167: Subglacial Drainage

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Subglacial drainage can occur wherever ice at a glacier bed reaches the pressure melting point. The subglacial drainage system is fed from a mixture of surface, englacial, subglacial, and groundwater sources that differ in terms of their spatial distribution and characteristic patterns of temporal variability. Subglacial drainage systems are not readily accessible, and knowledge of their characteristics is derived from a range of indirect methods including radio-echo sounding, the use of artificial tracers, monitoring, and manipulation of subglacial conditions via boreholes, and monitoring of glacial runoff properties. Subglacial water flow is driven by gradients in hydraulic potential, and occurs through either fast/channelized or slow/distributed systems located at the ice-bed interface, or through subglacial aquifers. Water can be stored subglacially in cavities, in the pore space of subglacial sediments, or in subglacial lakes. Drainage system structure evolves continually on various timescales in response to changing water inputs, evolving glacier geometry, and changes in glacier flow dynamics. Major hydrological events in such systems include the spring and fall transitions, outburst floods, and structural changes related to glacier advances and glacier surging. In the absence of variable water inputs from the glacier surface, subglacial drainage systems may be sensitive to forcing by earth, ocean, and atmospheric tides.

INTRODUCTION

Whenever ice at the base of a glacier or ice sheet reaches the pressure melting point, water may be present and a subglacial drainage system can develop. Subglacial water flow is of interest for a number of reasons, including its role in modulating glacier runoff, its potential as a source of geo-hazards, its influence on ice flow dynamics, and its role in chemical weathering (*see* **Chapter 168, Hydrology of Glacierized Basins, Volume 4** and **Chapter 169, Sediment and Solute Transport in Glacial Meltwater Streams, Volume 4**). If meltwater produced at the glacier surface penetrates to the glacier bed, the rate at which it is transported to the glacier terminus is determined in part by the character of the subglacial drainage system, which therefore mediates the relationship between surface melting and runoff. This is an important issue where glacier runoff has economic value (e.g. for the production of hydroelectricity) since efficient harvesting of the resource depends upon the ability to forecast runoff (Willis and Bonvin, 1995). Subglacial drainage systems may receive sudden inputs of large volumes of water from the drainage of supraglacial, ice-marginal, and subglacial lakes (Nye, 1976). The escape of this water from the glacier can be a significant hazard to population centers. Subglacial drainage can also exert a strong influence on glacier flow dynamics. Basal water reduces the friction at the ice-bed interface, affecting the rate of glacier flow by sliding (Iken, 1981) and the strength and deformation rate of unconsolidated subglacial sediments (Clarke, 1987a). Spatial variations in subglacial drainage system character may help explain the location of ice streams within large ice sheets (Bentley, 1987), and temporal variations have been implicated in the mechanics of glacier surging (Kamb, 1987). Subglacial water has access to abundant supplies of freshly ground and highly reactive "rock flour", creating a unique chemical weathering environment (Raiswell, 1984; Tranter *et al.*, 1993) (*see* **Chapter 169, Sediment and Solute Transport in Glacial Meltwater Streams, Volume 4**). The investigation of subglacial drainage has therefore become a major research focus in glacier hydrology.

SOURCES, SINKS, AND DIRECTION OF SUBGLACIAL DRAINAGE

Water Sources

Subglacial water has four main sources: surface, englacial, and basal melt, and subglacial groundwater. The relative

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importance of these sources depends upon the climatic regime at the ice surface, the temperature of the glacier ice, ice flow dynamics and the nature of the glacier bed. On annual timescales, surface melt is usually the dominant source in "temperate" glaciers, where ice is almost all at the pressure melting point. Basal melt may, however, be the dominant subglacial water source in large areas of Antarctica where surface air temperatures never rise above the freezing point but ice temperatures reach the pressure melting point near the glacier bed. It may also be important in areas of high geothermal heat flux, such as Iceland. Englacial melt due to strain heating may be a significant water source in fast flowing glaciers such as Jakobshavn Isbrae, Greenland. Groundwater inputs have received little study, but they may be significant where a glacier rests on an underlying aquifer, especially in winter when surface melt ceases.

The Distribution and Magnitude of Water Inputs

The distribution of water inputs to a glacier bed depends upon the water source and thermal structure of the glacier. Basal melting (driven by geothermal heat and by frictional heat produced as the glacier slides over its bed) occurs throughout warm-based areas of a glacier, where ice at the bed is at the pressure melting temperature. This usually results in a relatively uniform distribution of melt inputs, because spatial variations in the geothermal heat flux usually occur on length scales greater than the size of individual glaciers. In some areas, like Iceland, however, there can be large spatial variations in the basal melt rate related to the distribution of active geothermal areas. There can also be significant spatial gradients in basal melt rate due to sliding friction or strain heating. Water produced within the glacier may drain to the glacier bed via a network of veins located at crystal boundaries in temperate ice (Nye and Frank, 1973). This also results in a relatively uniform distribution of water inputs to the bed. Most surface melt, however, becomes channelized subaerially and penetrates to the bed via crevasses or vertical shafts known as *moulins* (Holmlund, 1988) (*see* **Chapter 166, Surface and Englacial Drainage of Glaciers and Ice Sheets, Volume 4**). These localized inputs of water may be restricted to limited regions of the bed, depending upon the distribution and density of crevasses and moulins. This is especially true for nontemperate glaciers, where crevasses are often rare due to low rates of ice flow (Hodgkins, 1997).

The magnitude and variability of surface water inputs will depend on the density and exposure of input sites and on the nature of the melting surface. Surface snow and firn reservoirs have considerable potential to store water, and they act to delay melt-induced runoff and damp diurnal and meteorologically driven variations in meltwater flux (Fountain, 1996). These reservoirs produce slowly varying water inputs. Ice surfaces have comparatively little storage capacity, such that water inputs to the glacier bed from bare ice surfaces track the surface melt rate (Willis *et al.*, 2002). An exception to this generalization occurs on some cold glaciers, where there can be significant surface storage in lakes and channel systems (Liestøl *et al.*, 1980). This stored water may drain rapidly to the glacier bed as a result of water-pressure induced propagation of water-filled crevasses (Van der Veen, 1998; Boon and Sharp, 2003).

