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ENGLACIAL AND SUBGLACIAL HYDROLOGY: A QUALITATIVE REVIEW

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Abstract:

Modern interest in water flow through glaciers can be dated from a pair of theoretical papers published in 1972. In one of these, Shreve (1972) discussed the influence of ice pressure on the direction of water flow through and under glaciers, and in the other Röthlisberger (1972) presented a theoretical model for calculating water-pressures in subglacial conduits. Through a combination of these theoretical considerations and field observations, we conclude that the englacial drainage system probably consists of an arborescent network of passages. The millimeter-sized finger-tip tributaries of this network join downward into even larger conduits. Locally moulins provide large direct connections between the glacier surface and the bed.

Beneath a valley glacier the subglacial drainage is likely to be either in a few relatively large and comparatively straight conduits or in a system of linked-cavities. The latter results in an exceedingly tortuous flow path. Both types of drainage pattern may coexist in the same glacier. The average flow direction is controlled by a combination of ice overburden pressure and bed topography, and in general is not normal to contours of equal elevation on the bed.

Although theoretical studies usually assume that subglacial conduits are semi-circular in cross section, there are reasons for believing that this ideal is rarely realized in nature. Broad low conduits may be the rule.

When a glacier is moving over a bed of unconsolidated sediment, some water may drain through the sediment. In addition, when high water pressures weaken the sediment, it may be squeezed into subglacial channels, blocking them.

FORORD

Vannets veier gjennom og under breer har vært gjenstand for forskningsvirksomhet i mange år. Fortsatt har man ikke nådd fram til en klar forståelse på området. Dette betyr bl.a. at man ennå ikke kan trekke sikre dreneringsgrenser i breområder, og at man dermed heller ikke kan angi feltarealer med rimelig sikkerhet. Også når det gjelder vannets drenering på bunnen av breene og stabiliteten på eventuelle vannkanaler er det stor uvisshet.

Dette er bakgrunnen for at Brekontoret i 1986 tok initiativ til å få laget en state-of-the-art rapport om emnet. Dette arbeidet ble påbegynt av Bjørn Wold. Det er senere overtatt og fullført av Professor Roger LeB. Hooke fra University of Minnesota, i nært samarbeid med Bjørn Wold.

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Arne Tollan
avdelingsdirektør

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1. INTRODUCTION

A great deal has been learned about water flow in and under glaciers during the past 16 years. Much of the progress has been theoretical, as experimental techniques for studying the englacial and subglacial hydraulic system (remote sensing in the broadest sense of the word) are few, and as yet not fully exploited, and observational evidence is difficult to obtain for obvious reasons.

Two papers, one by Shreve (1972) and the other by Röthlisberger (1972) mark the beginning of this period of more active interest in the glacier hydraulic system. Shreve showed that water in the englacial part of the system could be expected to flow normal to equipotential surfaces defined by:

$$\phi = \phi_o + \varrho_w gZ + \varrho_i g(H-Z) + p(\dot{r}) \qquad (1)$$

where ϕ is the potential, ϱ_w and ϱ_i are the densities of water and ice respectively, g is the acceleration due to gravity, H and Z are the elevations of the glacier surface and of a point within or on the bed of the glacier, respectively, $\dot{r}$ is the rate of closure of passages by plastic flow of the ice, and $p(\dot{r})$ denotes a contribution to the pressure (or potential) that is a function of $\dot{r}$. The first term on the right is a reference potential, the second is the potential energy of the water due to its height above a datum such as sealevel, the third is the pressure in the water due to the overlying ice, and the fourth is a pressure difference between the water and the ice that results in closure (or opening) of the tunnel by plastic flow. In the steady state, closure of the tunnel by plastic flow exactly balances melting on the tunnel walls by viscous dissipation of energy in the water. By differentiating equation (1) with respect to a horizontal distance, assuming the gradient in $p(\dot{r})$ is small, and setting the result equal to 0, it is readily shown that the equipotential surfaces within a glacier dip upglacier with a slope of about 11 times the slope of the glacier surface (Fig. 1).

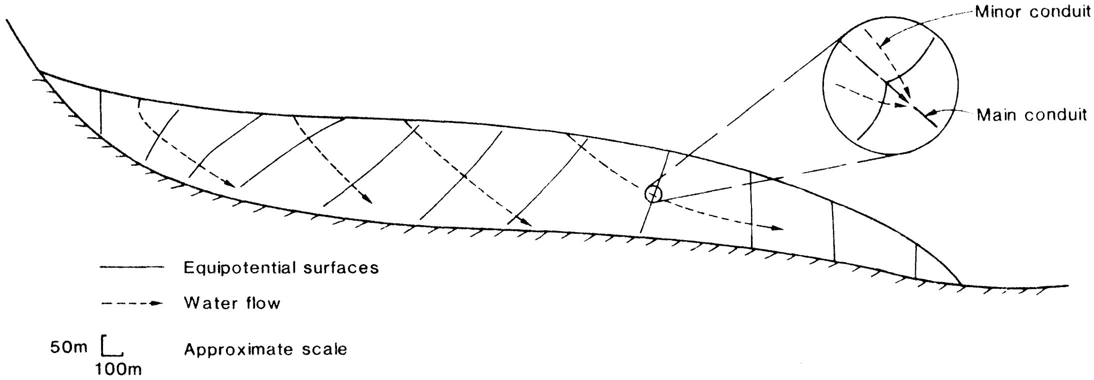


Fig. 1. Longitudinal section of a glacier showing upglacier-dipping equipotential surfaces and the theoretical directions of englacial water flow. Inset shows dimpling of an equipotential surface in the vicinity of an englacial conduit and consequent diversion of flow in smaller passages towards the conduit.

At the glacier bed, Shreve's theory predicts that water flow should be normal to equipotential contours formed by the intersection of these equipotential surfaces with the bed. This means that water may not flow in the direction of the maximum slope of the bed, but may instead flow diagonally down valley sides, or even uphill in the case of beds which slope against the direction of glacier flow.

