

MECHANISMS OF ENGLACIAL CONDUIT FORMATION AND THEIR IMPLICATIONS
FOR SUBGLACIAL RECHARGE

By

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To our Creator – The Flying Spaghetti Monster. May he bless us all with his noodled appendage.

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Ideas about the character and evolution of englacial drainage systems have been deeply influenced by the theoretical model developed by Shreve (1972). The Shreve model is based on three main assumptions: (1) englacial drainage is in steady state; (2) englacial water will flow along the steepest hydraulic gradient within the glacier; and (3) pressure head equals the pressure of the surrounding ice minus a small component due to melting of the walls. The Shreve model has been widely adopted as a fundamental component of englacial drainage theory. There is no evidence, however, that the model provides a realistic picture of actual glacial drainage systems. To evaluate Shreve's theory, we used speleological techniques to directly survey englacial conduits. We mapped a total of 8.25 km of passage in 27 distinct englacial conduits in temperate, polythermal, cold-based and debris-covered glaciers between 2005 and 2008. New information reported here is supplemented by published data on 40 other englacial conduits located worldwide and surveyed to ice depths of 176 m using speleological techniques. In all cases, englacial drainage systems consisted of a single unbranching conduit. Englacial conduit morphologies were found to be intimately linked to the orientation of a glacier's principal stresses or the presence of pre-existing lines of high

hydraulic conductivity. If a sufficient supply of water is available, hydrofracturing forms vertical conduits in zones of longitudinal extension and subhorizontal conduits where longitudinal stresses are compressive. On unfractured glacier surfaces, relatively shallow subhorizontal conduits with migrating nickpoints form by cut-and-closure provided channel incision is significantly faster than surface lowering. Conduits can also form along permeable debris-filled crevasse traces that connect supraglacial lake basins of different potential. Only conduits formed by extensional hydrofracture were found to be connected to glacier beds. Our results suggest that a Shreve-type englacial drainage system does not exist and implies that englacial conduits can only penetrate through thick ice to recharge the bed of the Greenland Ice Sheet where supraglacial water bodies either intersect, or are advected through, zones of acceleration.

CHAPTER 1 INTRODUCTION

Interest in englacial drainage systems has increased recently following the realization that surface-to-bed drainage can affect the dynamics of the Greenland ice sheet (Zwally et al., 2002; Catania et al., 2008; Das et al., 2008; Joughin et al., 2008). For many decades, it has been known that surface-derived meltwater can access the beds of temperate alpine glaciers and trigger transient velocity increases (e.g. Iken and Bindshadler, 1986; Iken and Truffer, 1997; Mair et al., 2003; Anderson et al., 2004), but it had been thought that thick, cold ice characteristic of the upper layers of polythermal glaciers (including the Greenland Ice Sheet) would provide an impenetrable barrier to englacial drainage. This conclusion arises from classical conceptualizations of englacial hydrology, which assume that conduits form from convergence of water flow in an integrated network of veins within permeable temperate ice (Shreve, 1972)

In contrast, recent evidence indicates that efficient englacial drainage systems traverse great thicknesses of impermeable cold ice in Greenland and high-latitude glaciers (e.g. Boon and Sharp, 2003; Copland et al., 2003; Catania et al., 2008; Das et al., 2008). As a result, a reassessment of englacial conduit formation is timely. Several conduit-forming mechanisms have been proposed, including the exploitation of fractures (Stenborg, 1969, 1973; Fountain et al., 2005) and permeable debris-filled structures (Gulley and Benn, 2007), hydrologically-assisted fracture penetration (Rothlisberger and Lang, 1987; Boon and Sharp, 2003; Alley et al., 2005; van der Veen, 2007), and incision and closure of supraglacial streams (Fountain and Walder, 1998; Gulley et al., in press). To date, however, no general theory of englacial conduit formation has been proposed.

In this paper, we propose a new general model of englacial drainage evolution, based on our direct observations of englacial conduits in a wide variety of glacier systems, together with published conduit descriptions from the literature.

Techniques and Study Sites

We surveyed conduits during ten expeditions between 2005 and 2008, using standard speleological techniques modified for glacier caves (Gulley and Benn, 2007; Palmer, 2008). The study sites span a wide range of thermal and structural regimes, including cold- and warm-based polythermal glaciers in Svalbard, Norway, the clean, temperate Matanuska Glacier, Alaska, USA, and polythermal debris-covered glaciers in the Khumbu Himal, Nepal. Surveys of englacial drainage networks were conducted at the end of the ablation season, when meltwater flow had largely ceased but passages remained fully open. Survey data were reduced using the COMPASS software program, from which planimetrically accurate maps were drawn. Patterns of strain in the ice were inferred at each site using a variety of methods, including measuring the change in distance between pairs of markers installed in englacial passages, feature tracking on ASTER imagery and crevasse pattern analysis (see Gulley et al., in press and Benn et al., in press for details).