The Location and Direction of Subglacial Drainage

Subglacial drainage can occur either at the interface between ice and the underlying substrate, or within subglacial sediments. The pattern of water flow is influenced by the topography of the subglacial surface, and by horizontal and vertical gradients in water pressure. In general, this pattern is strongly influenced by the geometry of the overlying ice because basal water often supports a substantial fraction of the ice overburden. Where the substrate is permeable, vertical gradients in water-pressure influence the magnitude and sign of water exchange between substrate and ice-substrate interface. For the case of a glacier resting on an impermeable substrate, water flow will be perpendicular to contours of equal hydraulic potential at the glacier bed. One simplified view of subglacial drainage assumes that the drainage system is a hydraulically connected water sheet, in which water pressure is equal to the ice overburden pressure (Shreve, 1972). In this case, the hydraulic potential can be written:

$$\Phi = \Phi_0 + \rho_I g(z_s - z_b) + \rho_w g\ z_b \tag{1}$$

Where Φ_0 is a reference potential, ρ_I and ρ_w are the densities of ice and water, g is the acceleration due to gravity, and z_s and z_b are the elevations of the ice surface and glacier bed, respectively. Shreve's model predicts that the slope of subglacial equipotential surfaces is $\sim$11 times the ice surface slope, but in the opposite direction. This implies that subglacial water can flow uphill out of over-deepened sections of glacier bed, so long as the magnitude of the water-pressure gradient exceeds that of the elevation potential gradient. Bedrock obstacles that are extensive in a direction transverse to ice flow will, however, likely cause substantial lateral diversions of water flow (Flowers and Clarke, 2002a), and water may refreeze if it emerges supercooled and quickly enough from over-deepenings (Alley *et al.*, 1998). Subglacial equipotential surfaces may contain depressions that define a closed interior drainage basin and become a focus of drainage, allowing the formation of subglacial lakes (Björnsson, 2002). Such depressions may be associated with localized sources of basal melting such as volcanoes. Changes in glacier geometry change the form of the subglacial hydraulic potential surface, allowing drainage capture and restructuring of subglacial drainage catchments (Fountain and Vaughn, 1995). The assumption on which Shreve's formulation of the hydraulic potential is based may, however, break down in practice because subglacial drainage systems do not consist of hydraulically connected sheets. In reality, they contain discrete elements, such as channels, in which water pressure may differ from the ice overburden pressure and vary with discharge (Röthlisberger, 1972; Walder, 1986).

THE INVESTIGATION OF SUBGLACIAL DRAINAGE SYSTEMS

Direct observations of active subglacial drainage systems are limited because these systems are typically located beneath tens of meters to kilometers of glacier ice. A variety of remote sensing and other approaches has therefore been employed to determine their character and behavior.

Characteristics of Recently Deglaciated Surfaces

Observations of recently deglaciated glacier beds have shown that glaciers can rest on both rigid bedrock and unconsolidated, permeable, and potentially deformable sediments that vary substantially in terms of key hydrologic parameters such as thickness, continuity, hydraulic conductivity, and porosity. Studies of deglaciated carbonate surfaces (Figure 1) have provided information on the size, morphology, density, and network characteristics of formerly subglacial channel systems eroded into bedrock (Walder and Hallet, 1979; Hallet and Anderson, 1980; Sharp *et al.*, 1989). They have provided evidence for extensive ice-bed separation (cavity formation) on the down-glacier side of bedrock bumps and steps (Figure 1), and revealed the widespread presence of laminated crusts composed of secondary calcite (Figure 1). These crusts are typically found on the lee side of small bedrock protuberances and are believed to have precipitated from a thin water film as a result of either refreezing of film waters or degassing of CO_2 rich solutions in the low-pressure region downstream of bed obstacles. The maximum size ($\sim 50\,\mu$m) of bedrock fragments trapped within individual laminae in the crusts sets an upper limit on the likely thickness of the water film (Hallet, 1976, 1979).

Such observations have motivated the development of several theoretical models of subglacial drainage (Walder, 1982, 1986; Kamb, 1987; Flowers and Clarke, 2002a,b). They are, however, limited in that they provide a temporally integrated view of drainage system structure and little insight into interactions between the various components of the drainage system (e.g. channels, cavities, film). Furthermore, the high solubility of carbonate minerals may mean that the structure of drainage systems developed on carbonate substrates is very different to that of systems developed on more resistant rock types.

Figure 1 Recently deglaciated carbonate bedrock surface at Glacier de Tsanfleuron, Switzerland, showing: (1) Roughened surface associated with an area of ice-bed separation (or cavity) on the downstream side of a bedrock obstacle. (2) Calcite deposits precipitated from the subglacial regelation water film. (3) A small Nye or "N" channel with solution scallops on its walls. (4) A polished and striated surface indicative of intimate ice-bedrock contact. A color version of this image is available at http://www.mrw.interscience.wiley.com/ehs

Radio-echo Sounding

Given the inaccessibility of active glacier beds, there has been considerable interest in remote sensing approaches to mapping their characteristics. Radio-echo sounding (radar) has been used to map the distribution of warm and cold-based ice using variations in the strength of the radar echo from the bed. The echo strength depends upon the dielectric contrast between ice and substrate, and is much greater for an ice-water interface than for an interface between ice and rock or dry, unfrozen sediment (Bentley *et al.*, 1998). Within warm-based areas of a glacier, the spatial pattern of bed reflection power may be indicative of the relative abundance of water at the bed (Copland and Sharp, 2001). To date, attempts to locate and map the distribution of individual channels at the glacier bed have only been successful in relatively thin ice (Moorman and Michel, 2000).