Röthlisberger's paper deals more explicitly with the character of the flow in the hydraulic system. Whereas Shreve, for the most part, assumes that the water pressure in conduits equals the ice overburden pressure, which would mean zero tunnel closure rate, Röthlisberger bases his theory on the equality between closure rate and melt rate. Thus his equations can be used to calculate conduit sizes and hydraulic grade lines, for example, for an assumed set of discharge, ice-thickness, bed-slope, and channel-roughness conditions.

Röthlisberger's theory may be used to study the magnitude of the term involving $\dot{r}$ in Shreve's model. It turns out that $p(\dot{r})$ increases rapidly toward the glacier terminus. For a conduit with a discharge of 1 m^3/s beneath a glacier with a flat bed and a parabolic profile, $p(\dot{r})$ is 10% of the overburden pressure 10 km from the margin and reaches 50% of the overburden pressure about 2 km from the margin. Qualitatively this results in upglacier-pointing dimples in the equipotential surfaces (Fig. 1, inset). The dimples are larger for larger channels, but the dependence on channel size is weak. Water in smaller conduits flowing normal to the equipotential surfaces will thus be deflected toward existing conduits, the more so the larger the conduit. At the bed, similar dimpling will likewise tend to divert water away from the average direction of normals to the equipotential contours and towards existing channels.

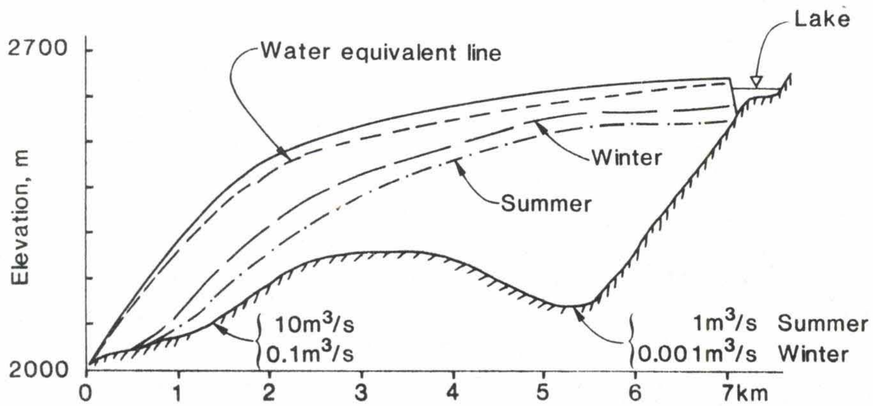


Fig. 2. Longitudinal section of Gornergletscher showing theoretically-calculated winter and summer water pressures in a conduit at the bed. Water pressure is represented by the height in the glacier to which water would rise in a piezometer. Discharge was assumed to change linearly between the two points marked with arrows and to have been constant upglacier and downglacier from those points. The water equivalent line is the height to which water must rise to float the glacier. (Modified from Röthlisberger, 1972, Fig. 8.)

Rötlisberger's theory predicts that steady-state water pressures should be higher during the winter when dicharges are lower (Fig. 2). Although perhaps contrary to expectation, this seems to be confirmed by field measurements indicating that after a period of readjustment of conduits to the reduced water discharge in winter, ice velocities, which vary directly with water pressure, increase during late winter and early spring on at least some glaciers (eg. Hodge, 1974, Fig. 13). The inverse relation between water pressure and discharge arises from the fact that melt rates on conduit walls are linearly dependent on discharge, and conduit closure rates are linearly dependent on conduit radius. Halving the discharge, for example, would halve the conduit radius, thus reducing the cross sectional area of the conduit by 75%. The water velocity would therefore have to double, and the pressure gradient (or more properly, the potential gradient) has to increase to provide this higher water velocity. When integrated back from the terminus, where the pressure is atmospheric, this leads to higher pressures throughout the conduit system. As a result of the higher pressures, the cross sectional area of the conduit is not reduced as much as just suggested, but it is constricted enough to require a higher potential gradient.

Both Shreve and Röthlisberger assumed that channels were completely filled with water. Lliboutry (1983), however, recognized that many passages might not be filled with water much of the time. Hooke (1984) studied this possibility in some detail, and summarized his findings in a graph (Fig. 3) showing that under moderate ice thicknesses on beds that sloped gently downward in the direction of flow, the energy released by even relatively small discharges would melt more ice than could be replaced by conduit closure. In this

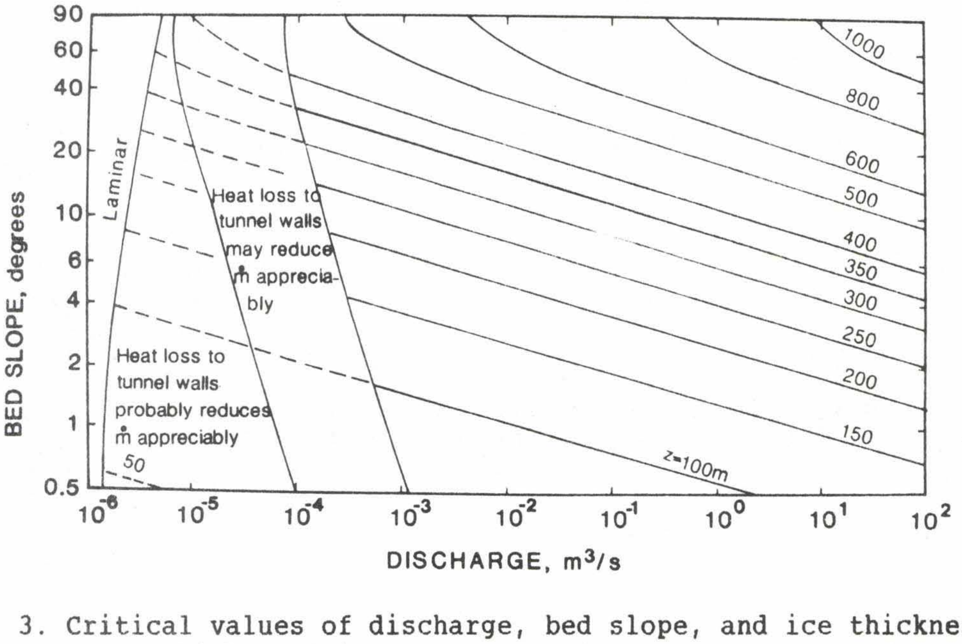


Fig. 3. Critical values of discharge, bed slope, and ice thickness, z_i at which conduits are expected to become open. If the discharge in a conduit is greater than the value read on the abscissa for a given bed-slope and ice-thickness, the conduit is likely to be open. $\dot{m}$ is the melt rate on the tunnel walls. (Modified from Hooke, 1984, Fig. 2.)

