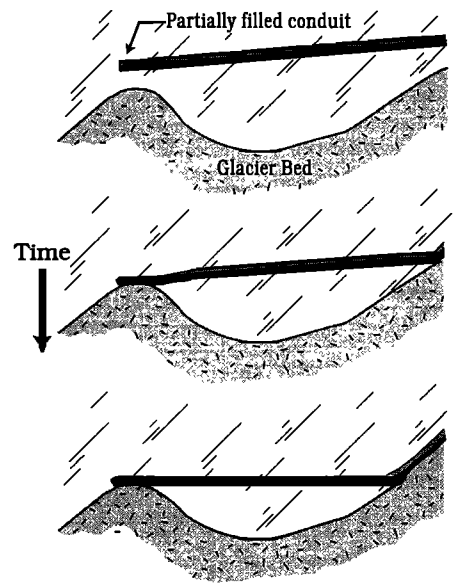

Figure 10. Longitudinal cross section of a glacier showing a hypothetical englacial conduit descending into an overdeepened region of a glacier. The conduit becomes pinned once it reaches the bed at the downstream margin of the overdeepening. Deepening upstream of this point continues until the channel becomes water filled.

4. SUBGLACIAL HYDROLOGY

The modern study of subglacial hydrology can plausibly be traced to *Mathews* [1964], who measured water pressure at the end of a shaft that reached the base of South Leduc Glacier, Canada, from a mine beneath the glacier. Mathews observed that water pressure was generally higher in winter than in summer, a situation that turns out to be common, and that abrupt increases in water pressure were correlated with periods of rapid ablation or heavy rain, reflecting the efficient hydraulic connections between the glacier surface and bed. These general conclusions were supported by investigators who reached the bed of Gornergletscher, Switzerland [*Béz-*

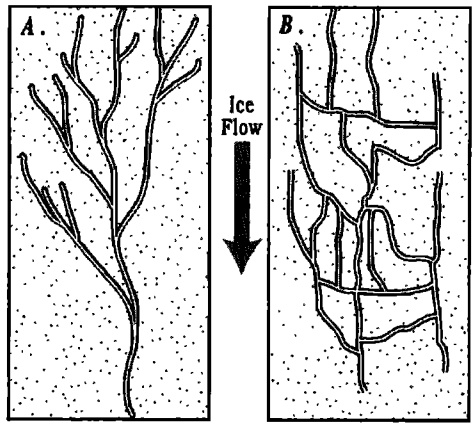

Figure 11. Idealized plan view of (a) an arborescent hydraulic (fast) system composed of channels and (b) a nonarborescent hydraulic (slow) system.

inge et al., 1973] and Glacier d'Argentière [*Vivian and Zumstein*, 1973] in connection with hydropower developments. *Vivian and Zumstein* [1973] also showed that the strong diurnal pressure fluctuations observed during the melt season were absent during winter.

Iken [1972] measured water pressure in moulins in the subpolar White Glacier and observed large diurnal variations. Subsequently, various investigators developed techniques to drill to the glacier bed and used the water level in boreholes as a piezometric measure of subglacial water pressure. The first systematic borehole studies of the subglacial drainage system were done at South Cascade Glacier [*Hodge*, 1976, 1979], Blue Glacier, Washington State, United States [*Engelhardt*, 1978; *Engelhardt et al.*, 1978], and several Swiss glaciers [*Röthlisberger et al.*, 1979]. In a minority of the boreholes in all of these glaciers, the water level dropped as soon as the drill reached the glacier bed. Water levels in these boreholes fluctuated diurnally but were usually not closely correlated. In a majority of the boreholes the water level remained high and nearly constant, commonly at a level corresponding to a water pressure greater than local ice overburden pressure, even after the drill reached the bed; a drop in the water level, if it occurred at all, was delayed for several days to a few weeks. Such experiences with borehole water levels seem to be ubiquitous. Glaciologists commonly describe the first sort of borehole as "connected" to the subglacial drainage system and, assuming that the borehole volume is small compared with the volume of accessible subglacial water, treat borehole water level as a manometric measure of water pressure at the bed. The second sort of borehole is termed "unconnected" and cannot be used as a manometer, as the hole volume is probably large compared with the volume of accessible subglacial water. Recently, *Waddington and Clarke* [1995] and *Murray and Clarke* [1995] have shown that valuable information can be collected from unconnected boreholes when the top of the borehole can be sealed, in their case, by freezing, and water pressure is measured directly at the bed, without relying on the manometric principle.

4.1. Components of the Subglacial Drainage System

Water emerges at the glacier terminus in a small number of conduits incised into the basal ice, and it is tempting to suppose that these conditions prevail subglacially as well. Reality is probably much more complicated. There is presently broad agreement among glaciologists that water flows at the glacier bed in one or both of two qualitatively different flow systems (Figure 11), commonly termed "channelized" and "distributed." This terminology is problematic because, as we shall point out, the distributed system often consists in part of what common sense dictates be called channels. We suggest that it makes more sense to refer to "fast" and "slow" drainage systems [*Raymond et al.*, 1995]. In the fast system, relatively small changes in total system volume produce relatively large changes in discharge. The fast system has a relatively low surface-to-volume ratio, covers a very small fraction of the glacier bed, and comprises an arborescent (converging flow) network of conduits, similar to a subaerial stream network. In the slow system, in contrast, relatively large changes in total system volume produce only small changes in discharge. The slow system has a relatively large surface-to-volume ratio, covers a relatively large fraction of the glacier bed, is nonarborescent, and may involve a variety of complicated flow paths at the glacier bed.

The distinction between fast and slow flow systems has been inferred from variations in borehole water levels, from the travel time and dispersion of tracers injected into glaciers, and from measurements of water flux and chemistry in streams flowing from glaciers. Under any particular glacier, part of the bed may host a fast drainage system while the rest hosts a slow drainage system, with transitional zones linking the two. Furthermore, the basal drainage system in any particular region may switch from one configuration to the other in response to perturbations in meltwater input.

4.1.1. Fast drainage system (Röthlisberger channels). An isolated, water-filled void in a glacier will tend to be closed by inward ice flow unless the water pressure p_w equals the ice overburden pressure p_i [*Nye*, 1953]. Englacial or subglacial channels may exist with $p_w < p_i$ if the flowing water dissipates enough energy as heat to melt the ice and thereby keep the channel open (Figure 12a). *Röthlisberger* [1972] presented the first reasonably complete analysis of the hydraulics and thermodynamics of steady flow in a subglacial channel, and glaciologists now commonly refer to such channels as "Röthlisberger" or "R" channels. Röthlisberger assumed that subglacial channels have a semicircular cross section and that the flowing water must gain or lose energy so as to remain at the pressure-melting temperature. He derived the following differential equation for steady flow in a channel:

$$\left(\frac{d\Phi}{ds}\right)^{p+1} - \alpha\left(\frac{d\Phi}{ds}\right)^{p}\frac{dp_w}{ds} = \beta Q^{-q}p_e^n \qquad (8)$$

where (Figure 12b) the coordinate s increases in an upglacier direction along the water flow path, z is the bed elevation relative to an arbitrary datum, $\Phi = p_w + \rho_w g z$ is the total hydraulic potential, Q is the discharge, ρ_w is the density of water, g is the acceleration due to gravity, and $p_e = p_i - p_w$ is the effective pressure in the channel. The term multiplied by α reflects the pressure-melting effect, and β involves channel roughness and ice rheology. The exponents p and q are both positive and depend weakly on the empiricism chosen to describe turbulent flow in the channel [*Lliboutry*, 1983]. The exponent $n \approx 3$ follows from empirical ice rheology.

Röthlisberger [1972] showed that the form of (8) immediately leads to an important conclusion, most easily seen for the case in which the local channel inclination ($\theta = \sin^{-1}(dz/ds)$) equals 0, in which case, $\Phi = p_w$ and (8) becomes

$$\frac{dp_w}{ds} = \left(\frac{\beta}{1-\alpha}\right)^{1/(1+p)} Q^{-q/(1+p)} p_e^{n/(1+p)} \qquad (9)$$

The greater the discharge, the smaller the pressure gradient; accordingly, if we envisage two nearby channels and integrate (9) over some finite distance x, the channel carrying the greater discharge will be at a lower pressure than the other. Röthlisberger concluded that if hydraulic connections between channels exist, the largest channels should capture the drainage of the smaller ones and an arborescent drainage network should develop. *Shreve* [1972] reached the same conclusion by a slightly different line of reasoning.

Röthlisberger's [1972] analysis applies strictly only for steady flow. This is certainly a poor approximation for channels fed from the ablation area, in which case, temporal variations of discharge are commonly large, and channels probably flow full of water only a small fraction of the time. The steady flow approximation is probably reasonable for channels fed from the accumulation area, but even in this case, channels may not be full of water. *Weertman* [1972] and *Hooke* [1984] both assessed theoretically the likely extent of open-channel flow in semicircular channels and predicted that open-channel flow under a constant discharge will occur if the ice thickness H_i is less than a critical value $H_{crit} = \text{const}\ Q^a(d\Phi/ds)^b$, with $a \approx 0.07$–0.08 and $b \approx 0.46$. Hooke concluded that open-channel flow should be common beneath steeply sloping alpine glaciers less than a few hundred meters thick. Data to assess this conclusion are sparse. *Fountain* [1993] and *Kohler* [1995] inferred from analysis of tracer injections that while open-channel flow probably did occur, it was restricted to very thin ice near the glacier terminus (i.e., in the ablation area) and thus not nearly as extensive as Hooke's calculations would have predicted.

Application of (8) to predict measured subglacial water pressures (as indicated by the water level in boreholes drilled to the bed) has been problematic. Predicted values of p_w are commonly much less than measured values [*Engelhardt*, 1978; *Hooke et al.*, 1990; *Fountain*, 1994]. *Lliboutry* [1983] argued that this primarily reflected an inadequate description of the channel-closure physics and that the correction entailed considering the complicated stress state created by glacier sliding past bedrock obstacles, as well as a proper choice of ice rheology parameters. *Hooke et al.* [1990] suggested that the discrepancy between measured and predicted p_w could be resolved if R channels are actually broad, and they proposed an ad hoc modification to *Röthlisberger*'s [1972] analysis in line with their idea about channel shape. Predicted water pressures for the modified channel shape were in good agreement with the observed borehole water pressures at Storglaciären. Finite element modeling of R channel evolution in response to variable water input [*Cutler*, 1998] supports Hooke et al.'s conjecture about channel shape.