Radar has played a key role in mapping the size, shape, and distribution of subglacial lakes in Antarctica (Robin *et al.*, 1970; Oswald and Robin, 1973). Lakes are identifiable on radar records from a very strong, mirrorlike bed reflection that is constant in strength along track, suggesting a very smooth basal interface. In Iceland, radar mapping of ice thickness and bedrock topography has allowed the reconstruction of subglacial equipotential surfaces beneath ice caps (Björnsson, 1982). Subglacial lakes, such as that in the Grimsvötn caldera beneath Vatnajökull, are located within internal drainage basins that can be identified from these surfaces.

Investigations with Artificial Tracers

Various tracers have been used to characterize subglacial drainage systems. These include salt solutions and fluorescent dyes, such as rhodamine or fluorescein (Burkimsher, 1983; Kamb *et al.*, 1985; Seaberg *et al.*, 1988; Willis *et al.*, 1990; Hock and Hooke, 1993; Nienow *et al.*, 1998). Where multiple streams enter and leave a glacier, the drainage catchment structure of the glacier can be mapped by delineating those areas of the glacier surface from which tracer inputs are recovered in each individual outflow stream (Sharp *et al.*, 1993). Measures of drainage system efficiency include the time of travel from input point to the glacier terminus, and the travel distance divided by the travel time (an index of flow velocity). If the discharge during the tracer test is known, the ratio of discharge to tracer velocity provides an index of the cross-sectional area of the drainage system. Since neither the channel sinuosity nor the water fluxes through individual channels are usually known, however, it is not possible to derive true measures of velocity or cross-sectional area. Nevertheless, the relationship between the indices of velocity and cross-sectional area may provide insight into the dynamics of the drainage system because the timescale of geometric adjustment of channels can be longer than that of water input variations. Thus, in a permanently water-filled conduit, variations in water flux on short timescales would be accommodated almost entirely by variations in flow velocity, while cross-sectional area would be relatively constant. By contrast, in a channel that was not full at low flows, the flow cross section would increase as discharge rose. Water flux increases would therefore be accommodated by a combination of increases in flow velocity and the cross-sectional area of the flow, resulting in a less sensitive relationship between flow velocity and discharge (Nienow *et al.*, 1996).

Continuous monitoring of dye concentrations in an outflow stream allows the construction of a dye return curve. This provides additional information about the character of the drainage system through which the tracer has passed. Curves with a single well-defined peak indicate that most dye was advected through a single major channel, whereas multipeaked curves are indicative of a more complex, multipath configuration (Willis *et al.*, 1990). Often, there is a correlation between peak shape and transit velocity, with single-peaked curves being associated with rapid transport and multipeaked curves with slower transport. Integration of the dye flux as a function of time allows calculation of the fraction of injected tracer that was recovered at the point of outflow. Typically, this fraction is much less than 1, indicating either significant retention of tracer within the glacier drainage system, or that much of the tracer emerged at concentrations below the fluorimetric detection limit. Retention may reflect diversion of water into subglacial storage or sorption of dye to fine-grained sediment (Bencala *et al.*, 1983).

As a tracer cloud passes through a drainage system, it becomes dispersed due to vertical and horizontal variations in water velocity and diversion of packets of tracer into storage sites such as backwater eddies. Dispersion results in reduced peak heights and increased peak widths. The rate of dispersion is described by the *dispersion coefficient*, D ($m^2\,s^{-1}$), but a more useful descriptor of peak characteristics is the *dispersivity* d(m), which is the ratio of the dispersion coefficient to the tracer velocity (Fischer, 1968). The dispersivity describes the rate of peak spreading relative to the rate of peak advection, and provides a characteristic length scale for the drainage system. Flow through major channels typically results in $d < {\sim}10$ m, while more complex, less efficient systems result in much higher values ($>{\sim}50$ m) (Seaberg *et al.*, 1988; Nienow *et al.*, 1998).

Borehole Manipulation and Monitoring

Boreholes, drilled with high-pressure hot water, provide direct access to glacier beds and are routinely employed in investigations of subglacial drainage. This has allowed measurements of the minimum thickness of subglacial sediment layers (by penetrometry; Hooke *et al.*, 1997; Harbor *et al.*, 1997), and *in situ* estimation of parameters that are important determinants of the hydraulic behavior of these sediments. These parameters include the porosity, hydraulic diffusivity and conductivity, and compressibility of the sediments. The techniques employed to measure them include analysis of the horizontal and vertical propagation and attenuation of natural pressure waves through subglacial sediments (Fountain, 1994; Hubbard *et al.*, 1995; Fischer *et al.*, 1998) and impulse-response tests (Stone and Clarke, 1993; Stone *et al.*, 1997; Kulessa and Murray, 2003).

Boreholes also provide an opportunity to measure water pressure and water quality within subglacial drainage systems. Boreholes sealed by freezing allow direct measurement of water pressure in the subglacial drainage system connected to the borehole. Water levels in open boreholes have also been used as a measure of subglacial water pressure on the assumption that the borehole functions as a manometer. Although there are potential problems with this assumption, a number of studies have recorded spatially coherent patterns of water level variation across arrays of open boreholes, suggesting that at least some boreholes do behave approximately as manometers (Hubbard and Nienow, 1997).

The behavior of borehole water levels during and after the time when the drill contacts the glacier bed has been used to characterize different morphologies of the subglacial drainage system. Boreholes differ in terms of their base water levels, the amplitude, and timescale of water level oscillations, and the phase relationship between these oscillations and glacier runoff (Murray and Clarke, 1995; Hubbard *et al.*, 1995; Smart, 1996; Gordon *et al.*, 1998). Boreholes that are close to drainage channels often display low minimum water levels and high amplitude diurnal water level fluctuations that are almost in phase with runoff. Boreholes sampling hydraulically resistant elements of the drainage system typically exhibit high background water levels with minimal variability. Water level fluctuations in such boreholes may be out of phase with those in boreholes located close to channels, indicating diurnal transfer of mechanical support for the glacier between areas close to and remote from major channels as channel water pressure varies (Murray and Clarke, 1995; Gordon *et al.*, 1998).