case the pressure in the conduits is atmospheric, and the direction of water flow is determined by the bed-slope and ice-flow direction only. We will return to this point in the course of the following discussion.

2. THE ENGLACIAL HYDRAULIC SYSTEM

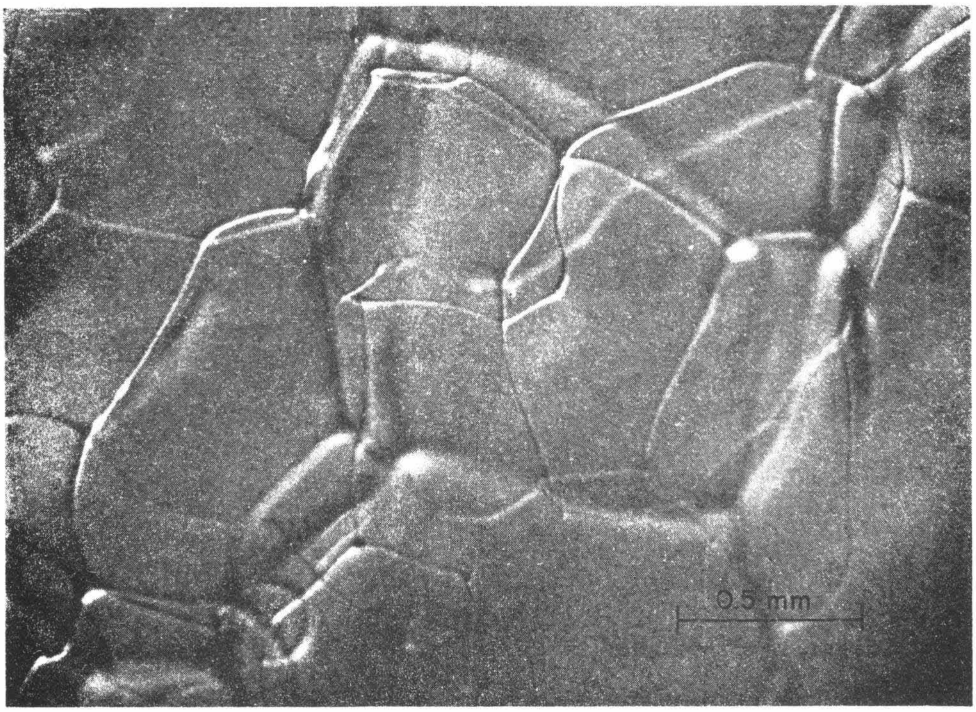
2.1 Initial development of passages

Nye and Frank (1973) have argued that veins should be present along three-grain intersections in ice, and that these veins should join together at four-grain intersections to form a network of capillary-sized tubes along which water can move. They thus concluded that temperate ice should be permeable. Raymond and Harrison (1975) observed such veins in ice samples collected from depths of up to 60 m in Blue Glacier, Washington (Fig. 4A). The estimates of the permeability given in these studies vary widely; expressed in terms of the thickness of a water layer that could be transmitted downward into the ice in a year's time, values range from ~1 mm for coarse-grained ice (Raymond and Harrison, 1975, p. 228) to 1 m (Nye and Frank, 1973, p. 160). Lliboutry (1971, p. 19) points out that the existence of supraglacial streams precludes the possibility of permeabilities significantly higher than these values, and argues further that at permeabilities near the upper end of this range and higher, the potential energy released by the descending water would rapidly enlarge the passageways to the point of completely melting the glacier. The question thus arises as to whether the veins are sufficiently well interconnected to result in appreciable permeability.

Lliboutry (1971) studied this problem and concluded that deformation and recrystallization would probably close the veins, rendering ice essentially impermeable. In laboratory experiments, Nye and Mae (1972) confirmed that temperature gradients produced by non-hydrostatic stresses could result in freezing of water in veins; the heat thus released caused melting and the formation of water-filled lenses on grain surfaces. If pervasive, this could have a significant effect on permeability. Lliboutry also considered the possibility that air bubbles might be present along veins in sufficiently high concentrations to block water movement, but concluded that this was unlikely. Raymond and Harrison (1975, p. 228) generally agreed that in fine-grained ice air bubbles would not block enough veins to appreciably reduce the permeability, but in coarse-grained ice they felt that a significant reduction was likely.

Hantz and Lliboutry (1983, p. 236) later interpreted some borehole water pressure measurements on Glacier d'Argentière as indicating that at depth fine-grained ice became slightly permeable. Although there may be other ways to interpret their measurements in particular, observations in subglacial cavities suggest that water is squeezed out of ice that is under pressure (Carol, 1947); presumably this water moves in capillary tubes, along three-grain intersections.