In considering the discrepancy between predicted and measured p_w it is worth reflecting on whether borehole water level should even be considered as a piezometric

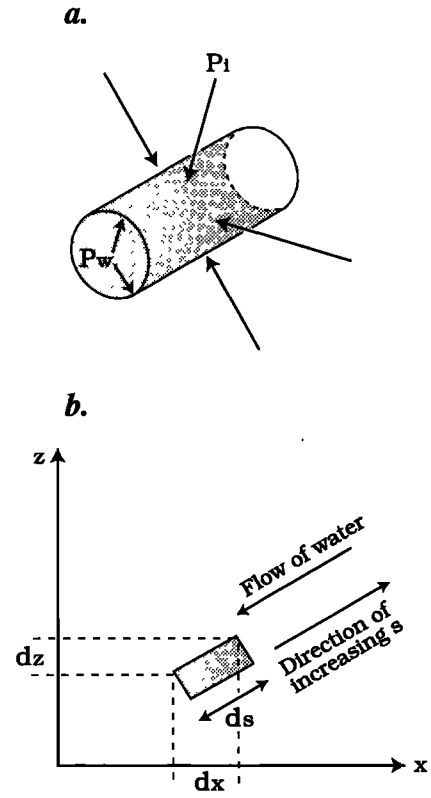

Figure 12. Schematic of a Röthlisberger channel [after *Röthlisberger*, 1972]. (a) The channel tends to close by creep of ice at overburden pressure p_i but tends to be opened by melting as the flowing water (pressure p_w) dissipates energy. (b) In general, the channel is inclined at some angle to the horizontal (the x axis). Reprinted from the *Journal of Glaciology* with permission of the International Glaciological Society.

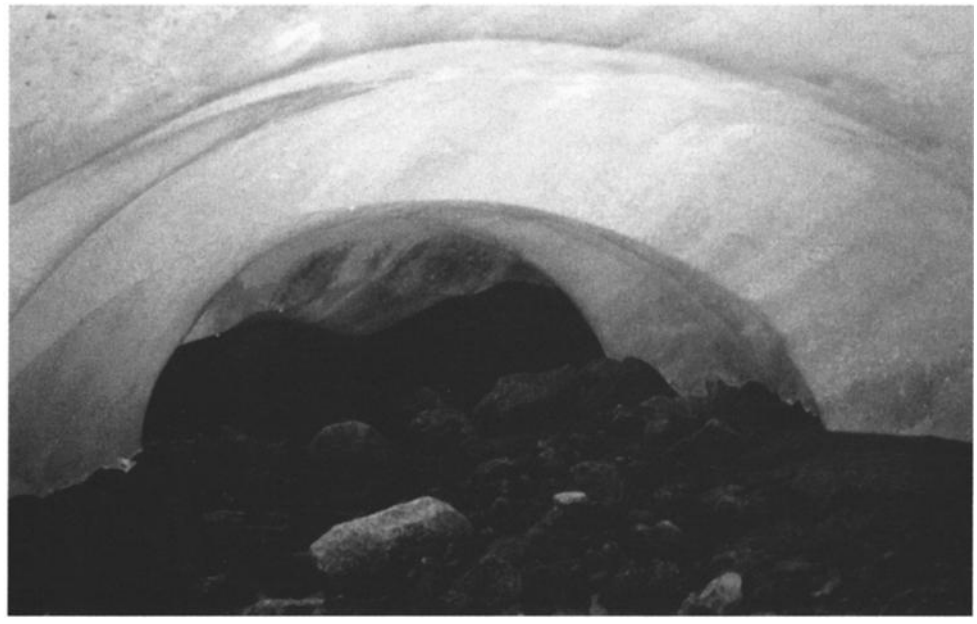

Figure 13. Subglacial conduit incised into ice near the margin of South Cascade Glacier. The height of the conduit is ~1.5 m. Note three important features: The ice is resting on unconsolidated sediments, the channel is not full of water, and the cross-sectional shape is flattened instead of semicircular.

measurement of water pressure in an R channel. Borehole water level does not reflect basal water pressure if water enters the borehole either at the glacier surface or englacially. Surface runoff can be diverted from a borehole, but the same is not true for water entering englacially, and there is a substantial body of evidence that boreholes do commonly intersect englacial channels during drilling. Moreover, although R channels have certainly been seen at the margins of glaciers (Figure 13), there is no unequivocal evidence that a borehole has ever intersected an R channel. If a borehole intersects a part of the basal hydraulic system that drains to a subglacial channel, then borehole water levels will necessarily reflect p_w greater than that in the channel [cf. *Engelhardt*, 1978]. Until a borehole can be drilled that unambiguously intersects a subglacial channel and that is not adversely affected by englacial water input, it is perhaps premature to discount the quantitative accuracy of (8). Several investigators seem to have drilled boreholes that came tantalizingly close to basal channels. Water level in borehole U of *Engelhardt et al.* [1978] fell to the glacier bed a few days after the borehole was completed, and a sounding float lowered into the borehole did not stop at the bottom of the borehole but ran out along some sort of basal passage as far as it was allowed to go. *Hantz and Lliboutry* [1983], *Fountain* [1994], and *Hubbard et al.* [1995] found that in a few boreholes, there were large diurnal pressure fluctuations, with minimum values of p_w close to atmospheric pressure, and concluded that these boreholes were efficiently connected to R channels.

Studies by *Fountain* [1994] and by *Hubbard et al.* [1995] at two different glaciers yielded intriguingly similar results bearing on the hydraulic connection between subglacial channels and the surrounding, presumably slow basal drainage system. At both glaciers, water level measurements in arrays of boreholes in the ablation area indicated the existence of a zone, elongated along the ice flow direction but only a few tens of meters wide, in which basal water pressure fluctuated greatly and commonly fell to atmospheric values. To either side of this zone, basal water pressure was generally high and fluctuated relatively little. A plausible interpretation is that an R channel existed and was efficiently connected to the ablation-zone input, whereas the adjacent (slow) hydraulic system was poorly connected to the channel and was fed from farther upglacier. We will elaborate on this idea in our discussion of the temporal evolution of the basal drainage system in section 5.

4.1.2. Slow drainage system. The slow drainage system comprises several morphologically distinct flow pathways. Most of the discharge in the slow system moves through cavities and subglacial sediment. A widespread, thin water film also forms part of the slow system; this film accommodates little water flux but may affect the glacier sliding speed as well as water chemistry [*Hallet*, 1976, 1979].

4.1.2.1. Subglacial cavities: A subglacial cavity forms where sliding ice separates from the glacier bed (Figure 14). Large cavities beneath thin ice are sometimes accessible from the glacier margin [e.g., *Anderson et al.*, 1982]. Cavity formation is favored by rapid sliding and high bed roughness [*Nye*, 1970]. *Lliboutry* [1965, 1968] was the first to propose that cavitation played a critical role in glacier sliding. In later papers [*Lliboutry*, 1976, 1978, 1979] he argued that there were two types of

cavities: "autonomous" cavities containing stagnant meltwater hydraulically isolated from the subglacial drainage system and "interconnected" cavities linked to R channels. *Lliboutry* [1976] also proposed that sliding speed should depend on the effective pressure p_e in a network of R channels and interconnected cavities.

Walder and Hallet [1979], *Hallet and Anderson* [1980], and *Sharp et al.* [1989] mapped small-scale geomorphic features on recently deglaciated carbonate bedrock surfaces and concluded that widespread cavitation must have occurred beneath some small alpine glaciers. Steep concavities downglacier of local bedrock knobs or ledges were often deeply scalloped and similar in appearance to the surfaces created by turbulent water flow over limestone in caves and subaerial environments. Because these concavities, which covered 20–50% of the deglaciated surfaces, could not hold water subaerially, they could have experienced extensive dissolution only if water had been confined over them, as in subglacial cavities. In every one of the mapped examples the overall basal drainage system was nonarborescent.

The hydraulics of steady flow through a cavity system was investigated theoretically by *Walder* [1986] and more completely by *Kamb* [1987]. A subglacial cavity opens as ice separates from the bed at the upglacier margin of the cavity and closes by the creep of ice into the cavity. Energy dissipated by flowing water also enlarges the cavity, but this effect is minor except in the orifices that link large cavities because nearly all of the head loss occurs in the orifices. Analyses by Walder and Kamb lead to the result

$$Q \sim u_b^m (d\Phi/ds)^{1/2} p_e^{-3} \qquad (10)$$

where u_b is the sliding speed and $m = 0.5$–1. The key feature of (10) is that water flux increases as p_e falls, that is, as p_w rises. Thus there is no tendency for many smaller cavities to drain into fewer, larger cavities, contrary to the situation with R channels. Another key feature of a cavity drainage system, shown particularly clearly by *Kamb* [1987], is that for a given discharge the water pressure in the cavity system must be much greater than in an R channel system.

Considering again Figure 14, it seems apparent that an arborescent R channel network should be much more efficient at evacuating meltwater than a nonarborescent cavity network. The channel network has shorter average flow paths, thus shorter travel times, than the cavity network. We also expect the behavior of tracers injected into the subglacial drainage system to be very different for the two cases: Tracers injected into a cavity network should tend to become highly dispersed, with multiple concentration peaks resulting from comparatively long travel times and multiple flow paths, whereas in a channelized system the travel times should be shorter, and dispersion should be much less. Field data such as those from Variegated Glacier, Alaska [*Brugman*, 1986; *Humphrey et al.*, 1986], support this conclusion.