In situ measurements of basal water quality have focused on the electrical conductivity (EC) and turbidity of borehole waters (Stone *et al.*, 1993; Hubbard *et al.*, 1995; Stone and Clarke, 1996; Kavanaugh and Clarke, 2001). High conductivity is usually associated with waters that have a long subglacial residence time, while low conductivity and low turbidity are associated with surface meltwater. High turbidity indicates basal water and/or high flow velocities. Temporal variations in conductivity and turbidity can thus provide insight into interactions between drainage system components (Hubbard *et al.*, 1995) and processes of drainage reorganization (Gordon *et al.*, 1998; Kavanaugh and Clarke, 2001). For open boreholes in particular, interpretation of measurements made at the base of boreholes demands an understanding of the system geometry and patterns of water exchange between the borehole and the subglacial drainage system (Gordon *et al.*, 2001). There have also been attempts at *in situ* hydrochemical characterization of borehole waters (Tranter *et al.*, 1997, 2002). Such measurements provide an alternative means of identifying different components of the subglacial drainage system, on the assumption that different modal water chemistries reflect differences in parameters such as the degree of water-rock contact and access to supplies of dissolved gases.

Properties of Glacial Runoff

Several studies have used the characteristics of glacial runoff to make deductions about subglacial drainage systems. The form of the diurnal discharge hydrograph from a glacier changes systematically over the melt season (Röthlisberger and Lang, 1987). Advances in the timing of the daily discharge peak and increases in the amplitude of the diurnal cycle reflect increased efficiency of water transfer through the glacier (Richards *et al.*, 1996). In part, this is a consequence of removal of the supraglacial snowpack and its delaying effect on runoff of surface meltwater, but it may also reflect the development of large drainage channels at the glacier bed. In some cases, the two processes may be closely connected (Nienow *et al.*, 1998) (*see* **Chapter 168, Hydrology of Glacierized Basins, Volume 4**).

Glacial runoff typically shows strong seasonal variations in solute concentration and composition (Tranter *et al.*, 1993). Solute concentrations in water leaving glaciers are enhanced relative to concentrations in the snowpack and supraglacial streams, indicating significant solute acquisition from water-rock interaction at or near the glacier bed. The nature and extent of this interaction depend upon the duration of water-rock contact, the rock:water ratio in the drainage environment, the extent to which water has access to supplies of gaseous O_2 and CO_2, and the availability of other potential sources of protons such as organic carbon and sulfide minerals (Raiswell, 1984; Tranter *et al.*, 1993). Glacial runoff appears to acquire solute in two distinct types of environment. Environments characterized by long transit times, high rock:water ratios, and proton supply from oxidation of sulfides and/or organic carbon result in water with high solute concentrations. This water type dominates runoff in the winter, the early melt season, and periods of recession flow in summer, and may also be characteristic of sudden outbursts of subglacially stored water at any stage of the melt season (Anderson *et al.*, 1999; Wadham *et al.*, 2001). Such waters are inferred to have drained via a predominantly distributed drainage system (see below; Tranter *et al.*, 1993). More dilute waters that have acquired significant amounts of solute by rapid reactions with suspended sediment are characteristic of periods of high flow later in the melt season (Brown *et al.*, 1994). These waters appear to have drained primarily and rapidly through large subglacial channels, though they may have mixed with and diluted waters of the more concentrated type in the process (*see* **Chapter 169, Sediment and Solute Transport in Glacial Meltwater Streams, Volume 4**).

The suspended sediment content of meltwaters also provides insight into subglacial drainage system structure. Waters that pass slowly through distributed drainage systems are unable to entrain significant amounts of sediment even though it may be abundant within the drainage environment. Water passing through large channels may have more limited contact with sediment but is better able to entrain what is available. Thus, sediment evacuation increases as channelized drainage becomes established in summer, but sediment exhaustion may occur towards the end of the melt season (Swift *et al.*, 2002). Sudden releases of sediment may result from outburst events and reorganization of drainage systems at the glacier bed (Humphrey and Raymond, 1994; Wadham *et al.*, 2001) (*see* **Chapter 169, Sediment and Solute Transport in Glacial Meltwater Streams, Volume 4**).

TYPES OF SUBGLACIAL DRAINAGE SYSTEM

Two types of subglacial drainage system have been widely recognized, though each can assume a diversity of forms. These have been characterized as *distributed* or *slow* systems, and *channelized* or *fast* systems (Fountain and Walder, 1998; Raymond *et al.*, 1995). In addition, some glaciers rest on permeable sediments that function as a groundwater aquifer (Boulton *et al.*, 1995; Flowers and Clarke, 2002b).

In distributed/slow systems, drainage pathways are tortuous, anastomosing, and widely distributed across the bed. Such systems are hydraulically resistant, and changes in water flux result in large changes in flow cross section but little change in flow velocity, which is generally low. By contrast, in channelized/fast systems most of the water flux passes through a few major channels that have an arborescent structure and occupy only a small fraction of the bed. Such systems offer much less resistance to water flow, and changes in water flux result in major changes in flow velocity. In the steady state, distributed systems exhibit an inverse relationship between water pressure and discharge, while the reverse is true for channelized systems (Walder, 1986; Kamb, 1987; Walder and Fowler, 1994; Röthlisberger, 1972). Both drainage morphologies can exist beneath the same glacier, and they will interact with each other (Hubbard *et al.*, 1995; Alley, 1996). The relative importance of the two systems in transmitting the imposed water flux may change on timescales ranging from hours to decades (Kamb *et al.*, 1985; Nienow *et al.*, 1998).