A



B

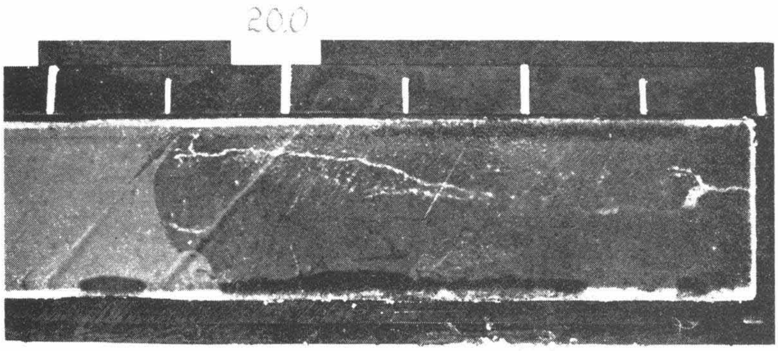


Fig. 4. A. Veins in ice from Blue Glacier, Washington (from Raymond and Harrison, 1975, Fig. 1). B. Millimeter-sized tubes in ice from a depth of 20 m in Blue Glacier, Washington (from Raymond and Harrison, 1975, Fig. 8).

Enlargement of passages by the release of potential energy is another important process to consider. Shreve (1972, p. 208-209) showed that because the ratio of discharge to wall area was larger in larger passages, such passages would increase in size at the expense of smaller ones. Röthlisberger (1972, p. 180) reached the same conclusion, using an argument based on the head loss in competing parallel channels of different size. Thus a short distance below the surface we may expect to find that some of the passages are no longer capillary veins with widths of ~25 microns (Raymond and Harrison, 1975), but instead are small tubes with dimensions of millimeters. Raymond and Harrison (1975, p. 221), in fact, observed such tubes in two of the Blue Glacier samples that they collected. In one of these, a slab about 0.6 m long obtained from a depth of 20 m, they found an upward-branching arborescent network of tubes that were up to a "few millimeters" in diameter (Fig. 4B).

2.2 Connections to the surface

In the accumulation area one can visualize continuous connections between the vein system along three-grain intersections and the overlying porous firn. As these veins normally would not be able to transmit all of the melt water percolating down through the firn, one might expect to find a local water table in the firn. The existence of such a water table has been confirmed by Vallon and others (1976) in a study on a glacier on Mont Blanc.

In the ablation area, however, there may be a near-surface layer of ice, a few meters or more in thickness, in which the temperature is some degrees below the melting point (Liestøl, 1967, p. 15; Hooke and others, 1983a). This layer forms on glaciers in more continental climates where snow fall is low enough to allow appreciable cooling of the ice by conduction of heat to the surface during the winter. It does not form on glaciers in more maritime climates where large snow falls form an effective insulating layer. When present, it is likely to persist well into the melt season, if not entirely through it, and thus forms an effective barrier to penetration of surface melt water along grain boundaries.

Crevasses form where tensile stresses exceed the tensile strength of ice, and complicate the drainage pattern. In the accumulation area they result in depressions in the water table, and in the ablation area they can transmit water through any cold layer that may exist, down to the temperate ice below.

When crevasses form in the ablation area they normally cut across the surface drainage. Surface streams running into a newly-formed crevasse may initially fill it with water. This, however, is relatively uncommon. Usually the water rather rapidly finds a way out of the crevasse, sometimes by flowing laterally to a point where the intersection between the crevasse and the glacier surface is at a lower elevation, thus returning the water to the surface, but more often through passages that lead further down into the glacier. It seems likely that in the latter cases the crevasse intersected one or more millimeter-sized englacial passages such as those observed by Raymond and Harrison (1975), and that energy released by the descending water rapidly enlarged the passage until it could accommodate all of the water flowing into the crevasse.

Crevasses do not remain open indefinitely, of course. As the tensile stress decreases, either through flow of the ice to areas of lower stress, or through seasonal changes in the stress field resulting from changes in subglacial water pressure, they close. The energy released by water in streams flowing into the crevasse is sufficient to maintain a connection with the englacial hydraulic system as the crevasse closes, however. Such a connection, called a moulin, normally persists, sometimes for more than one melt season, until another crevasse opens farther upglacier and captures the water flowing to it. Then, through a combination of slow closure by plastic flow of the ice, and infilling by snow during the winter and by water from local surface drainage during the summer, the moulin ceases to exist, at least as an open hole. However, the processes leading to its demise result in distinctive structures in the ice that can be recognized years later after ablation has lowered the surface many meters.

Through descents into moulins in the winter, using standard mountaineering techniques, and careful mapping of structures such as those mentioned above over a period of several years, Holmlund (1988) has studied the geometry of some moulins on Storglaciären, Sweden, and of the connections between these moulins and the englacial drainage system. He has found (1) that the moulins are normally 30 to 40 m deep, (2) that passages leading away from the bottoms of moulins typically meander and trend in the direction of the crevasse in which the moulin formed, (3) that when two or more moulins formed along the same crevasse, the conduits leading from them usually join together a few meters below the bottoms of the moulins, and (4) that after some distance, the meandering channels end in vertical conduits leading deeper into the glacier.

2.3 The deeper part of the englacial drainage system

Based on the above observations and theoretical considerations, we visualize the englacial drainage system as consisting of an upward-branching arborescent network of passages. This network starts near the surface with capillary tubes, which then join downward to form increasingly larger channels. At intervals, crevasses or moulins provide much larger direct connections between the surface and deeper parts of this system. The question we now face is to determine the geometry of the system beneath the level reached by Holmlund's observations.

One possibility is that at some depth the conduits begin to slope downglacier normal to the equipotential surfaces defined by equation (1), according to Shreve's theory. However, Hooke's (1984) calculations suggested that once channels reach a diameter of 3 to 4 mm, the mechanical energy dissipated by the descending water would be able to melt the walls of the conduits faster than they could close by plastic flow of the ice. Under these circumstances the water would not fill the conduit, but would instead flow on the gravitationally lowest side of it. This side would thus melt; for a conduit 3 mm in diameter and initially sloping at an angle of 45 degrees, the melt rate would be on the order of 0,5 m/a, but as the conduit diameter increases downward, the melt rate increases rapidly, reaching, for example, about 2,5 m/a for a 10 mm diameter channel. Thus such conduits should tend to become more nearly vertical. This, in turn increases the melt rate further; a 10 mm diameter conduit sloping 60 degrees will melt at a rate of about 3,5 m/a. In addition to this effect, a downglacier-sloping conduit will be gradually steepened by differential flow of the ice, as ice near the surface of a glacier flows faster than that at depth.