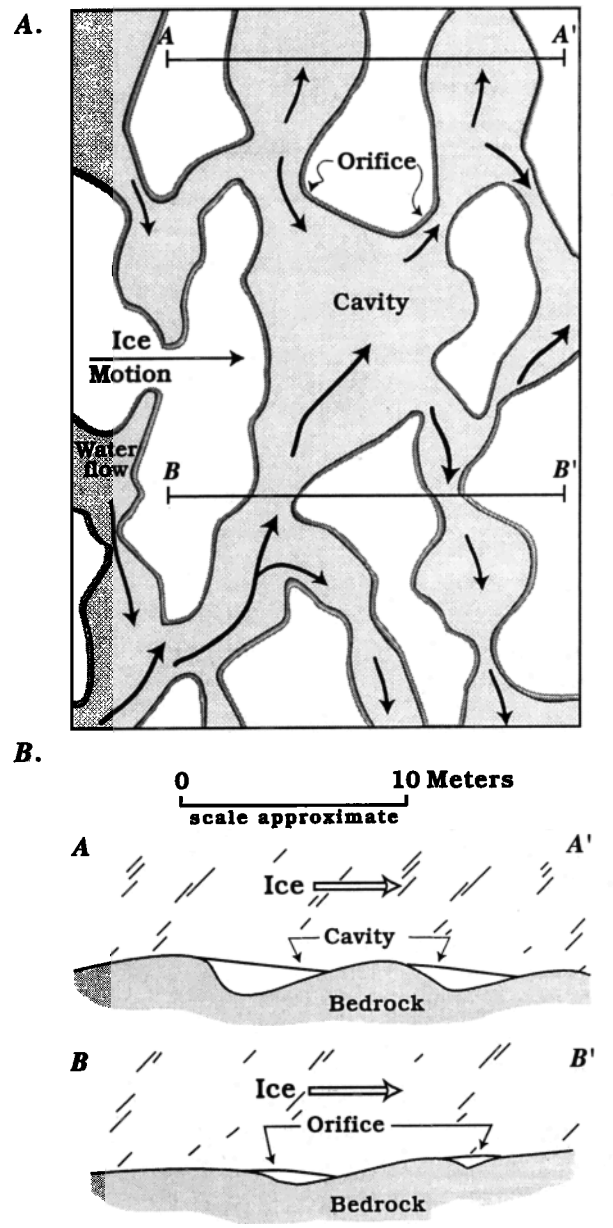

Figure 14. Idealized subglacial cavity network in (a) plan view and (b) cross section [after *Kamb*, 1987]. Unshaded areas are regions of ice-rock contact; shaded areas are regions of ice-rock separation (cavities). Flow directions in the cavities are indicated by arrows. Orifices are the most constricted parts of the cavity network and account for most of the energy losses.

4.1.2.2. Subglacial water film: *Weertman* [1962, 1964, 1966, 1969, 1972] argued that meltwater drainage involved a widespread, thin water layer at the glacier bed (Figure 15). He argued [*Weertman*, 1972] that basal channels were inefficient at capturing meltwater generated at the glacier bed (by geothermal heat and energy dissipated by basal sliding) and that basally generated water must flow in a thin layer, typically ~1 mm thick. *Weertman*'s [1972] argument for the inability of channels to capture meltwater generated at the glacier bed relied on peculiarities of the stress distribution near a channel

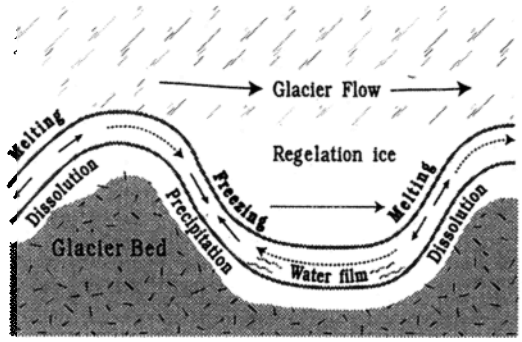

Figure 15. Regelation film of water at the ice-rock interface.

in a material with the rheological properties of glacier ice; he concluded that except near the margins of a channel the pressure gradient at the glacier bed would drive water away from the channel and that water layer drainage was the only plausible alternative. *Weertman*'s [1972] argument was formulated with the implicit assumptions that (1) the glacier-bedrock interface is planar and free of rock debris and (2) the bedrock is impermeable. Actual glacier beds are rough on a range of length scales, and it is probable that there would always be flow paths through the zone of supposedly adverse pressure gradient. Moreover, if there is a permeable layer of rock debris at the glacier bed, and this seems to be typically the case, then meltwater can flow to a channel through the permeable rock debris. The water layer concept has two other problems. *Nye* [1973] pointed out a fundamental inconsistency in postulating a widespread water layer that provides both a path for local redistribution of meltwater involved in the regelation-sliding process, which requires a layer thickness of ~1 μm, and a path for basally generated, through-flowing meltwater, with a characteristic layer thickness of ~1 mm. Some other paths must exist to allow for the net downglacier flow of meltwater. Finally, a water layer at the glacier bed is not stable against perturbations in layer thickness. The instability, proposed by *Nye* [1976] and analyzed by *Walder* [1982], arises because the thicker the layer, the more energy is dissipated by viscous forces and thus the greater is the melt rate. Variations in water layer thickness are therefore magnified, and protochannels tend to develop.

4.1.2.3. Nye channels: *Nye* [1973] suggested that R channels must be transient features, being squeezed shut as they are advected against the upglacier sides of bumps on the glacier bed, and that the maintenance of continuous subglacial meltwater drainage requires the presence of channels incised into the bed (Figure 16). Studies of recently deglaciated bedrock surfaces cited in section 4.1.2.1 in connection with the distribution of cavities [*Walder and Hallet*, 1979; *Hallet and Anderson*, 1980; *Sharp et al.*, 1989] also revealed large numbers of Nye channels, typically ~0.1 m deep, preferentially oriented along the former ice flow direction and commonly connecting cavities to one another. Because these observations were all made on highly soluble carbonate bedrock, it is uncertain how representative they are of glacier beds in general. However, it seems clear that beneath at least some small alpine glaciers, Nye channels constitute a morphologically distinct part of the slow drainage system. Within the context of present-day theory, however, their hydraulic behavior, at least on the mapped glacier beds mentioned above, is simply that of elongated orifices in a cavity network.

4.1.2.4. Subglacial sediments: Recently deglaciated bedrock surfaces nearly devoid of unconsolidated sediment are rare, and it is likely that most glaciers are in fact underlain by a spatially variable, perhaps discontinuous layer of rock debris (Figure 17), which for simplicity (and without sedimentological connotations) we will call glacial till. The till acts as a confined aquifer as long as it is much more permeable than the underlying bedrock (Figure 18); this seems to have been first recognized by *Mathews and Mackay* [1960], although it was not widely appreciated by glaciologists until the work of *Boulton and Jones* [1979]. Even a pervasive basal till layer, however, can evacuate only a small fraction of the total water flux through the glacier. Consider, for example, a till layer with a thickness $B = 0.1$ m and a hydraulic conductivity κ in the range of 10^{-8}–10^{-4} m/s, values that are plausible in light of the field evidence discussed below. Assuming that this layer covers the entire glacier bed, of width W transverse to the ice flow direction, the net meltwater flux through the layer is

$$Q_{\text{Darcy}} = \left(\frac{\kappa B W}{\rho_w g}\right)\left(\frac{d\Phi}{ds}\right) \tag{11}$$

The gradient $(1/\rho_w g)(d\Phi/ds)$, determined largely by the ice-surface slope [e.g., *Shreve*, 1972], is typically ~0.1 for alpine glaciers. Taking $W = 1$ km, we find $Q_{\text{Darcy}} \approx 10^{-7}$–$10^{-3}$ m^3/s, as compared with typical melt-season discharges of ~1–10 m^3/s and winter discharges of perhaps 0.01–0.1 m^3/s [cf. *Lliboutry*, 1983]. Analogous calculations with similar conclusions have been presented for till aquifers beneath ice sheets [*Boulton and Jones*, 1979] and ice streams [*Alley*, 1989]. We conclude that most of the water draining from a till-floored glacier either avoids the bed altogether and simply flows engla-

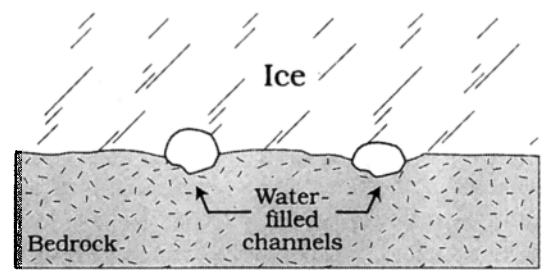

Figure 16. Nye channels (channels incised into subglacial bedrock). A Nye channel may be accompanied by a Röthlisberger channel incised into the overlying ice.

Figure 17. Basal ice resting on unconsolidated sediments at South Cascade Glacier. The photograph was taken in a Röthlisberger channel near the glacier margin. The tape measure at the right shows the scale in both inches (numbers in front) and centimeters (numbers in rear).

cially or else passes directly from englacial conduits into basal conduits.

The characteristics of subglacial channels coexisting with a till aquifer (Figure 19) were analyzed by *Walder and Fowler* [1994]. The geometry of a sediment-floored channel develops in response to flow interactions with both the ice roof and the sediment floor. As in the case of a rock-floored R channel, the channel is enlarged by melting of the ice and shrunken by inward creep of the ice. In addition, the channel may be enlarged by fluvial erosion and closed by inward creep of the till [*Boulton and Hindmarsh*, 1987]. Walder and Fowler showed that a network of sediment-floored channels may exist in two asymptotically distinct conditions: either an arborescent network of sediment-floored R channels at $p_e > \tilde{p}$ or a nonarborescent network of wide, shallow, ice-roofed canals eroded into the sediment at $p_e < \tilde{p}$, where $\tilde{p}$ is a "critical" effective pressure. Which drainage network is stable depends on the magnitude of $\tilde{p}$ (determined primarily by the creep properties of the ice and sediment) and the hydraulic gradient, which is approximated by the ice-surface slope. For very low hydraulic gradients typical of ice streams and ice sheets the drainage network should comprise nonarborescent canals. For large hydraulic gradients typical of alpine glaciers the drainage system should comprise arborescent R channels over relatively "stiff" till, with properties like those of the Breidamerkurjökull till in Iceland [*Boulton and Hindmarsh*, 1987]; however, as was pointed out by *Fountain and Walder* [1993], for alpine glaciers floored by relatively "soft" till [e.g., *Humphrey et al.*, 1993] the stable drainage system would consist of nonarborescent canals.

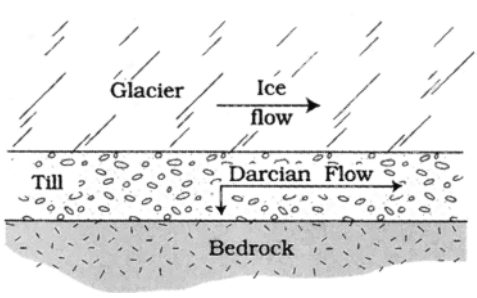

Figure 18. Drainage through subglacial till. Beneath most temperate alpine glaciers a thin layer of unconsolidated till lies between the base of the glacier and the underlying bedrock.