Overall, we have explored more than 45 distinct conduit systems and surveyed a total of 8.25 km of passage in 27 conduits (Table 1-1). We have explored another ~3 km of passages that were not formally surveyed. Repeat surveys of several conduits allowed aspects of their evolution to be directly monitored. Our survey data are supplemented by published data on 40 other englacial conduits located worldwide and investigated using speleological techniques to ice depths of 176 m. The deepest known

descent of englacial conduits was to 203 m below ice surface by French speleologists in Greenland (Badino, 2007).

Shreve Model of Conduit Development

The theoretical model developed by Shreve (1972) has deeply influenced glaciological ideas about englacial drainage (e.g. Röthlisberger and Lang, 1987; Paterson, 1994; Fountain and Walder, 1998; Hooke, 2005). The model is based on three main assumptions: (1) the englacial drainage system is in steady state (i.e. discharge rates and system geometry are constant) and recharge is uniformly distributed; (2) englacial water will flow along the steepest hydraulic gradient through cylindrical conduits developed within permeable, isotropic glacier ice; and (3) water pressure in conduits is approximately equal the pressure of the surrounding ice. The assumption that water pressure equals ice overburden pressure stems from the idea that conduits will expand or contract in response to differences between pressure in the conduit and the surrounding ice, and that the conduit will therefore adopt a geometry that equalizes these pressures. Hydraulic potential ϕ at any elevation z within the glacier is defined as:

$$\phi = \phi_o + \rho_i g(H - z) + g(\rho_w - \rho_i)z \quad (1)$$

where ϕ_o is the potential at an arbitrary datum, ρ_w and ρ_i are the respective densities of water and ice, g is the acceleration due to gravity, H is the elevation of the glacier surface. The second term on the right hand side is the water pressure (assumed equal to ice pressure), and the third term is elevation potential. Differentiating (1) with respect to the horizontal direction x (taken parallel to the maximum slope of the glacier

surface) and evaluating for $\frac{\partial \phi}{\partial x} = 0$ yields an expression for the form of equipotential surfaces within the glacier:

$$\frac{\partial z}{\partial x} = \left(\frac{\rho_i}{\rho_w - \rho_i} \right) \frac{\partial H}{\partial x} \quad (2)$$

Taking $\rho_w = 999.84 \text{ kg m}^{-3}$ and $\rho_i = 916 \text{ kg m}^{-3}$, Equation 2 states that equipotential surfaces should dip upglacier at nearly 11 times the glacier surface gradient. Water is assumed to flow down the maximum potential gradient, so conduits will plunge toward the glacier bed normal to the equipotential surfaces. Importantly, Shreve (1972) states that his model only defines water pressure in conduits but that it is permissible and convenient to assume the equipotential surfaces are defined throughout the ice.

Water flow through conduits melts the walls by viscous heat dissipation. This tends to lower the pressure in the conduit. For steady state, wall melting is compensated by conduit contraction by ice creep. Because large conduits have greater discharge and dissipate more heat per unit wall area, they should also experience a greater pressure drop than small conduits. In consequence, larger conduits will form areas of relatively low potential, and will tap smaller passages in which the hydraulic potential is higher. Shreve's model, therefore, predicts that water-filled englacial conduits form an upward branching, tree-like (arborescent) system, in which large numbers of small tributary conduits link up with one or a few master conduits (Figure 1-1).

It is important to note that the equipotential surfaces defined by Shreve are not the same equipotential surfaces familiar to those acquainted with flow in granular aquifers. The condition that water pressure equals ice over burden pressure stems from the

simplifying assumption that conduits expand and contract in response to changes in conduit water pressure and that these pressures will therefore tend to equalize. This applies a “hidden” work function to Shreve’s calculation of equipotential surfaces (strictly speaking, they are actually equal ice pressure surfaces) and define flow paths that would not occur in a freely draining porous medium (one on which no work is performed). Flow paths in porous media can be derived from the Bernouli equation, which restates the Law of Conservation as:

$$\frac{\rho_w}{\gamma} + \frac{V^2}{2g} + z = \text{constant along streamline } C \quad (3)$$

states that the sum of the pressure head $\frac{\rho_w}{\gamma}$, where γ is the specific weight, the velocity head $\frac{V^2}{2g}$, and the elevation head z , remain constant along a streamline. This constant (C) is the total head of the system.

Because the velocity head component is very small in common aquifer porous media, it is usually neglected and the equation is restated as:

$$P + \gamma z = C \quad (4)$$

Where P is the pressure head and γz are the elevation head components and C is a constant (*total head*). The term *total head* was arrived at by dividing the first and second terms of equation (4) by the specific weight (γ) and has units of meters. Contours of equivalent head in an aquifer are equipotential surfaces and water flow is normal to these contours. Equipotential surfaces for a freely-draining, glacier-shaped porous medium are shown in Figure 1-1 and are rotated nearly 90 degrees to Shreve’s equipotential surfaces. This occurs because Shreve’s model sets the conduit closure

rates equal to their enlargement rates and this process of conduit closure performs work on the system. This is why Shreve states that his model only defines water pressure in conduits,

The Shreve model has been widely adopted as a fundamental component of englacial drainage theory (e.g. Röthlisberger and Lang, 1987; Fountain and Walder, 1998; Hooke, 2005), and forms the foundation of many glacier hydrological models (Palli et al., 2003; Rippin et al., 2003; Pattyn et al., 2005; Price et al., 2002; Fricker et al., 2007; Boulton et al., 1995; Bindschadler and Choi, 2007). Furthermore, the assumption that water pressure and ice pressure are equal has been used as the basis to explain a host of glacier-hydrological processes, including glaciohydraulic supercooling (Lawson et al., 1998; Alley et al., 1997, 1998, 2003), the formation of tunnel valleys (Shreve, 1985), and the formation and stability of subglacial lakes (Nye, 1976; Pattyn, 2008).