The processes visualized by Hooke will not occur if there are backwater effects from constrictions in the hydraulic system deeper in the glacier. In this case the flow rates in the passages near the surface depend on conditions in the vicinity of the constriction, and are independent of the size and slope of the passage near the surface. Data from boreholes drilled to study water pressure variations in glaciers may give some idea of the range of depths at which conduits become full as a result of such constrictions. In holes in which water levels fluctuate widely, and that thus are believed to have connected with the subglacial hydraulic system, water levels typically range from about 30 m to slightly more than 100 m below the surface (Hodge 1979; Hantz and Lliboutry, 1983; Iken and Bindschadler, 1986). Englacial conduits that are only partially filled with water much of the time may thus be expected within about 100 m of the surface.

3. THE SUBGLACIAL HYDRAULIC SYSTEM

As noted, the simplest model of the subglacial drainage is that channels should be normal to contours formed by the intersection of the upglacier-dipping equipotential surfaces with the bed. Studies of esker geometry demonstrate that this is undoubtedly a good first approximation for channels far from the margin beneath continental ice sheets (Shreve, 1972, 1985a, 1985b). Beneath valley glaciers, however, the geometry is complicated by the dimpling of the equipotential surfaces in the vicinity of channels mentioned above, by ice flow, by microtopography on the bed, and, where bed-slopes are appreciable and the ice is less than a few hundred meters thick, by melting of the channel walls to the point where the ice pressure no longer influences the direction of water flow. In some cases a layer of deformable till, decimeters or even meters in thickness, may be present between the sole of the glacier and the bedrock, further complicating the problem. We begin this discussion by considering the role of microtopography.

When ice flows past a bump on a glacier bed, the pressure on the stoss side of the bump is higher than the hydrostatic pressure in the ice, and that on the lee side is lower. If the ice flow is fast enough over a given bump, the ice may separate from the bed in the lee of the bump, leaving a cavity. Water may find its way into this cavity, and if the water is under some pressure, the size of the cavity will increase. Such pressurized water may also gain access to the lee sides of bumps from which the ice has not separated, and may cause separation in these areas.

In the high pressure region on the stoss side of the next bump down-glacier, such cavities, or channels leading from them, may be constricted (Lliboutry, 1983, p. 222), or even completely pinched off if the discharge is low enough (Weertman, 1986). Constriction would have the effect of increasing the potential gradient required to drive the flow, and hence water pressures in general. In the limit, in which the only sources of water were basal melting by geothermal and frictional heat, pinch-off would be likely (Weertman, 1986); in this case water pressures would rise so high that the water would spread out in a thin film under the glacier.

As long as the water supply is adequate, small channels apparently form between such cavities, leading to development of an interconnected or linked-cavity network (Fig. 5). On the average, the flow in this network should be normal to the equipotential contours, but the path is exceedingly tortuous, with long stretches in which flow is likely to be nearly parallel to the contours. We can be quite certain that such linked-cavity networks exist beneath some if not most valley glaciers, as evidence for separation is abundant on deglaciated bedrock surfaces, and evidence for the connecting channels can sometimes be found (see for example, Walder and Hallet, 1979).

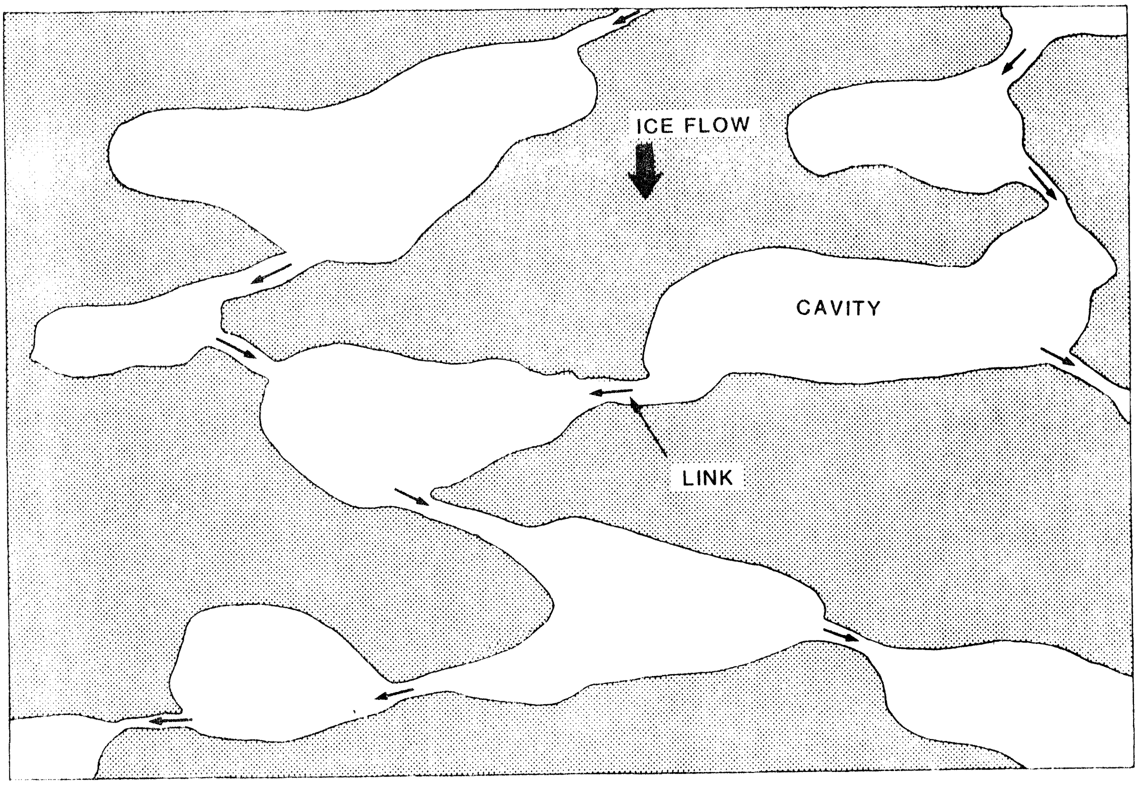


Fig. 5. Schematic drawing of a linked-cavity drainage network. Ice is in contact with bed in stippled areas. (Modified from Kamb, 1987, Fig. 1.)