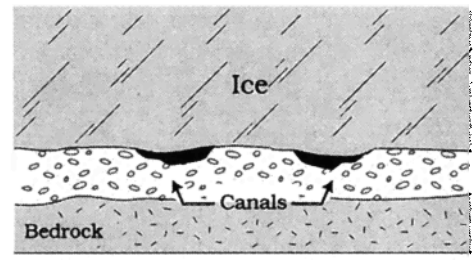

Figure 19. Subglacial "canals" coexisting with subglacial till. Canals tend to be enlarged as the flowing water removes sediment as both bed load and suspended load. Enlargement is counteracted by the tendency for canals to be closed by inward movement of sediment by creep or mass failures from the canal walls. Energy dissipated by the flowing water also melts the overlying ice and counteracts the tendency for ice creep to close off the canal.

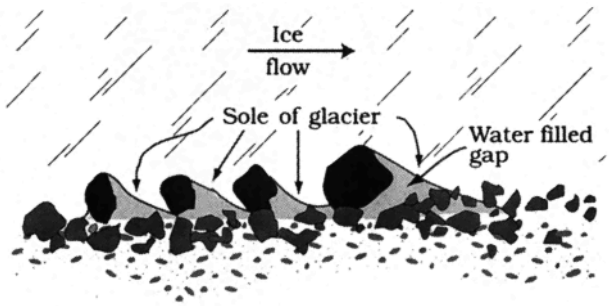

Figure 20. Hypothetical microcavity network at the ice-till interface. Small cavities may form at the lee of relatively large clasts that protrude above the mean surface of the till into the flowing ice.

Basal drainage over a till bed may also involve linked cavities, even if all of the bedrock irregularities are smothered by till. The cavities in this case would simply be gaps on the downglacier side of clasts protruding up into the ice (Figure 20), as was pointed out by *Kamb* [1987].

Clearly, there must be important interactions between the subglacial till layer and the basal conduits, regardless of the exact geometry of the latter. Although the amount of water actually exchanged between the till and the basal conduits may be small, the till provides a pathway for smoothing out water pressure differences between distinct conduits. Depending on the efficiency of the subglacial conduits, the pore pressure p_p within the till aquifer may be close to p_i, with potentially important implications for the mechanical properties of the till.

A consistent set of measurements is beginning to emerge for the hydrological characteristics of subglacial till. For tills that seem to be dilated owing to active shear deformation the porosity is typically near 0.4; nondeforming tills have a porosity more commonly of ~0.25–0.3 (Table 2). There is considerably more variability in the apparent hydraulic conductivity (Table 3), as one might expect from the 6-order-of-magnitude range (10^{-12}–10^{-6} m/s) reported for subaerial till aquifers [*Domenico and Schwartz*, 1990, p. 48]. Laboratory measurements of tills from beneath Breidamerkurjökull [*Boulton et al.*, 1974], ice stream B, Antarctica [*Engelhardt et al.*, 1990], and Storglaciären [*Iverson et al.*, 1994] yield values in the range of 10^{-9}–10^{-6} m/s. It is difficult to compare these values without knowledge of (1) the till grain-size distribution, which was given for the Breidamerkurjökull till [*Boulton and Dent*, 1974] but for neither of the other two samples, and (2) the p_e value during testing, considering the strong dependence of κ on p_e at low p_e [cf. *Boulton et al.*, 1974]. The measured κ values probably all correspond to very low values of p_e.

Estimated hydraulic conductivity values of till inferred from in situ measurements vary substantially. Several in situ measurements yield values broadly consistent with laboratory data and values for subaerial till aquifers. For South Cascade Glacier, *Fountain* [1994] estimated the hydraulic conductivity range from $\kappa = 10^{-7}$ to 10^{-4} m/s on the basis of the migration rate (as observed in several boreholes) of diurnal water pressure fluctuations. The boreholes intersected a region where dye-tracer tests and water level measurements suggested that a basal till layer probably provided hydraulic communication to a low-pressure basal channel. Using the same approach, *Hubbard et al.* [1995] inferred the same range of κ for till at the base of Haut Glacier d'Arolla and also inferred that κ decreased with distance away from a subglacial conduit. A much larger value of hydraulic conductivity for till beneath Gornergletscher has been inferred from borehole-response tests by *Iken et al.* [1996]: 0.02 m/s, a conductivity value typical of medium to coarse gravel [*Domenico and Schwartz*, 1990, p. 48]. Similarly large (or even larger) values were originally inferred by *Stone and Clarke* [1993] from Trapridge Glacier, Canada, borehole-response tests, but more recent work by *Stone et al.* [1997] involving the detailed application of inverse theory has yielded a revised estimate of 5×10^{-4} m/s in the connected part of the drainage system. Much smaller values ($\kappa \approx 10^{-9}$–10^{-8} m/s) have been inferred by *Waddington and Clarke* [1995] for till in unconnected regions of the bed of Trapridge Glacier on the basis of long-term borehole water level variations. The situation at Trapridge Glacier is somewhat unusual because of several factors: (1) the glacier is probably on the verge of surging, (2) the highly correlated behavior of water levels in connected boreholes, with water pressures commonly near (or greater than) the ice overburden pressure, suggests that locally, the glacier is nearly floating on a layer of water, and (3) the glacier margins are frozen to their bed, forcing all meltwater to drain downward through the subglacial till and underlying bedrock.

Another estimate of in situ κ for basal till is wildly at variance with all other data discussed above. This "anomalous" κ value follows from our interpretation of a tracer test at the bottom of a borehole in ice stream B [*Engelhardt et al.*, 1990]. The flow rate of subglacial water

TABLE 2. Porosity of Subglacial Till

Location	*Porosity*	*Reference*
Trapridge Glacier, Yukon Territory, Canada	0.35–0.40	*Stone and Clarke* [1993]
Ice stream B, West Antarctica	0.32–0.40	*Blankenship et al.* [1987]
Ice stream B, West Antarctica	0.40	*Engelhardt et al.* [1990]

TABLE 3. Apparent Hydraulic Conductivity of Subglacial Till

Location	*Hydraulic Conductivity, m/s*	*Reference*
	In Situ Estimates	
Trapridge Glacier	5×10^{-4}	*Stone et al.* [1997]
Trapridge Glacier	10^{-9}	*Waddington and Clarke* [1995]
South Cascade Glacier	10^{-7}–10^{-4}	*Fountain* [1994]
Haut Glacier d'Arolla	10^{-7}–10^{-4}	*Hubbard et al.* [1995]
Breidamerkurjökull	10^{-6}	*Boulton et al.* [1974]
	Laboratory Measurements	
Storglaciären	10^{-7}	*Iverson et al.* [1994]
Ice stream B	10^{-9}	*Engelhardt et al.* [1990]
	Subaerial Till Aquifers	
Various tills	10^{-12}–10^{-6}	*Domenico and Schwartz* [1990]

between two boreholes, inferred from tracer injection, was 30 m/h. Assuming the validity of Darcy's law and estimating the hydraulic gradient from the surface slope of the ice stream (3.5×10^{-4} to 1.25×10^{-3}, according to *Shabtaie et al.* [1987]), we estimate κ = 6.6–24 m/s, about an order of magnitude greater than for coarse gravel [*Domenico and Schwartz*, 1990].

It is not entirely clear how to interpret the various in situ values of κ cited above. The values were all based on simple models that assumed a homogeneous, isotropic till layer of constant thickness. Furthermore, one must be careful in comparing the derived κ values with those for subaerial till aquifers, which may have been affected by consolidation [e.g., *Boulton and Dent*, 1974] or diagenetic processes subsequent to being exposed by retreating ice. Even with these caveats, however, it seems likely that the exceptionally large value of κ for ice stream B represents something other than Darcian flow through the till; for one thing, the Reynolds number is too high for Darcy's law to be valid [cf. *Stone and Clarke*, 1993, p. 332]. In line with *Kamb*'s [1991] analysis the large apparent κ probably reflects the flow through a network of connected "microcavities" at the ice-till interface. This situation is analogous to water movement through fractured rock, in which case, flow through fractures dominates flow through the rock's pore space [*Domenico and Schwartz*, 1990].

As a simple illustration of the relative conductivities of a till layer and a microcavity network, consider the situation sketched in Figure 20 with the microcavities idealized for simplicity as gaps of uniform width h covering a fraction f of the bed. The effective hydraulic conductivity κ_{eff} of a slit of width h is $\rho_w g h^2/12\mu_w$, where μ_w is the viscosity of water; correcting for the fraction of the bed covered by microcavities gives $\kappa_{eff} = f\rho_w g h^2/12\mu_w$. For f = 0.1–0.5 we find κ_{eff} = 0.1–10 m/s for h = 0.3–15 mm. If we account for the tortuosity of the actual flow paths along the ice-till interface, the estimates of h would be perhaps a factor of 2 greater. For comparison, *Kamb* [1991] suggested that the microcavity gaps have a characteristic width of ~1 mm. (We again note that the Reynolds number may be great enough that flow is actually turbulent.) The microcavity concept seems better grounded physically than *Alley*'s [1989] suggestion that flow takes place in a nonuniform "Weertman" water layer. On theoretical grounds it seems plausible that drainage at the bed may also involve wide, shallow, nonarborescent canals incised into the till [*Walder and Fowler*, 1994], particularly if water reaches the bed from the glacier surface. Realistically, the hydraulics of these various scenarios are probably indistinguishable. The precise geometry of the flow paths is less important than the conclusion that a high-conductivity, nonarborescent flow path probably does exist.

In closing our discussion of the subglacial till layer we should emphasize that the hydraulic properties of the subglacial till, and perhaps the till itself, seem to be patchy, with a characteristic length scale of ~10–100 m beneath alpine glaciers [*Engelhardt et al.*, 1978; *Stone et al.*, 1994; *Fountain*, 1994; *Harper and Humphrey*, 1995] and ~100–10^4 m under ice sheets [*Alley*, 1993]. Until more is learned, categorical statements about the properties of subglacial till should be regarded with skepticism.

4.2. Synopsis and Implications

Until about the mid-1980s, subglacial water flow was almost invariably interpreted within the context of an assumed R channel dominated drainage system. The R channel concept had successfully formed the basis for a quantitative theory of outburst floods from glacier-dammed lakes [*Nye*, 1976; *Clarke*, 1982] and was widely applied; for example, *Bindschadler* [1983] discussed the link between basal hydrology and ice dynamics of an Antarctic ice stream and a surging glacier within the context of classic R channel theory. It was known both theoretically [e.g., *Lliboutry*, 1968; *Nye*, 1970] and from studies of exhumed glacier beds [*Walder and Hallet*, 1979; *Hallet and Anderson*, 1980] that cavitation was probably common, but the hydrologic significance of cavities had barely been explored. An analogous statement could be made about the hydrologic significance of

unconsolidated sediment at the glacier bed [*Engelhardt et al.*, 1978].