Distributed/Slow Subglacial Drainage Systems

One of the earliest configurations proposed for a subglacial drainage system was a pervasive *water film* (Weertman, 1964). Such a film might transport water produced at the bed by geothermal heating and sliding friction, or waters involved in regelation (pressure-induced phase change) around small obstacles on the glacier bed (Hallet, 1979). Weertman (1972) argued that the film might be the primary means of subglacial drainage because the pressure distribution around incipient channels at the glacier bed would impede their ability to capture water from the film. Since this argument only holds for a bed that is planar, impermeable, and free of debris, however, it is probably not applicable in reality. Thin water films are inherently unstable in the face of perturbations in film thickness because rates of viscous energy dissipation increase with film thickness, enhancing the local melt rate of the overlying ice, and causing a tendency towards channelization of the flow (Nye, 1976; Walder, 1982). Thus, while films are likely components of many subglacial drainage systems, their role in water transport is probably minor.

Where a glacier slides over a rough substrate, ice may separate from the bed to form cavities downstream of bedrock obstacles (Lliboutry, 1968) (Figure 1). Two types of cavity have been recognized: *autonomous cavities*, which contain ponded meltwater but transmit no water flux, and *interconnected cavities* which form an active part of the subglacial drainage system (Lliboutry, 1976). The status of individual cavities probably changes over time in response to changes in the imposed water flux and ice dynamics (Iken and Truffer, 1997). Nonarborescent cavity networks, linked by small bedrock (or *Nye*) channels oriented parallel to ice flow and incised up to 0.2 m into the glacier bed, have been mapped on recently deglaciated bedrock surfaces (Figure 1) (Walder and Hallet, 1979; Hallet and Anderson, 1980; Sharp *et al.*, 1989). Analyses of the hydraulics of linked-cavity systems demonstrate that cavities open by sliding of the overlying ice and close by ice deformation (Walder, 1986; Kamb, 1987). Melting of the ice roof by energy dissipated by flowing water makes only a minor contribution to the enlargement of cavities. The major stimulus to cavity growth is a discharge-induced rise in water pressure within the cavity system. The positive relationship between water flux and water pressure in cavity systems (and other forms of distributed system) has the consequence that water tends to be at higher pressure in larger cavities than in smaller cavities. It thus tends to drain from large to small cavities. This equalizes water pressures and ensures that there is no tendency for master channels to develop within the cavity network, thus acting as a negative feedback on the development of efficient (fast) drainage systems. The tendency for cavities to grow in response to rising water fluxes means that cavity-based drainage systems have significant potential to store water.

Glaciers often rest on unconsolidated sediments, rather than bedrock. The subglacial sediment layer may vary substantially in thickness and continuity, and the sediment characteristics can range from coarse gravel to fine-grained silt and even clay. Porous and permeable subglacial sediment functions as a confined aquifer that can drain of some of the water that penetrates to or is produced at the glacier bed. Measured hydraulic conductivities of *in situ* subglacial sediments are in the range 10^{-4} to 10^{-9} m s^{-1} (Fountain and Walder, 1998). With these hydraulic conductivities, most calculations suggest that Darcian flow through aquifers of thickness $\sim$0.1 m under a hydraulic gradient of $\sim$0.1 is unable to transport discharges of the magnitudes typical of the summer melt season, or even the winter period (Boulton and Jones, 1979; Alley, 1989; Fountain and Walder, 1998). Thus, the aquifer may become saturated and a drainage system will develop at its upper surface. Elution of fines from the upper layers of the sediment (Hubbard *et al.*, 1995) may form a *macroporous horizon* at the surface of the sediment layer (Stone and Clarke, 1993; Fischer and Clarke, 1994). Alternatively, a network of interconnected *microcavities* may form on the downstream sides of larger particles protruding up through the sediment surface (Kamb, 1991).

Such a system could have an effective hydraulic conductivity several orders of magnitude higher than the values quoted above (Fountain and Walder, 1998), and would be capable of adjusting its porosity by dilation in response to water flux variations (Clarke, 1987a).

Alternatively, a system of *canals* may coexist with the subglacial aquifer (Walder and Fowler, 1994). It may be fed directly by englacial conduits, in which case much of the surface-derived runoff would drain through the canal system without entering the aquifer and the aquifer would play a limited role in overall drainage. The canals would have an ice roof and a sediment floor, and would grow by roof melting or basal erosion, and close by deformation of ice or sediment. Canal systems could adopt two very different configurations: an arborescent network incised upwards into the ice, or a nonarborescent network incised into the sediment. The type of system that develops depends upon the magnitude of the effective pressure within the canal system ($P_e = P_{ice} - P_{water}$) relative to some critical value (P_{crit}) which depends upon the hydraulic gradient and the relative creep properties of ice and sediment. Nonarborescent networks are expected when the hydraulic gradient is low (as under ice sheets and ice streams), while arborescent networks are more likely beneath valley glaciers where hydraulic gradients are high. A canal network may exist beneath Ice Stream B, Antarctica (Engelhardt and Kamb, 1997).

Channelized/Fast Subglacial Drainage Systems

Large ice-walled conduits (*Röthlisberger or "R" channels)* are the major component of channelized or fast subglacial drainage systems (Figure 2). They form where the rate of wall melting due to viscous dissipation of energy in flowing water outstrips the rate of creep closure of the overlying ice. The original analysis of the hydraulics of R-channels assumes steady water flow through conduits of semicircular cross section (Röthlisberger, 1972). This analysis predicts an inverse relationship between the water flux through an R-channel and the pressure gradient that drives the water flow. This implies that when two parallel conduits transporting different water fluxes coexist, the water pressure at a given distance from the glacier terminus along the conduits will be greater in the conduit with the lower water flux. Water from the smaller conduit will therefore tend to be captured by the larger conduit, a process that can account for the arborescent character of R-channel networks.