Kamb (1987) has made a detailed theoretical study of the stability of such a linked-cavity system. He found that if the sliding velocity is high enough and the water pressure low enough, such a drainage system should be stable. The key to this stability is the cross-sectional shape of the links (Fig. 6), which are inferred to form in the lee of low bumps in the bed, much as larger cavities form in the lee of larger bumps, and thus tend to be oriented transverse to ice flow. With high sliding velocities, ice flow towards the roofs of these links, both from upglacier and vertically downward from above, is adequate to balance melting from viscous dissipation in the water. However, with lower sliding velocities or higher water pressures, the roofs of the links melt so fast that influx of ice cannot keep pace. The links then grow into tunnels and become more nearly parallel to the ice flow.

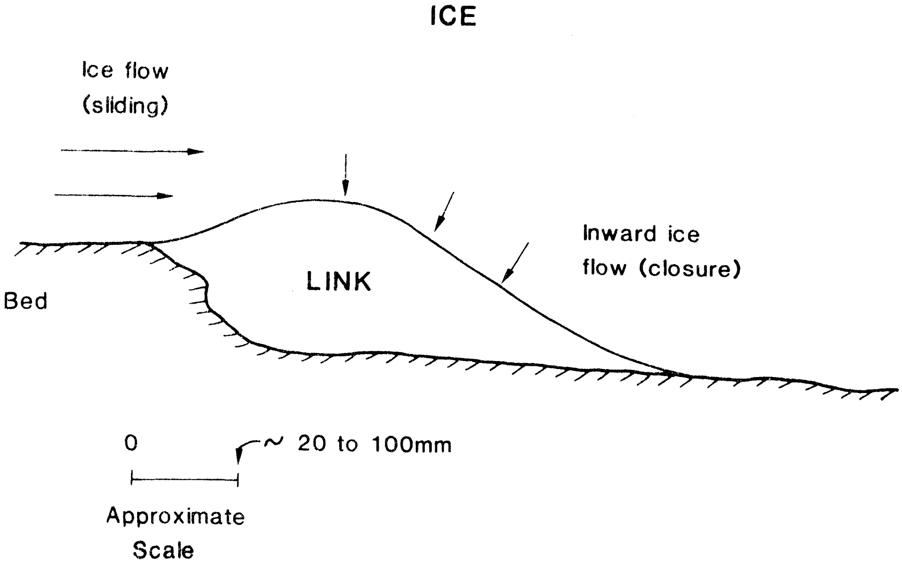


Fig. 6. Schematic cross section of a link between two cavities. (Modified from Kamb, 1987, Fig. 5.)

Contrary to the situation in a Röthlisberger tunnel system in which, as noted, water pressures decrease as discharge increases, the water pressure in a linked-cavity system varies directly with discharge. This is because the flow of ice towards the roofs of the links is such that although halving the discharge would halve the melt rate on the roof, this would result in only a small lowering of the roof. The cross-sectional area of the link thus decreases by less than a factor of two, and the lower discharge can be driven by a lower pressure (potential) gradient. This has an important consequence; whereas larger Röthlisberger tunnels have lower pressures and therefore tend to draw flow away from smaller conduits with higher pressures, the converse is true for the linked-cavity system. Thus whereas a Röthlisberger tunnel system tends to evolve into an arborescent network of channels with smaller tributaries feeding larger trunk streams, the linked-cavity system can remain anostomosing. By "anostomosing" we mean that from any given cavity there may be two or more possible links that the flow can pass through, leading to distinctly different downstream cavities, and perhaps ultimately to different outlets at the glacier terminus. A second significant difference between a Röthlisberger tunnel system and a linked-cavity system is that once a certain threshold discharge has been exceeded, pressures in the linked-cavity system are higher than in the tunnel system at the same discharge.

The conditions required for stability of a linked-cavity system may not be realized very often in nature. From numerous experiments in which tracers have been put into moulins or boreholes on glaciers, for example, we know that the velocity of water flow in the subglacial drainage system is often far higher than would be possible in such a tortuous system. Velocities of 0,2 to 0,7 m/s are common (Stenborg, 1969; Behrens and others, 1975; Burkimsher, 1983; Seaberg and others, 1988). Thus it seems clear that larger straighter channels are an important part of the drainage system of most glaciers.

If water has ready access to a linked-cavity system, increases in water pressure are expected to result in increases in sliding speed, due in part to increased separation of ice from the bed, and in part to an hydraulic jacking effect, by which water in cavities exerts a downglacier force on the downglacier-sloping roofs of the cavities (Iken, 1981). Iken and Bindschadler (1986) have studied the relation between surface velocity and subglacial water pressure in the ablation area of Findelengletscher. They recorded water pressure in several boreholes drilled to the bed of the glacier over an area measuring about 300 x 600 m, and determined the velocity of a number of stakes at the glacier surface several times a day. They found a good correlation between water pressure and velocity over a significant area, and thus inferred that water did have ready access to the cavity network.

Iken and Bindschadler (1986, p. 109) also reported, however, that dye-injection studies yielded water-flow velocities of 0,1 to 0,5 m/s, indicating the existence of relatively direct connections between the boreholes and moulins in which the dye was injected and the glacier terminus. Theoretically such channels, because they are larger, should have significantly lower water pressures than the interconnected cavity network (Röthlisberger, 1972). Bindschadler (1983) and Iken and Bindschadler (1986, p. 113) show that this should lead to rather steep gradients in water pressure as water courses in the linked-cavity system approach the main channels. This presents a problem, however, as there was very little spatial variation in water levels in the boreholes that Iken and Bindschadler monitored.