By about 1990 the way that glaciologists envisaged basal water flow had greatly changed, largely owing to the realization that R channels could not form the basis for an explanation of observations from surging Variegated Glacier [*Kamb et al.*, 1985] and rapidly moving ice stream B in Antarctica [*Blankenship et al.*, 1987]. Theories were developed to elucidate the hydraulics of water flow through linked cavities [*Walder*, 1986; *Kamb*, 1987], deformable till [*Alley et al.*, 1987], and till-floored channels [*Walder and Fowler*, 1994]. The interpretive framework shifted to one in which glaciologists usually tried to explain field data from any particular glacier in terms of a basal drainage system presumed to be dominated by one or another of three morphologically distinct components: R channel, cavity, or till.

We believe that the most significant general conclusion to be drawn from the last 3 decades of investigations is that the basal drainage system is highly heterogeneous in both space and time. It is probable that the various components of the subglacial drainage system have now all been identified, and their hydraulics have been reasonably well described. However, the distribution, spatial extent, and seasonal evolution of each drainage-system component under any particular glacier are still uncertain. The way in which the drainage-system components interact also remains poorly understood. These issues need to be addressed to improve our understanding of both glacier dynamics and hydrology. The link between the subglacial drainage system and groundwater flow also remains unexplored aside from gross generalizations [cf. *Lliboutry*, 1983].

To summarize, the drainage system under any given glacier comprises several or all of the morphologically distinct components described in this section. A slow, nonarborescent drainage system, comprising a mixture of elements including cavities, permeable till, and conduits incised into the bed (i.e., Nye channels and canals), probably covers most of the glacier bed and is nearly fixed relative to the bed. The water pressure in the slow drainage system is commonly close to the ice-overburden pressure. A fast drainage system consisting of arborescent R channels may also exist. The fast drainage system, being incised into the base of the glacier, is advected by glacier movement and probably undergoes continuous rearrangement as the sliding ice interacts with the rough bed beneath. The water pressure in the fast drainage system is commonly much less than the ice overburden pressure; indeed, the fast drainage channels may be only partly full most of the time, in which case the flow is unpressurized except at times of peak diurnal discharge. Both the fast and slow components of the basal drainage system probably undergo major temporal changes, particularly at the beginning and end of the melt season, as discussed in section 5.

5. TEMPORAL EVOLUTION OF THE DRAINAGE SYSTEM

Meltwater discharge from temperate alpine glaciers varies typically by about 2 orders of magnitude from winter to summer, so it seems plausible that the subglacial drainage system must also undergo seasonal changes. Data bearing on this question are sparse, as little information is available except for the ablation season. Moreover, there is not much of a theoretical foundation for understanding time-dependent discharge.

It is very unlikely that a robust system of R channels can survive from year to year except perhaps beneath ice only a few tens of meters thick. This is readily seen by considering the fate of a channel of radius R that becomes empty of water at the end of the melt season. The channel will tend to be closed by inward creep of ice, with the channel radius as a function of time given by [*Weertman*, 1972]

$$R(t) = R_0 \exp(-t/\tau) \tag{12}$$

where R_0 is the radius at the end of the melt season. The characteristic time τ is proportional to p_i^{-n}, where $n \approx 3$. For ice thicknesses ≥ 150 m a channel that has survived the winter will have a radius at the beginning of the melt season of $\leq 10^{-3} R_0$, probably no more than a few millimeters for even the main trunk channels. A similar argument can be made for the closure of Nye channels and sediment-floored canals. In contrast, a cavity network ought to be able to survive from 1 year to the next because cavities are maintained open primarily by basal sliding and only secondarily by melting. As long as sliding does not cease in winter and the accumulation of sediment is not too great, cavities stay open, although some of the links between the cavities may become quite constricted. Thus we suggest that the subglacial drainage system beneath the entire glacier at the beginning of the melt season should typically be cavity dominated, with the cavities probably poorly connected. A similar scenario has been discussed by *Raymond* [1987] in connection with glacier-surge initiation, which we will discuss in section 8. In the case of a sediment-floored glacier the cavities will be located at the downglacier sides of isolated bedrock protuberances or relatively large boulders that protrude above the mean local till surface [*Kamb*, 1987].

At the beginning of the melt season, once meltwater penetrates the winter snowpack, it flows into crevasses and preexisting englacial channels linked to the crevasses. In the event that such links were severed during winter, water will pond in the crevasses while slowly escaping through intergranular passages or microfractures in the ice; the escaping water eventually enlarges these flow paths by melting, and the cycle of englacial channel development starts anew. The englacial network of channels begins to fill again, and water makes its way

to the glacier bed. Initially, the basal drainage system is unable to cope with the increased meltwater flux. The initial response will essentially be like that proposed by *Iken et al.* [1983], who measured vertical uplift of the surface of Unteraargletscher, Switzerland, at the beginning of the melt season: Water pressure will increase, and cavities will enlarge. During this period of time the glacier will store large volumes of water, consistent with the observations of *Tangborn et al.* [1975] and *Willis et al.* [1991/1992]. The rate of water input to the glacier varies slowly in time as long as the available storage in the firn and the seasonal snowpack are great enough. Therefore, in the ablation zone the water flux into the glacier interior must begin to exhibit large diurnal variations once the seasonal snow has melted [*Nienow*, 1994]. Once this happens, englacial channels fed from the ablation zone enlarge and cut down rapidly, and the parts of the basal hydraulic system fed by these channels become subjected to large daily variations in water pressure. The transient increase in water pressures, as well as in the melting caused by the increase in flow, enlarges the orifices in the basal cavity system. If the englacial water flux, and thus the pressure perturbation, is great enough, cavity orifices will enlarge unstably and spawn an R channel system [*Kamb*, 1987]. However, if the flux reaching the bed through an englacial conduit is sufficiently small, the local cavity system will remain stable. Thus we envisage that both the fast and slow components of the basal drainage system receive water directly from the glacier surface and that R channels represent the continuation at the glacier bed of the largest englacial conduits. These R channels probably endure throughout the melt season [*Sharp et al.*, 1993a] as long as the water supply from the glacier surface is sufficient to melt the ice walls and thus counteract creep closure.

Because the hypothetical process of R channel development described above is driven by water input from the surface, it seems reasonable that this sort of development in the basal drainage system should progress upglacier as the melt season progresses. Evidence for such a spatial progression has been presented by *Nienow* [1994], who inferred from dye-tracer studies at Haut Glacier d'Arolla that the boundary between a slow, nonarborescent, wintertime drainage system and a fast R channel system moved upglacier through time as the snow line on the glacier surface retreated.

Data bearing on the seasonal evolution of the basal drainage system are sparse. *Seaberg et al.* [1988] and *Hock and Hooke* [1993] injected dye into moulins in the lower part of the ablation area of Storglaciären. They calculated subglacial flow speeds mostly in the range of 0.1–0.2 m/s and also found that the dispersivity progressively decreased. They inferred that the basal drainage system included shallow but very wide conduits (Figure 21), not at all like the classic conception of an R channel but similar to the predicted geometry of canals on a sediment bed (Figure 19). *Seaberg et al.* [1988] and *Hock and Hooke* [1993] suggested that the conduits were essentially ice-roofed braided streams and interpreted the progressive decrease in dispersivity as reflecting a decrease in the degree of braiding. *Fountain* [1993] injected dye mostly into crevasses at South Cascade Glacier and compared the measured travel time of tracers with that derived from calculations based on a model of turbulent flow in conduits. He also concluded that the subglacial conduits were shallow and broad and that they were pressurized early in the melt season but evolved toward open-channel flow as the season progressed.

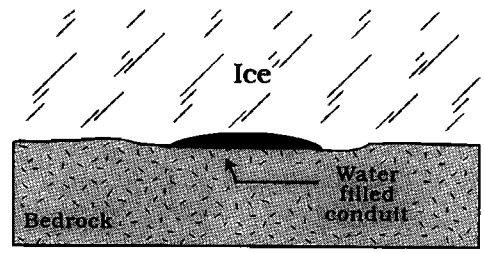

Figure 21. Hypothetical broad, shallow conduit incised into the basal ice, a variant of the standard concept of a semicircular Röthlisberger channel. Basal channels may be broad and shallow because ice creeps inwardly more quickly at the top of the channel (which is not full at times of low flow) than at the sides of the channel (where the creep rate is reduced by the effect of drag at the ice-rock interface).

6. WATER STORAGE

Glaciers store and release large volumes of water on daily to seasonal timescales [*Tangborn et al.*, 1975], making short-term runoff prediction difficult. Outburst floods, resulting from englacial and subglacial water storage and release, are probably common to most glaciers; small floods, with a peak discharge not much larger than the typical maximum daily flow, probably occur frequently but are rarely detected [cf. *Warburton and Fenn*, 1994], whereas large, destructive floods are rare. Water storage also affects glacier movement. Surging glaciers store large volumes of water, and major surge slowdowns as well as surge terminations are correlated with the release of stored water [*Kamb et al.*, 1985; *Humphrey and Raymond*, 1994]. The rapid flow of Columbia Glacier, Alaska, seems to be controlled by the volume of stored water [*Meier et al.*, 1994; *Kamb et al.*, 1994]. More generally, the ability of glaciers to store water modulates rapid changes in the basal water pressure and may help to increase and sustain glacier motion by distributing high water pressure over large bed regions [*Humphrey*, 1987; *Alley*, 1996].

Small, temperate alpine glaciers seem to attain a maximum seasonal water storage of ~200 mm of water averaged over the area of the glacier bed, with daily fluctuations of as much as 20–30 mm [*Tangborn et al.*, 1975; *Willis et al.*, 1991/1992]. It is tempting to assume that subglacial cavities provide the capacity for most of

the storage, but this is questionable. The volume of the former linked cavity system beneath Castleguard Glacier, Alberta, was inferred to be equivalent to an average water layer thickness of only ~27 mm [*Hallet and Anderson*, 1980], and this figure is within a factor of 2 for exhumed cavity systems at Blackfoot Glacier, Montana, United States [*Walder and Hallet*, 1979] and Glacier de Tsanfleuron, Switzerland [*Sharp et al.*, 1989]. The discrepancy between the estimated stored volume and the likely storage capacity of subglacial cavities is also evident when we consider the situation at Variegated Glacier, where the volume of water stored during the glacier surge of 1982–1983 was equivalent to a layer of water ~1 m thick [*Humphrey and Raymond*, 1994]. This seems like an implausibly large amount of water to store in subglacial cavities, even if a majority of the glacier was separated from the bed, unless the glacier bed was extraordinarily rough.