In reality, flow through R-channels fed by surface melt is unlikely to be steady as discharge varies with surface melt rates on timescales that are shorter than those on which channel cross section can adjust by a combination of wall melting and creep closure. Thus, R-channels may cease to be water-filled at low flows, and water pressures may rise to values in excess of the local ice overburden pressure during peak discharges (Hubbard *et al.*, 1995). Although

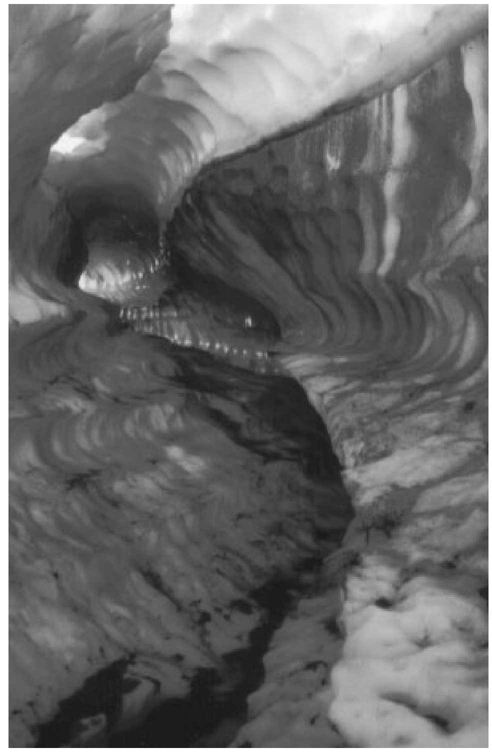

Figure 2 Röthlisberger or "R" channel at Matanuska Glacier, Alaska. Note scalloping and sediment deposits on the channel walls. A color version of this image is available at http://www.mrw.interscience.wiley.com/ehs

there have been attempts to calculate the conditions under which open channel flow might occur within subglacial conduits (Hooke, 1984), there are few data with which to validate the calculations. In general, however, open channel flow is most likely where ice is thin and steeply sloping, and at times when discharge has dropped substantially from a preceding period of high flow.

There is also evidence that the cross-sectional shape of R-channels may not be semicircular. Attempts to forecast the distribution of water pressure within subglacial conduits using Röthlisberger's (1972) analysis have often predicted water pressures substantially below measured values. This might be explained if channels are broad and low in cross section, since such channels would tend to close more readily than semicircular channels, resulting in higher average water pressures (Hooke *et al.*, 1990). Conduits with this geometry might be expected because ice melt would be concentrated low down on channel walls at times of low discharge. In addition, creep of ice into partially full channels would be most rapid at the tunnel ceiling (especially when channels are only partly full), rather than at the base of the tunnel walls where creep is resisted by friction with the glacier bed. These arguments are supported by modeling of the evolution of channel cross section in response to time-varying water fluxes (Cutler, 1998).

Subglacial Groundwater Systems

In some cases, a complex hydrostratigraphy exists beneath a glacier, consisting of interbedded layers of contrasting grain size and hydraulic conductivity. Some of these layers may function as aquifers and others as aquitards. Water flow will tend to be predominantly horizontal within the aquifers and vertical across the aquitards (Flowers and Clarke, 2002a). In such cases, groundwater flow may represent an important sink or source for subglacial water.

A common case is that of a low conductivity till aquitard overlying a higher conductivity sand or gravel aquifer. For such a system, the pressure and elevation gradients across the aquitard drive the exchange between the glacier bed and the aquifer. Upward flow of groundwater commonly occurs where the glacier bed profile is concave (Tulaczyk *et al.*, 2000) and at the glacier margin (Flowers and Clarke, 2002a). The magnitude of the water flux into the aquifer is very sensitive to the aquifer conductivity, and it decreases as the aquitard and aquifer conductivities decrease. The mean water pressure and water fluxes at the glacier bed are therefore strongly influenced by the hydraulic conductivities of both the aquitard and the aquifer, decreasing as the conductivities rise. Thus, these properties of the subglacial sediments act as valves that control the routing of subglacial water. For most reasonable values of till conductivity, fluctuations in water pressure and flux recorded at the ice-bed interface will be transmitted to the aquifer, but with a discernible lag (Flowers and Clarke, 2002a).

Subglacial Lakes

Subglacial lakes can represent a significant form of subglacial water storage. 77 subglacial lakes have been identified beneath the Antarctic Ice Sheet (Siegert *et al.*, 1996). These are mostly located beneath regions of low surface slope close to ice divides, with major clusters in the Dome C and Ridge B regions. Those located away from ice divides are found close to regions where enhanced ice flow begins. The largest lake so far identified is the 230-km long Lake Vostok in East Antarctica (Kapitsa *et al.*, 1996), which has an estimated depth of at least 510 m, an area of $\sim$14 000 km^2, and a volume of 1500 to 7000 km^3. The mean observed length of the 77 lakes identified is, however, only 10.8 km (Siegert, 2000), and the total water volume stored in Antarctic subglacial lakes is likely in the range 4000–12,000 km^3 (Dowdeswell and Siegert, 1999). It is not known whether these lakes formed *in situ* beneath the ice sheet as a result of the progressive accumulation of water produced by basal melting, or whether they represent preglacial water bodies that were overridden by the growing ice sheet in the past. Little is known about whether and how these lakes drain under present-day conditions.

Subglacial lakes are common beneath ice caps in the volcanically active region of Iceland (Björnsson, 2002). High rates of heat flow associated with subglacial hydrothermal systems result in rapid melting of basal ice and a local depression in the ice cap surface. This can create closed depressions in the subglacial equipotential surface in which water can accumulate. Some lakes, such as Grimsvötn beneath Vatnajökull, fill caldera depressions in the bedrock, but the lake water level may rise above the caldera rim so long as the potential gradient continues to direct water flow towards the lake from surrounding areas. Other lakes form more rapidly as a result of extreme rates of basal melting during subglacial volcanic eruptions. These Icelandic subglacial lakes drain by episodic outburst floods (*see* **Chapter 168, Hydrology of Glacierized Basins, Volume 4**).

MAJOR EVENTS IN SUBGLACIAL DRAINAGE SYSTEMS

Spring Transitions

Where surface melting represents a significant input to subglacial drainage systems, there tends to be a strong seasonality to the magnitude of the water input. Subglacial drainage systems adjust to these variations in water flux and may thus have very different configurations in summer and winter. Thus, the subglacial drainage system will undergo major periods of evolution in spring and fall.