Iken and Bindschadler (1986, p. 110) used electrical conductivity measurements in proglacial streams and other observations to draw some inferences regarding differences between the subglacial drainage of Findelengletscher and nearby Gornergletscher, which is comparable in size. They concluded that the drainage of Gornergletscher is in a few large conduits, whereas that of Findelengletscher is in a large number of smaller conduits, but do not suggest any explanation for this difference. This serves to emphasize that significant differences may be found among different glaciers.

Whereas the above results suggest that some combination of the linked-cavity and tunnel systems may be the norm, there is at least one situation in which the linked-cavity system may exist alone, at least for a few months: beneath a glacier in surge. During build up to a surge, a glacier becomes steeper and its sliding velocity increases (Kamb and others, 1985). During the autumn prior to the surge, water drains out of the existing tunnel system faster than the tunnels can close, leading to low water pressures, and the low-water-pressure and high-sliding-velocity conditions required for stability of the linked-cavity system are then met. The tunnel system collapses into a linked-cavity system, and as higher pressures are required to drive the discharge through a linked-cavity system, pressures increase. This increases the sliding velocity, and the surge begins. Eventually pressures become so high that the linked-cavity system is no longer stable, and it reverts back to a tunnel system, ending the surge. This, at least, is the model put forth by Kamb (1987) to explain the 1982-3 surge of Variegated Glacier, Alaska.

In a few cases, dye-trace experiments have suggested a bilateral division of the subglacial drainage, with major conduits on either side of the glacier centerline (Stenborg, 1969). Shreve's theory provides a possible explanation for this geometry. Because transverse profiles of glaciers are normally convex upward in the ablation area, equipotential surfaces near the margins will tend to dip inward toward the glacier centerline. If the glacier bed is sufficiently flat, the intersections between these surfaces and the bed will therefore be deflected downglacier near the centerline, thus driving water away from the centerline. Hooke (1984, p. 185) also proposed a possible explanation for such a division; he suggested that as water entering a glacier along the sides penetrated to greater depths along the bed, conduit closure rates would increase until all of the energy released by the water would be required to maintain the conduit diameter, leaving none available to offset flow of the ice. In this situation, the conduits would be parallel to the ice-flow direction.

The drainage pattern is not static. It was inferred over a decade ago (Hodge, 1974) that both englacial and subglacial channels should be smallest in the spring after closing by plastic flow of the ice during the winter months when water input was low, and that these channels should increase in size rapidly during the early part of the melt season as water input increased. Water pressures, on the average, should thus be highest in the late winter when, as noted, high potential gradients are required to drive the water through the small conduit system (Röthlisberger, 1972, p. 181), and especially in the spring when water inputs begin to increase, but before the channel system becomes enlarged. Measurements of seasonal variation in surface velocity lend strong support to this hypothesis (Hodge, 1974; Hooke and others, 1983b).

As inputs increase in the spring and the conduits begin to increase in size, pressures decrease, and under suitable conditions of bedslope and ice thickness, may become atmospheric (Fig. 3). The energy dissipated by the flowing water is then more than enough to maintain the channel size required to carry the water supplied, and the excess energy can begin to offset flow of the ice. In this case conduits may migrate laterally, as appears to happen annually beneath Bondhusbreen (Hooke, Wold, and Hagen, 1984) resulting in significant changes in channel location. Such seasonal changes in channel location have also been documented by dye-trace experiments (Burkimsher, 1983; Seaberg and others, 1988).

Temporal variations in the linked-cavity network are also expected, both because the cavities expand and contract in response to variations in water pressure, and because the small channels connecting the cavities, if transverse to the ice flow direction, may be gradually carried away by ice flow (Iken and Bindschadler, 1986, p. 115). As these channels are destroyed, pressures rise in the cavities and new connections are formed.

Before concluding, it may be well to return briefly to the role of microtopography, and in particular consider how a bump, some meters in height, on a glacier bed would distort the equipotential contours. The stoss side of the bump would clearly be a zone of elevated pressure, so Φ (Eq. 1) would be higher here than in the absence of the bump. Because Φ decreases downglacier, it is clear that the gradient in Φ would be less on the stoss side of the bump than in adjacent areas, or in other words, the equipotential contours will be dimpled downglacier over the bump. Troughs in the contours will thus occur on either side of the bump, and water will be deflected into these troughs. The high pressure zone may extend far enough upglacier, however, so that the water does not flow to the lowest side of the bump. Alternatively, if a trough between two laterally-adjacent bumps is relatively narrow and there is a strong convergence of the ice flow toward this trough, pressures may develop here that are higher than those on the stoss sides of the bumps. In this case troughs in the equipotential contours may develop over the bumps, and water flow may be over rather than around them. Unfortunately analytical techniques capable of dealing with such three-dimensional pressure distributions in a non-Newtonian fluid like ice are only just being developed, so more quantitative analyses are presently impossible.

3.1 Conduit shape

Heretofore we have tacitly if not explicitly assumed that englacial conduits were circular in cross section and subglacial ones were approximated as either circular or semi-circular. There is reason to believe that in many cases the latter approximation is a poor one. Shoemaker (in prep) for example, has suggested that drag on the bed will inhibit closure of semi-circular tunnels, and J. Kohler (unpublished) is studying the effect of low discharges on the shape of conduits formed under higher discharges.

The case of bed drag is easy to visualize, but difficult to analyze quantitatively. Basically, the theory of closure of a semi-circular tunnel assumes that there is no drag on the bed. It is clear, however, that there is always some drag, and that as a result the inward flow of ice towards a tunnel will be less along the bed than it will be higher on conduit walls. The amount of drag varies considerably depending on the roughness of the bed. It will be low on smooth bedrock beds and particularly high on beds composed of heterogeneous till if water pressures in the till are low enough to ensure that the till does not deform. As a result of such drag, melt will exceed closure low on the tunnel walls if the melting is distributed more-or-less uniformly over the tunnel cross section as is usually assumed. Thus the tunnel will tend to become broad and low.