Subglacial till may play an important, although not predominant, role in seasonal water storage. An unconsolidated till layer covering the entire glacier bed, a questionable scenario, and having a thickness of, say, 0.25 m and a porosity of 0.30 would provide a storage volume equivalent to a 75-mm-thick layer of water. Accommodating as much as 200 mm of stored water in a basal till layer would require that the till be substantially thicker than that suggested by the limited borehole data from alpine glaciers [e.g., *Engelhardt et al.*, 1978; *Stone et al.*, 1997]. Furthermore, diurnal variations in water storage probably cannot be explained with reference to basal till. The fractional change in storage in a saturated till layer is given approximately by $S_s \Delta h$, where S_s is the specific storage, a measure of the water volume stored or released as the till layer is strained, and Δh is the variation in hydraulic head. With $S_s \sim 10^{-4}$/m for sandy or gravelly till [cf. *Stone et al.*, 1997; *Domenico and Schwartz*, 1990] and $\Delta h \sim 100$ m (at most) the fractional change in storage is ~0.01 or ~1 mm. By process of elimination we conclude that a substantial fraction of the water storage, both short- and long-term, is probably englacial [cf. *Jacobel and Raymond*, 1984; *Humphrey and Raymond*, 1994].

Borehole-video studies [*Pohjola*, 1994; *Harper and Humphrey*, 1995] suggest that englacial voids and conduits in small temperate glaciers constitute a macroporosity of ~0.4–1.3%, although some of this probably comprises isolated, water-filled voids. *Fountain* [1992] estimated a macroporosity of 1% to maintain reasonable calculated subglacial water pressures. Englacial water storage is an attractive hypothesis because a macroporosity of only 0.1% in hydraulic communication with the bed would be equivalent to a 100-mm-thick water layer for a glacier with an average thickness of only 100 m. Filling and draining of englacial passages have been detected by radar [*Jacobel and Raymond*, 1984], and the filling of moulins has been measured [*Iken*, 1972]. Thus it seems both qualitatively and quantitatively likely that englacial storage may exceed subglacial storage in many cases. Water storage in surging glaciers also involves water-filled surface potholes [*Sturm*, 1987] and crevasses [*Kamb et al.*, 1985]. Near-surface storage of this sort implies that water is in fact backed up in englacial passages.

7. OUTBURST FLOODS

A glacial outburst flood, sometimes called by the Icelandic term "jökulhlaup," may be broadly defined as the sudden, rapid release of water either stored within a glacier or dammed by a glacier. Although outburst floods are perhaps best known for the hazards they pose in alpine regions, they are not limited to such glaciers but are also associated with large tidewater glaciers in Alaska [*Mayo*, 1989] and Icelandic ice caps [*Björnsson*, 1992]. The Pleistocene Missoula floods, the largest known terrestrial floods, were caused by periodic drainage of an enormous lake dammed by the Cordilleran ice sheet in western North America. Floodwaters swept across an area of $\sim 3 \times 10^4$ km^2 in present-day Washington State, creating the unique landscape known as the Channeled Scabland [*Waitt*, 1985].

Most observed outburst floods involve drainage of glacier-dammed lakes. The water either drains through a subglacial tunnel and appears at the glacier terminus, or drains through a breach between the glacier and valley wall. For drainage through subglacial tunnels, drainage occurs, to a first approximation, when the lake level rises sufficiently to nearly float the ice dam [*Björnsson*, 1974]. As water begins to leak under the dam, frictional energy dissipation causes flow to localize in a channel. Initially, the channel enlarges rapidly by melting, but as the lake level drops, water pressure in the channel falls, the rate of inward ice creep increases, and the channel closes, thereby terminating the flood. The flood hydrograph commonly has a long, gentle ascending limb and a steep, abrupt falling limb, with the flood lasting a few days to weeks (Figure 22a). Details of the physics have been elucidated by *Nye* [1976], *Glazyrin and Sokolov* [1976], *Spring and Hutter* [1981], and *Clarke* [1982]. Clarke's analysis has the greatest utilitarian value. He showed that the exit hydrograph is determined primarily by (1) the temperature θ_{LAKE}, volume V, and hypsometry of the lake, (2) the ice overburden pressure p_i where the tunnel meets the lake, and (3) the tunnel roughness $\tilde{n}$ and mean hydraulic gradient G. These factors are incorporated into two dimensionless parameters, a "tunnel closure parameter" α and a "lake temperature parameter" β. Exit hydrographs can be calculated if plausible bounds on α and β can be estimated. For many alpine glaciers, α is relatively small, and plausible bounds can be placed on peak discharge Q_{MAX} even without calculating the complete exit hydrograph. The lower bound corresponds to the case in which $\theta_{LAKE} = 0$°C; the upper bound corresponds to the case in which tunnel enlarge-

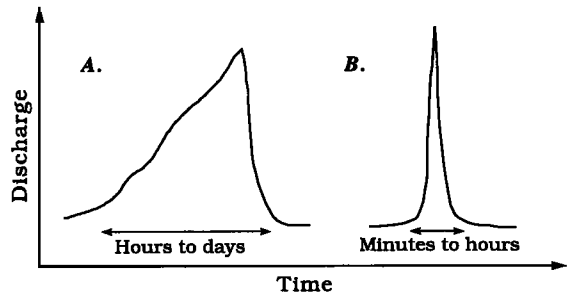

Figure 22. Outburst-flood hydrographs for two distinct types of water release: (a) flood caused by lake water tunneling under a glacier and (b) flood caused by release of subglacially stored water or by subaerial breaching of an ice dam [after *Haeberli*, 1983]. Reprinted from *Annals of Glaciology* with permission of the International Glaciological Society.

ment is dominated by the lake's thermal energy and frictional dissipation is negligible.

Where a valley is blocked by a glacier advancing from a tributary valley (Figure 23), the glacier-dammed lake commonly drains through a breach between the ice dam and an adjacent rock wall [*Walder and Costa*, 1996]. The way in which drainage begins is enigmatic. In some cases, drainage may begin through a subglacial tunnel near the valley wall because the ice is normally thinnest at that point. Tunnels formed in this way seem to be prone to roof collapse, and marginal breaches develop [e.g., *Liss*, 1970]. Alternatively, because the ice-wall contact is typically irregular, seepage through the gaps erodes the ice through frictional heating, thereby initiating a breach (M. F. Meier, personal communication, 1994). A theoretical description of breach-drainage outburst floods has been given by *Walder and Costa* [1996]. Their analysis parallels *Clarke*'s [1982] analysis of tunnel-drainage outbursts in many important respects, particularly in the assumption that breach enlargement proceeds by melting of the ice. Calving also widens the breach [*Liss*, 1970] but is not hydraulically important unless calved ice actually blocks the breach. Breach closure resulting from ice movement is negligible. The flood hydrograph commonly exhibits a steep rising limb (Figure 22b), similar to floods caused by the failure of constructed dams [*MacDonald and Langridge-Monopolis*, 1984]. The hydrograph depends primarily on lake hypsometry and two dimensionless parameters, a "breach roughness parameter" γ and a lake temperature parameter δ. Both γ and δ depend on the initial lake volume and depth. Again, approximate bounds can be placed on Q_{MAX} for cases in which the effect of the lake's thermal energy on breach erosion is either negligible or dominant.

Simple empiricisms have also been useful for estimating Q_{MAX} from lake drainage. *Clague and Mathews* [1973] were the first to present a regression relation between Q_{MAX} and V:

$$Q_{MAX} = bV^a \tag{13}$$

where Q_{MAX} is given in cubic meters per second and V is given in millions of cubic meters and with values of $b = 75$ and $a = 0.67$. *Walder and Costa* [1996] updated the Clague-Mathews relation by developing separate regressions for tunnel drainage ($b = 46$ and $a = 0.66$) and breach drainage ($b = 1100$ and $a = 0.44$). For a given value of V, Q_{MAX} for a breach-drainage flood is commonly much greater than for a tunnel-drainage flood. The values of the constants for a breach drainage are quite similar to the regression developed by *Costa* [1988] for floods from constructed earthen dams ($b = 1200$ and $a = 0.48$). The similarity of constants is probably not fortuitous but rather reflective of the fact that both were developed for floods that drained through rapidly formed subaerial breaches. The differences in material properties of the dams and in erosion mechanisms are evidently less important than the flow hydraulics, which are the same in both cases.

Another class of outburst floods involves the abrupt release of water from subglacial or englacial storage [*Haeberli*, 1983; *Driedger and Fountain*, 1989]. Outburst

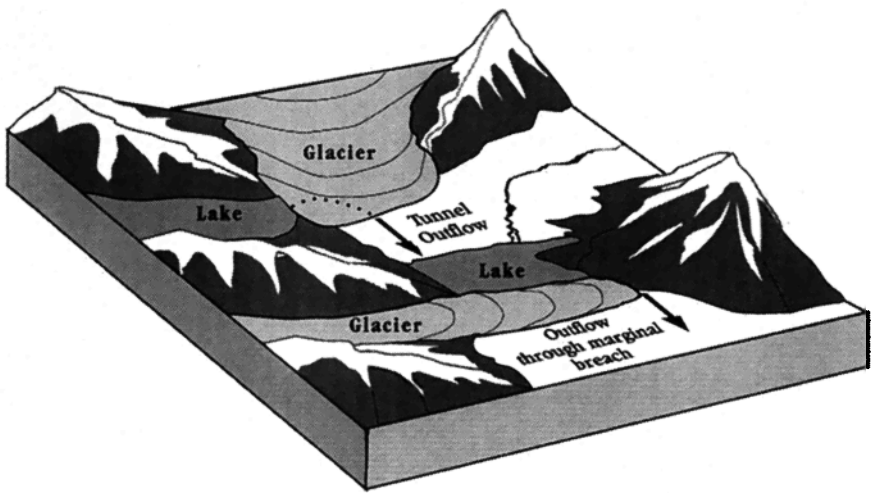

Figure 23. Contrasting modes of glacier-dammed lake formation. A lake impounded by the advance of a glacier across the mouth of a tributary valley (left side of figure) drains through a subglacial channel, but a lake formed by the advance of a glacier from a tributary valley typically drains through a subaerial breach, usually at the terminus. Based on Figure 3 of *Walder and Costa* [1996]; copyright John Wiley and Sons Ltd.; reproduced with permission.

floods of this type, which are by their nature unanticipated and poorly described, seem to be triggered by rapid input of rain or meltwater to the glacier. *Walder and Driedger* [1995] suggested that the release mechanism probably involves unstable enlargement of the orifices in a basal cavity network that transforms into one that drains water rapidly. This mechanism, initially proposed in connection with glacier-surge termination [*Kamb*, 1987], is also considered important to the annual reestablishment of an R channel network [cf. *Nienow*, 1994; M. J. Sharp, personal communication, 1996]. Borehole measurements showing an abrupt reorganization of the basal drainage system, consistent with the scenario discussed here, have been collected at Trapridge Glacier [*Stone and Clarke*, 1996]. Alternatively, water could be stored englacially in passages temporarily isolated from the subglacial drainage system, then released when rapid input of water to the glacier forces reconnection with the bed.