The driving force behind drainage evolution in spring is the progressive increase in the amount of surface meltwater delivered to the glacier bed. The temporal pattern of water inputs depends upon the end of winter snowpack distribution, meteorological conditions at the glacier surface, and the ease with which water can penetrate to the bed. Snow, with its high albedo, melts relatively slowly in comparison to glacier ice, and it acts as a potential storage site for water. As snow is removed from the glacier surface, a flood can result from the release of water stored in the snowpack (Flowers and Clarke, 2002b). Where drainage pathways from surface to bed are not immediately available, water is stored in supraglacial lakes, channels, and crevasses at the start of the melt season, and may be delivered to the bed very suddenly (Flowers and Clarke, 2000; Boon and Sharp, 2003). As glacier ice is exposed and subjected to surface melting, both the total daily runoff and the amplitude of the diurnal variation in runoff increase (Willis *et al.*, 2002). These changes in water input drive the evolution from a slow/distributed subglacial drainage system that is characteristic of winter, to a system that is usually dominated by fast/channelized components (Nienow *et al.*, 1998).

Drainage development in spring seems to occur in two phases. Because the drainage system is initially underdeveloped, the first water to penetrate to the bed forces increased separation between ice and substrate not exclusively, but primarily, downstream from points where it reaches the bed (Mair *et al.*, 2002). This generates an increase in the subglacial drainage system volume that acts as a preferred pathway for subglacial drainage, and may allow significant water storage at the bed (Nienow *et al.*, 1998). As ice is exposed and surface meltwater production increases, water fluxes to the bed increase, water backs up in the drainage system, water pressures tend to rise, and major subglacial channels develop rapidly within regions of preferred drainage (Nienow *et al.*, 1998; Gordon *et al.*, 1998; Cutler, 1998). Rainstorms during this period may be a major stimulus to channel growth (Cutler, 1998; Gordon *et al.*, 1998), which is accompanied by the onset of strong diurnal water-pressure cycling in areas close to channels (Gordon *et al.*, 1998). Locally, channel growth is probably directed down-glacier from each point of water input. At a larger scale, however, the subglacial channel network extends upglacier over time at a rate that mirrors the retreat of the transient snowline. Where crevasses and moulins are distributed widely across a glacier, this can be a relatively smooth process. Where they are sparsely distributed, however, the process can be more episodic, with large areas of glacier bed developing channelized drainage in relatively short periods of time.

Where a frozen glacier margin effectively dams the outflow of basal water, a wave of high subglacial water pressures may spread upglacier from the margin in response to surface water input. Eventually, these high pressures are dissipated by either increased water flux into the subglacial groundwater system (Flowers and Clarke, 2002b), or by a sudden release of water at the glacier terminus (Skidmore and Sharp, 1999; Flowers and Clarke, 2000). In this case, channel development may follow the outburst.

End of Summer Transitions

The end of summer is marked by a decline in available melt energy. As water inputs decline, channels cease to be water-filled and begin to contract by ice creep (Cutler, 1998). Initially, water pressures tend to fall, allowing drainage of water previously stored in other components of the subglacial drainage system (Flowers and Clarke, 2002b). Eventually, the drainage system (both fast and slow components) contracts to the point that water pressures begin to rise again (Hubbard and Nienow, 1997), and the distributed/slow components come to dominate the system. In some areas, there is evidence that both the upstream and downstream ends of major channel systems may freeze shut very rapidly when surface melt ceases (Skidmore and Sharp, 1999; Copland *et al.*, 2003). This can preclude further input of surface water and trap water at the glacier bed. In other areas, including Svalbard, drainage continues throughout the winter (Hodgkins, 1997).

Hydrological Changes During Glacier Surges

Glacier surges involve periodic, abrupt one to three order of magnitude increases in glacier flow velocity that are sustained for months to years and separated by periods of quiescence lasting years to decades (Meier and Post, 1969; Raymond, 1987). Ice builds up in a reservoir area during quiescence and then is transferred rapidly down-glacier during surges. Rapid flow during surges is usually attributed to an increase in the basal velocity of the glacier. This results from a reduction in ice-bed coupling due to the buildup of high-pressure water at the bed (Clarke, 1987b). Two models have been proposed to explain how changes in ice flow dynamics might be associated with a reorganization of subglacial drainage (Clarke *et al.*, 1984; Kamb, 1987; Raymond, 1987), though available observations are currently insufficient to validate either model. In both models, transitions in drainage system character between the quiescent and surge phases are linked to the evolution of glacier geometry over a surge cycle. The buildup of ice in the reservoir area during quiescence results in local thickening of the glacier, an increase in the mean surface slope, and a rise in shear stress at the glacier bed (Raymond and Harrison, 1988).

One model of surging assumes the glacier rests on a largely impermeable substrate, and suggests that rapid surge phase motion occurs primarily by basal sliding. This model proposes that during surges, the basal drainage system is distributed in character and may consist of linked cavities (Kamb, 1987). During quiescence, this distributed system is replaced by a low-pressure system dominated by major subglacial channels. This temporal switching of drainage morphologies is analogous to seasonal changes but operates on longer timescales. This model of surging proposes that thickening of the reservoir area during quiescence increases the rate of closure of subglacial channels by ice creep. This drives up the channel water pressure (especially over winter when water fluxes are low) until it exceeds the mean water pressure in the surrounding drainage system. Once this happens, the channels lose water to the cavity system, initiating a positive feedback whereby decreasing water flux results in a further increase in water pressure and further water loss from the channels (Raymond, 1987). This process may be facilitated if the glacier contains sufficient englacial water storage capacity that downward drainage of englacially stored water over winter eventually overwhelms the constricted subglacial system and initiates rapid sliding (Lingle and Fatland, 2003). If the cavity system is sufficiently stable to withstand the following melt season's discharge perturbation, rapid sliding may be sustained (Kamb, 1987). If the surge is propagating down-glacier into thinner ice, water inputs into the drainage system under the surging region increase, and episodic releases of water occur (Kamb *et al.*, 1985). Initially, rapid ice creep shuts down channels that start to form during these releases, but eventually the ice becomes too thin to allow this. Permanent drainage of water stored beneath the surging region occurs and the channel system becomes reestablished. Subglacial water pressure is lowered and the surge phase of motion comes to an abrupt end (Kamb *et al.*, 1985).