Similarly, if the discharge varies so that a conduit is alternately full and partially full (or open), melting during the latter period will be concentrated along the lower parts of the walls. This again will lead to broad low cross-sectional shapes. Variations in discharge are to be expected in conduits that receive a large part of their flow from the surface where temporal variations in melt rate and rainfall control the water supply. Whether such variations will lead to open conduits will depend, in part, on ice thickness; under thicker ice, closure rates may be sufficiently high that water pressures rarely drop to atmospheric.

The problems encountered in a theoretical study of conduit shapes are legion. First, the closure rate is no longer uniform around the periphery of the tunnel. Shoemaker (in prep) has used crack theory to approach this problem, but then needs to assume a linear rheology for ice. Whether such an approach will eventually prove feasible is uncertain. Secondly, drag on the bed would be difficult to predict, even if bed roughness were known, which it is not in most cases. Thirdly, melting is probably not uniform around the periphery of even a semi-circular tunnel, and furthermore the heat used for melting is produced some distance up-stream from the point where the melting occurs. Existing theory generally assumes that melting occurs where the heat is generated, and is greatest where water is deepest. These assumptions, together with arguements based on symmetry, lead to a semi-circular cross section, other things being equal. How reality differs from this model is unknown. Finally, discharge variations are weather dependent, so channel shape may be also.

Perhaps this section should be concluded with observation that channels emerging from glacier termini are broad and low. Thus the question can be posed: "How far upglacier does this conduit shape persist?"

3.2 Subglacial till layers

A further complication may arise if there is a layer of deformable till between the sole of the glacier and the bedrock. When water pressures are low, such till may be squeezed up into any channels or cavities that were developing between the ice and the bed, constricting them severely. There is increasing evidence for the existence of such deformable till beds (Hodge, 1979, p. 315; Haeberli and Fisch, 1984; Brand and others, 1987). In at least two of the documented cases, such till was found in overdeepenings in the glacier bed (Hodge 1979; Brand and others, 1987). In one of these cases (Brand and others, 1987) tracer tests have suggested that water-flow through the overdeepening is restricted to englacial channels (Hooke and others, 1988), that water-flow velocities are generally lower than those downglacier from the overdeepening, and that dispersivity is higher (Wiberg, unpublished). Despite these low water velocities, the velocity of the overlying ice responds to changes in water input which are inferred to result in changes in subglacial water pressure (Hooke and others, 1983b; Hooke and others, in preparation).

4. CONCLUSIONS

We have reviewed the state of knowledge of the englacial and subglacial drainage systems. The processes in the englacial drainage system are obviously complex and at present it is impossible to predict the geometry of this part of the hydrologic system more precisely. We can speculate that conduits that are less than a year old and that are not connected directly to moulins at the surface are likely to be normal to equipotential planes which, on the average, dip upglacier at about 11 times the slope of the glacier surface. Conduits that are more than a year old, and particularly those in areas of gentle glacier-surface slope (where equipotential surfaces should be more nearly horizontal) and those within about 100 m of the surface may well slope more steeply than normals to the equipotential surfaces, having been steepened by differential melting in channels at atmospheric pressure. In the limit such conduits would be vertical, or might even dip upglacier if they were deformed sufficiently by ice flow.

The character of the subglacial drainage system is at least equally unpredictable. It seems fairly certain that in areas where the ice is in direct contact with a horizontal or downglacier-sloping, non-deformable, undulating bed, a tortuous linked-cavity system will be present, transected by and in hydraulic content with one to several larger straighter conduits. It is also clear that the positions of some or all of these conduits may change seasonally and, from time to time, permanently (Hantz and Lliboutry, 1983, p. 229). The latter changes are apparently due to secular changes in thickness and surface slope of the glacier that are sufficient to change the positions of the equipotential contours. In general, however, the average trend of conduits should be normal to the equipotential contours on the bed. In areas of deformable bed, the subglacial drainage may be either through a permeable layer of till, through a large number of quite small subglacial passages, or through englacial conduits.

5. DIRECTIONS FOR FUTURE RESEARCH

While the problems that we face in improving our understanding of the englacial and subglacial hydraulic systems are clear, the approaches that should be taken are not. Some of the experiments that might be undertaken will be expensive, and others require as yet undeveloped remote-sensing techniques.

Dye-trace experiments yield information on the average properties of the hydraulic system. They are straight forward and relatively inexpensive, and can be undertaken on any glacier for which specific information is required. The experiments should include multiple injections at the same point to monitor temporal changes in the hydraulic system during the year, and injections at different points to study spatial variations. Sampling in the outlet streams must be frequent enough to define the shape of the concentration-time curve accurately, and analysis of the data should include determination of the dispersivty in order to study its temporal and spatial variation. Discharges should be monitored in the outlet streams during tests, and used to estimate the percentage of the dye recovered. Care must be taken to account for virtually all of the dye from one test before the next is started, as otherwise interpretation may be highly ambiguous. Such tests have the potential for contributing significantly to our basic understanding of the hydraulic system.

Detailed studies of the relation between velocity and water pressure, such as that carried out by Iken and Bindschadler (1986) are also valuable, but they require a major effort and the theoretical basis for their interpretation is still weak. Such studies may, however, point out significant differences between glaciers, such as those between Findelengletscher and Gornergletscher mentioned above.

An effort should be made to develop a neutrally-bouyant electronic "fish" that can be sent down a moulin or borehole and that can transmit information on its position to the surface. With such a sensor one could map at least parts of the englacial and subglacial hydraulic systems. Dr. Michael Walford of the University of Bristol, England, is undertaking such a project with NVE support.

On a more ambitious level, detailed experiments need to be undertaken from subglacial laboratories, such as that under Bondhusbreen (Hagen and others, 1983) or that planned for Engabreen. With proper facilities, one could, for example, build an artificial glacier bed with numerous pressure and temperature sensors and with viewing ports that would permit direct observation of some of the processes that we believe occur in the linked-cavity network. This would provide basic data needed for theoretical studies of the growth and collapse of cavities and of connections between cavities, such as those discussed by Kamb (1987).

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