Floods from internally stored water are largely unpredictable. *Walder and Driedger* [1995] used statistical methods to show that for South Tahoma Glacier (Mount Rainier, Washington State), which released 14 or 15 floods during a 6-year period, the probability of a flood increased as the input rate of water to the glacier (as rain or meltwater) increased. These results agree with the observations of *Warburton and Fenn* [1994]. Unfortunately, a relationship of this sort developed for a particular glacier is unlikely to be applicable elsewhere. A physically plausible, albeit crude, estimate of Q_{MAX} is nonetheless possible. Glaciological experience [e.g., *Haeberli*, 1983; *Walder and Driedger*, 1995] suggests that the water volume released from storage during an outburst flood is likely to be of a magnitude corresponding to a water layer ~10–100 mm thick over the entire glacier bed and that the release typically occurs during a period τ equal to ~15–60 min, although sometimes as long as a day [*Warburton and Fenn*, 1994]. Estimating the released water volume as the product of the glacier-bed area A and equivalent water layer thickness d and assuming a triangular exit hydrograph, we estimate

$$Q_{MAX} \approx \frac{2Ad}{\tau} \quad (14)$$

As an example, consider a small alpine glacier with A = 1 km^2. The base flow from such a glacier is probably ~1 m^3/s [e.g., *Fountain*, 1993]. From (14) we estimate an upper bound for Q_{MAX} of ~100 m^3/s. Flood peaks of this magnitude can be extremely destructive in small alpine drainage basins, particularly if the water floods transform to debris flows [*Driedger and Fountain*, 1989; *Walder and Driedger*, 1995].

8. LINKS BETWEEN HYDROLOGY AND GLACIER DYNAMICS

8.1. Effect of Glacier-Surface Morphology

Variations in water input to a glacier should affect basal water pressure. If water moves rapidly from the glacier surface to the bed, a proposition that we examine in section 9, the water pressure in bed areas supplied from the ablation zone should respond rapidly (probably within a few minutes to a few hours) to variations in the water input at the surface. In contrast, water pressure in bed areas supplied from the accumulation zone should respond slowly (probably on a timescale of days to a week or more) to varying water input at the surface, owing to delayed transport through snow and firn. A change in supply region from accumulation to ablation zone should therefore be reflected at the base of the glacier by a relatively sharp gradient in variations of subglacial water pressure. This may have important implications for glacier dynamics because the rate of glacier sliding is, in part, related to the effective pressure (ice pressure minus water pressure) at the base of the glacier [*Iken and Bindschadler*, 1986; *Jansson*, 1995]. We expect the sliding speed in the ablation zone to be greater than that in the accumulation zone during peak diurnal melt periods but less when the melt rate is at a minimum. The ablation zone would then "pull" the accumulation during midday, and the accumulation zone would "push" the ablation zone during the night. Similarly, the ablation zone would move fastest during the first few days of a rainy period before the water percolated through the accumulation zone and slowest just after the rain stopped. This scenario would be somewhat modified if most of the surface water input were routed directly to subglacial channels. Not only do the channels only pressurize a small part of the bed, but during their largest development in midsummer the conduits may only be pressurized during a short time each day. Under these conditions the accumulation zone may more or less constantly push the ablation zone. Because spatial variations in glacier movement are smoothed by longitudinal stress-gradient coupling over a distance related to the glacier thickness [*Kamb and Echelmeyer*, 1986], differences in flow speed between the accumulation and ablation zones caused by variations in water input should tend to increase with the length of the glacier.

8.2. Subglacial Hydrology

A large body of data has accumulated suggesting a link between variations in the basal drainage system and perturbations in glacier movement, but the physical nature of the coupling remains elusive. The best known evidence suggesting the hydrology-dynamics link involves seasonal variations in glacier-surface velocity, first observed by *Forbes* [1846] at Mer de Glace, France. Generally, the surface velocity peaks in late spring to early summer in the ablation area; in the accumulation area the seasonal variation may be of the opposite phase. The usual interpretation [e.g., *Hodge*, 1974] is that changes in surface velocity are too large to be explained by mass balance induced changes in applied stress and that changes in surface velocity therefore reflect changes in sliding velocity. Such an interpretation requires caution. *Balise and Raymond* [1985] showed theoretically

that the transfer of basal-velocity variations to the glacier surface is sensitively dependent on the length scale of such variations. They concluded that broad-scale variations in basal sliding should be reflected by similarly broad-scale variations in surface speed but that very localized basal-velocity variations cannot be unambiguously resolved by glacier-surface observations.

A key point of contention has been whether sliding speed is controlled primarily by the volume of stored water or by basal water pressure. *Hodge* [1974] showed that the surface speed of Nisqually Glacier, Washington State, peaked before the meltwater discharge from the glacier and also that the speed actually increased throughout the winter, even while meltwater discharge was falling. He interpreted this to mean that sliding speed was controlled by the amount of water stored at the glacier bed, with the maximum storage occurring early in the melt season before an efficient basal drainage system had developed (in line with our discussion in section 5). In contrast, *Iken et al.* [1983] found that the maximum sliding speed coincided with times when the glacier surface was rising most rapidly, the surface rise being thought to indicate water going into storage, rather than with the time of maximum surface elevation; they interpreted this to mean that sliding speed was a function of subglacial water pressure rather than storage. *Iken and Bindschadler* [1986] subsequently showed a correlation between velocity fluctuations and borehole water pressures at Findelengletscher, and similar measurements have been made at Storglaciären [*Jansson*, 1995].

Probably, the most detailed observations dealing with the link between basal hydrology and glacier dynamics are those from Columbia Glacier, a rapidly moving tidewater glacier [*Meier et al.*, 1994; *Kamb et al.*, 1994]. Surface-velocity fluctuations at two sites 7 km apart were strongly correlated with each other and fairly well correlated with the borehole water level at the upglacier of the two sites but not with the borehole water level at the lower site. Using estimates for recharge to and discharge from the basal drainage system, *Kamb et al.* [1994] concluded that variations in ice velocity were best correlated with variations in the amount of water stored at the glacier bed.

We appear to be faced with a conundrum. Models of the basal-cavitation process [*Lliboutry*, 1968; *Iken*, 1981; *Fowler*, 1986, 1987; *Kamb*, 1987] predict an increase in basal storage with an increase in basal water pressure, yet glacier movement seems sometimes to correlate with storage, sometimes with water pressure, but not with both. *Kamb et al.* [1994] suggested a resolution of this conundrum, as follows: Glacier sliding speed u_b and basal storage are controlled by $\langle p_w \rangle$, the basal water pressure averaged over the distance l, the length scale over which the basal shear stress is effectively averaged by glacier dynamics [*Kamb and Echelmeyer*, 1986]. The basal water pressure p_w measured at a point, however, is the sum of the spatial mean value $\langle p_w \rangle$ and a local fluctuating value p'_w, where p'_w is controlled by rapid, local reorganization of the basal drainage system. Thus one expects u_b to correlate with storage but not necessarily with local p_w.

8.3. Glacier Surging

Surging glaciers exhibit [*Raymond*, 1987, p. 9121] "a multi-year, quasi-periodic oscillation between extended periods of normal motion and brief periods of comparatively fast motion." A thorough review of glacier surging is beyond the scope of this paper (we refer the reader to the review by *Raymond* [1987]), but we do want to highlight recent developments related to the study of the 1982–1983 surge of Variegated Glacier [*Kamb et al.*, 1985; *Humphrey et al.*, 1986; *Kamb*, 1987; *Humphrey and Raymond*, 1994] that point to regulation of the surge process by basal water flow.

Kamb [1987] noted the following observations from Variegated Glacier as the basis for his surge model:

1. Borehole measurements demonstrate directly that rapid glacier motion during the surge is due to basal sliding.
2. Basal water pressure during the surge was close to the overburden pressure and notably higher than during the nonsurging state. Peaks in water pressure corresponded with peaks in sliding motion in both surging and nonsurging states.
3. Major decreases in surge motion, as well as surge termination, were accompanied by large flood peaks in outlet streams and a lowering of the glacier surface, indicating that the high sliding speed and water pressure during the surge are coupled with water storage within and at the bed of the glacier.
4. Dye-tracing experiments [*Brugman*, 1986] showed that the mean flow of water in the basal drainage system was ~25–30 times faster after surge termination than during the surge. Moreover, dye appeared at a number of locations across the width of the glacier during the surge, but in only a single stream after surge termination.
5. Water discharged from the glacier during the surge was extremely turbid; suspended-sediment concentration was much higher, and the average grain size of suspended sediment was finer during the surge than in the nonsurging phase [*Humphrey and Raymond*, 1994].

A physical model of surging that accounts for these observations was developed by *Kamb* [1987], who proposed that the basal drainage system during surge comprised a linked-cavity network, whereas the drainage system during the nonsurging phase consisted of arborescent R channels. The cavity system is associated with high water pressure and multiple, tortuous flow paths leading to prolonged, highly dispersed dye returns. Surge slowdowns and surge termination result from large transient increases in basal water pressure that destabilize part of the cavity system, thereby releasing water from storage. Sediment concentration in the meltwater discharged from the glacier increased during the surge because a linked-cavity drainage system brought a

large fraction of the glacier bed into contact with flowing water, but the mean suspended-sediment size dropped during the surge because the sluggishly flowing water in the cavity system could not suspend as much coarse sediment as could rapidly flowing, channelized water. [*Humphrey and Raymond*, 1994].