The second model assumes the glacier rests on an unconsolidated, permeable bed, and that rapid motion is due in part to deformation of water-saturated sediments (Clarke *et al.*, 1984). Sediment deformation may be negligible during quiescence if the sediment is drained by canals at the till surface, or a network of pipes within the sediments. Such types of system promote more efficient drainage, a reduction of sediment pore water pressure, and an increase in sediment strength. In this model, the rise in basal shear stress due to the evolution of glacier geometry during quiescence may initiate or increase the rate of sediment deformation. Although this might induce dilation of the sediment, increasing its porosity and hydraulic conductivity, it could also destroy pipes and surface canals, reducing the bulk transmissivity of the drainage system, driving up pore water pressures and weakening sediments (Clarke *et al.*, 1984). As the sediments weaken, they will deform more rapidly, initiating the surge phase of motion. Eventually, the down-glacier transfer of ice by the surge results in thinning of ice and a reduction in surface slope in the reservoir region, and a decrease in the basal shear stress acting on subglacial sediments. The basal shear stress may become too low to deform the sediments, and the cessation of deformation may allow drainage pipes and canals to become reestablished. This will further reduce pore water pressures and increase sediment strength, bringing an end to the surge motion.

Glacial Outbursts

Sudden releases of stored water are a common feature of the discharge regimes of glacier-fed rivers. The most common water sources for these "glacier outbursts" are ice-dammed lakes along glacier margins, but releases from supraglacial, subglacial, and englacial sources are also known. Often, these outbursts are routed through the subglacial drainage system and can be a major stimulus for changes in its character (Nye, 1976; Stone and Clarke, 1996; Flowers and Clarke, 2000; Björnsson, 2002). For the case of drainage of an ice-marginal lake, drainage commonly begins when the lake water level approaches that required to float the ice dam. Once water starts to leak into the subglacial drainage system, channel development likely occurs in a manner similar to that described for the spring transition. In some cases, channel development is initiated close to the ice dam and channels grow in a downstream direction over a period of days to weeks. In other cases, initial drainage may take the form of a turbulent sheet flood and channel development first occurs near the glacier margin. This facilitates drainage in this region, and provides a positive feedback for channel growth by reducing back-pressure in the system (Flowers *et al.*, 2004). When there is insufficient water in the lake to keep the channel full, the channel begins to shrink and eventually closes. This terminates the flood, which has a characteristically asymmetric hydrograph, with a gradual rise to peak discharge and abrupt termination. The flood characteristics depend upon the volume of water in the lake, the lake hypsometry, lake water temperature, ice overburden pressure, hydraulic gradient, and wall roughness of the tunnel (Clarke, 1982).

Outbursts from supraglacial and englacial water pockets may have a very different form, with a more abrupt and symmetric hydrograph (sudden break outbursts, Haeberli, 1983). This may reflect the sudden delivery of water to the subglacial drainage system as a result of water pressure-induced propagation of fractures from the void that contains the water pocket (such as a crevasse) to the glacier bed (Van der Veen, 1998). Abrupt increases in water delivery to the bed result in over-pressuring of the subglacial drainage system and hydraulic uplift of the surrounding ice. This in turn permits the disturbance to propagate away from its point of origin (Engelhardt and Kamb, 1997). This may activate new areas of the bed, increase the connectivity of the drainage system and allow flushing of stored water. Such events tend to propagate down-glacier and will eventually culminate as an outburst at the glacier margin. Connections created by these events may persist after the outburst, resulting in a more efficient drainage system (Flowers and Clarke, 2000). In some polythermal Arctic glaciers, such outbursts are the normal means by which subglacial outflow begins in spring (Skidmore and Sharp, 1999).

Tidal Forcing of Subglacial Drainage

Variability in subglacial water pressure and electrical conductivity on diurnal timescales is typically attributed to forcing by inputs of meltwater from the glacier surface. Such variability has, however, also been recorded in winter and in environments that are too cold for surface melting to occur. Semidiurnal fluctuations have also been observed in these environments. This may result from forcing of the subglacial drainage system by earth, atmospheric, or ocean tides. Diurnal variations in basal water pressure of Haut Glacier d'Arolla, Switzerland, in winter resulted from water flow between a borehole and the subglacial drainage system that was forced by deformation of the glacier substrate and the glacier ice induced by the luni-solar diurnal earth tide (Kulessa *et al.*, 2003). Diurnal velocity variations of Ice Stream D, Antarctica, appear to be driven by the ocean tide beneath the Ross Ice Shelf, into which the ice stream flows (Anandakrishnan *et al.*, 2003).

SUMMARY

Subglacial drainage can occur wherever ice at a glacier bed reaches the pressure melting point. The subglacial drainage system is fed from a mixture of surface, englacial, subglacial, and groundwater sources that differ in terms of their spatial distribution and characteristic patterns of temporal variability. Subglacial drainage systems are not readily accessible, and knowledge of their characteristics is derived from a range of indirect methods including radio-echo sounding, the use of artificial tracers, monitoring, and manipulation of subglacial conditions via boreholes, and monitoring of glacial runoff properties. Subglacial water flow is driven by gradients in hydraulic potential, and occurs through either fast/channelized or slow/distributed systems located at the ice-bed interface, or through subglacial aquifers. Water can be stored subglacially in cavities, in the pore space of subglacial sediments, or in subglacial lakes. Drainage system structure evolves continually on various timescales in response to changing water inputs, evolving glacier geometry, and changes in glacier flow dynamics. Major hydrological events in such systems include the spring and fall transitions, outburst floods, and structural changes related to glacier advances and glacier surging. In the absence of variable water inputs from the glacier surface, subglacial drainage systems may be sensitive to forcing by earth, ocean, and atmospheric tides.

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