The conditions that cause surge initiation can also be explained, at least qualitatively, in the context of the channel-cavity dichotomy. *Raymond* [1987] and *Kamb* [1987] suggested that during winter, R channels collapse and a high-p_w linked-cavity network develops. Usually, as the melt season begins, the flux of meltwater to the bed causes water pressure transients that destabilize parts of the linked-cavity network, and an R channel network reforms. (We have suggested in section 5 that this scenario is probably common to all temperate glaciers, not just those that surge.) The stability of the cavity network to pressure perturbations is controlled by a parameter Ξ [*Kamb*, 1987] that depends on roughness characteristics of the glacier bed, ice rheology, and glacier geometry; in a surging glacier, as the glacier geometry (primarily the thickness) changes with time during the nonsurging phase, a point is reached at which Ξ attains a small enough value to stabilize the wintertime cavity network against early melt-season water pressure perturbations. When this occurs, the high-pressure linked-cavity system persists and enlarges, and surging begins. Complications in this scenario have been discussed by *Humphrey and Raymond* [1994].

9. UNIFIED MODEL OF WATER FLOW THROUGH A TEMPERATE GLACIER

Water enters the body of a glacier primarily through crevasses and moulins. The englacial drainage system comprises a complex combination of gently inclined passages spawned by water flow along crevasse bottoms and steeply inclined passages formed by water enlarging intergranular veins. In general, water flows englacially for long distances, perhaps equal to several times the glacier thickness, before reaching the bed, although the common presence of moulins in the ablation zone indicates that water can sometimes descend vertically through a significant fraction of the ice thickness.

The englacial conduit system supplied from the accumulation zone is of relatively limited extent compared with the system supplied from the ablation zone because of the role of the firn in damping diurnal variations in water input. Much of the water that enters the glacier in the accumulation zone probably reaches the bed in the ablation zone. Thus the subglacial area of influence of each zone is shifted downglacier, and the firn influences a subglacial area greater than the area it actually covers.

The supply of surface water to the bed is inhibited in overdeepened parts of the glacier because the gently inclined parts of the englacial conduit system become pinned by the downglacier margin of the overdeepening. Basal water flow in an overdeepening is essentially restricted to the water already in the basal drainage system upglacier of the overdeepening; basal conduits may tend to freeze shut where they encounter the adverse slope coming out of the overdeepening. Basal conduits should be most frequently located along the margins of the overdeepening.

The morphology of the subglacial drainage system is controlled by a number of factors, including the distribution of englacial conduits reaching the bed, ice thickness, glacier sliding speed, bed lithology, and bed roughness. Any one of these factors may be of relatively greater or lesser importance at any particular glacier. Generally speaking, the morphology of the basal drainage system is heterogeneous. Slow drainage systems, involving linked cavities, permeable till, and channel segments incised into the bed and trending along the ice flow direction, cover most of the bed. The slow drainage system is in poor hydraulic communication with a fast system of R channels incised into the basal ice. The R channel system largely collapses during winter and is reformed in the spring as a flush of water reaches the bed and destabilizes parts of the linked-cavity network. In relatively thick ice, say, 200 m or more, there is probably ample opportunity for englacial drainage to become concentrated into a relatively small number of trunk conduits, each carrying a large water flux, whereas in thin ice, say, 50 m or less, the englacial flow is relatively more diffuse, with a large number of englacial conduits, each carrying a small flux of water, reaching the bed.

10. DIRECTIONS FOR FUTURE RESEARCH

Glaciologists need to adopt a holistic perspective in studying glacier hydrology. Indeed, although we have written separately about near-surface, englacial, and subglacial water flow, the three phenomena are obviously coupled. Influences in the glacier drainage system nearly always move from the glacier surface downward. Forcings imposed on the englacial and subglacial passages are distinctly different, depending on whether water is supplied from the accumulation zone or the ablation zone.

Coupling between the near-surface and englacial drainage systems needs to be investigated much more thoroughly. There are almost no data available showing how water flux to crevasses and moulins is distributed over the glacier surface and how this distribution evolves temporally. These flux data constitute a fundamental boundary condition for the englacial part of the drainage system.

The water flux delivered to the bed at the points of coupling between the englacial and subglacial drainage systems constitutes the "upstream" boundary condition on the subglacial drainage system. This flux distribution obviously cannot be directly measured, but it may still be

investigated once we recognize that surface water supplied to the englacial drainage system almost certainly becomes concentrated into a relatively limited number of trunk conduits by the time it reaches the glacier bed [*Shreve*, 1972]. It may be possible to use tracers to delineate the "drainage basins" of the englacial trunk conduits (i.e., the ones that reach the bed) and thereby to estimate the distribution of recharge to the subglacial drainage system, much as tracers have been used to delineate the gross drainage-basin structure of entire glaciers [e.g., *Stenborg*, 1973; *Fountain*, 1992, 1993; *Fountain and Vaughn*, 1995].

Some of the theoretical foundations of glacier hydrology theory need to be revisited. *Clarke* [1994] has recently begun doing this for the case of Röthlisberger channels by critically examining one of the key simplifying assumptions (the neglect of heat advection by the water) in *Röthlisberger*'s [1972] analysis. (At the time of writing, Clarke's recent work has appeared only as an abstract, and we cannot assess it critically.) There is also a distinct need to understand how the drainage system should respond to time-varying water input. This topic has been touched upon by *Spring* [1980], who explored the pressure-discharge relation for sinusoidally varying flow in R channels, and by *Kamb* [1987] in his analysis of the stability of cavities to pressure transients.

The time seems to be ripe for constructing theoretical models that fully couple glacier sliding and basal hydrology, accounting properly for both the long-range spatial averaging imposed by ice dynamics and the complex, time- and space-dependent variations within the basal drainage system. Some important studies that we believe can jointly serve as a springboard are those of *Humphrey* [1987], *Murray and Clarke* [1995], and *Clarke* [1996].

Humphrey [1987] presented the only analysis to date of the dynamic coupling between a glacier and its basal drainage system, albeit within the context of a highly idealized view of glacier-bed geometry. He argued that a complete description of the coupling between subglacial water flow and glacier dynamics requires one to specify the following: (1) a force balance at the bed, (2) the coupling between the basal shear stress and the stresses in the body of the glacier, with careful attention to longitudinal stress gradients, (3) a relation between cavity size, sliding speed, and basal water pressure, and (4) a description of the hydraulics of water flow in the linked cavities. The mathematical model is complicated, and its consequences have not yet been fully elucidated, but the results are tantalizing. Most importantly, from the perspective of glaciologically meaningful measurements, Humphrey showed that the model predicts no simple relation between the variation in sliding speed and the variation in water pressure as a function of distance along the glacier, although variations in sliding speed are predicted to correlate with basal water storage. These model predictions are in line with field data and interpretation from Columbia Glacier [*Meier et al.*, 1994; *Kamb et al.*, 1994].

Murray and Clarke [1995] developed a "black-box" model of the subglacial drainage system to explain peculiarities of the borehole water level data from Trapridge Glacier, but the concepts they developed are more widely applicable. Murray and Clarke showed that observed, time-dependent coupling between connected and unconnected boreholes could be modeled by thinking of water pressure in a connected borehole as a forcing function to which water pressure in an unconnected borehole must respond. Although their mathematical formulation was somewhat ad hoc, they argued plausibly that their model coefficients could be interpreted in terms of physical processes at the glacier bed, namely, dilation/compaction of porous subglacial sediment, diffusion of water pressure disturbances through the sediment, and uplift of the glacier from its bed. Subsequently, *Clarke* [1996] has shown that conceiving of the subglacial drainage system as consisting of linked "lumped elements," analogous to an electrical circuit, provides a powerful basis for explaining many of the complicated data collected during ~25 years of borehole studies. This approach seems to have great potential for elucidating the details of basal hydrology at relatively small spatial scales and short time periods. In this sense it complements *Humphrey*'s [1987] approach, which is directed at explaining large-scale, long time period behavior.

A key issue that needs much more thorough investigation is how the various components of the glacial drainage system interact in space and time. The system components (snow, firn, and surface streams; crevasses, moulins, and other englacial passages; and basal channels, cavities, and till) are in a state of flux throughout the year and are unevenly distributed.

GLOSSARY

Ablation: All forms of mass loss including sublimation, evaporation, melting, and calving. For alpine glaciers, the term "ablation" is often used, incorrectly, to mean melt because that is the dominant means of mass loss.

Ablation zone: The part of the glacier where yearly mass loss exceeds that gained by snow accumulation and the surface exposed in the late summer is ice.

Accumulation zone: The part of the glacier where yearly mass gain generally exceeds that lost by ablation and the surface consists of either snow or firn.

Albedo: The ratio of reflected energy flux to incident energy flux from solar radiation.

Arborescent: Tree like, used to describe a network of channels that converge as the branches of a tree converge to a trunk.

Confined aquifer: A water-bearing formation confined on the top and bottom by nearly impermeable formations.

Crevasse: A gaping crack in a glacier formed by tensile stresses resulting from glacier movement.

Englacial: Within the body of a glacier.

Equilibrium line: Line dividing the ablation and accumulation zones, where net annual mass change is zero.

Firn: A metamorphic transition stage between snow and ice. By definition, firn is seasonal snow that has survived the summer melt season.

Glacierized: Landscape currently covered by glaciers.

Glaciated: Landscape once acted upon by glaciers.

Hydraulic conductivity: A measure of the ability of a porous medium to transmit fluids.

Ice stream: Fast-moving (hundreds of meters per year) river of ice within an otherwise slow-moving (tens of meters per year) ice sheet. Ice streams are found in Antarctica and Greenland.

Intergranular: Between the grains. Used here in connection with veins to indicate small passages between the grains (crystals) of ice, specifically at the boundary where three or more grains meet.

Moulin: A natural vertical shaft in the glacier, formed from the heat of meltwater flowing into the body of the glacier.

Nonarborescent: Refers to a network that does not converge from many flow paths to a few flow paths. Paths will converge and diverge with little or no change in the number of paths.

Overdeepening: A large depression in the bedrock under a glacier. If the glacier receded, water would fill the overdeepening to form a lake.

Subglacial: At the base of the glacier.

Supraglacial: On the glacier surface.

Surging glacier: A glacier that rapidly accelerates its motion. These glaciers typically exhibit a cycle lasting decades with quiescent "normal" flow (~0.1 m/d) interrupted by one or two seasons of rapid flow (tens of meters per day).

Temperate glacier: A glacier in which the temperature at the bed and within most of the body of the ice is at its melting point. Most glaciers in the temperate zones of the Earth are temperate glaciers. A polar glacier is one that is frozen to the bed and in which meltwater does not penetrate to significant depths.

Till: Unconsolidated sediment composed of silt, sand, and cobbles, formed by glacial erosion. As used in this paper, the term does not have any sedimentological connotations.

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