It is doubtful, however, whether Shreve's model provides a realistic picture of englacial drainage systems, for several reasons. First, the assumption of steady state conditions is clearly not applicable to systems that are fed by surface melt, which undergo large fluctuations on diurnal, seasonal and annual timescales (e.g. Iken and Bindschadler, 1986; Fountain, 1993; Schuler et al., 2004; Bartholomaus et al., 2007). Second, the model requires that recharge is distributed evenly across the surface of the glacier. If recharge is concentrated at a few discrete points, such as large moulins, equipotential surfaces will have a very different distribution. Third, the assumptions that water pressure equals ice pressure and that gradients in ice pressure determine hydraulic gradients have been controversial (Liboutry, 1996), partly because they have

no analogue in other deformable media (Fowler, 1984). Fourth, the Shreve model describes flow in idealized conduit systems without consideration of how flow becomes sufficiently concentrated in a homogenous and isotropic aquifer to form a conduit in the first place. In this respect, the Shreve model is similar to early models of conduit (cave) formation in limestone where an initiation conduit normal to theoretical equipotentials was required and mechanisms for initial conduit inception were never considered (Rhoades and Sinacori, 1941; Thrailkill, 1968).

The initiation of a Shreve-type englacial drainage system depends on water flow through an interconnected network of water-filled veins formed at ice crystal boundaries (Nye, 1989; Mader, 1992). Nye and Frank (1973) calculated theoretical water fluxes through idealized vein networks, and derived velocities of the order of $0.009 - 90 \text{ cm yr}^{-1}$, depending on water content. Lliboutry (1971), however, pointed out that the energy dissipation implied by such water fluxes would cause runaway ice melt, enough to melt the entire glacier in only a few years. He therefore concluded that some other factor impedes the connectivity of ice veins, reducing the bulk permeability of temperate ice. Capillary forces associated with air bubbles could have this effect, opposing the forces driving water flow. In addition, Lliboutry argued that melting and refreezing at grain boundaries during ice deformation will exert a strong influence on the distribution of water within temperate ice, isolating water-filled voids and locally reducing film thickness to levels too low to support flow. Further factors limiting the permeability of temperate ice were considered by Lliboutry (1996). Experimental work shows that the bulk hydraulic conductivity of glacier ice is extremely low ($5.48 \cdot 10^{-11} \text{ m s}^{-1}$, Jordan and Stark, 2001) or about the same as unfractured recrystallized limestone or marine clays (Ford

and Williams, 2007). Lliboutry (1983) argued that significant englacial water flow can occur near glacier beds, where ice has high water content and pressure gradients are large. The high strain rates in such settings, however, will mitigate against the formation of stable englacial conduit networks, resulting in flow occurring in a constantly reorganizing system of veins and other transient voids. Thus we conclude that the potential for conduit development from vein systems is minimal, even where water content and hydraulic gradients are large.

In addition to the above considerations, there is no observational evidence to support the idea that englacial conduits can evolve by the enlargement and integration of vein networks. Although millimetre-scale tubes have occasionally been observed in ice cores (Raymond and Harrison, 1975) and by video-imaging of borehole walls, the majority of observed macroscopic voids have planar, fracture-like geometries or are associated with veins of clear ice interpreted as healed fractures (Pohjola, 1994; Harper and Humphrey, 1995; Fountain et al., 2005). Furthermore, englacial drainage systems investigated by speleological techniques and ground-penetrating radar have characteristics very different from those predicted by the Shreve (1972) model. Englacial drainages show no discernable tendency to follow theoretical potential gradients, but commonly form steeply plunging shafts or meandering, low-gradient passages interspersed with steep steps (e.g. Holmlund, 1988; Pulina and Rehak, 1991; Badino, 2002; Stuart et al., 2003; Mavlyudov, 2006; Gulley and Benn, 2007; Catania et al., 2008; Gulley et al., in press; Benn et al., in press). Finally, in the course of our exploration and survey of englacial passages, no instances of conduits wholly entombed in intact, unfractured glacier ice were found. Despite the fact that conduits at

atmospheric pressure from regions of very low hydraulic potential, no small infeaser conduits were observed to drain into the passage walls. Importantly, neither dendritic networks nor infeaser conduits were discovered in warm ice at depth within polythermal glaciers or in temperate glaciers. The weight of evidence, therefore, leads us to conclude that intact glacier ice is insufficiently permeable for conduits to form by the mechanisms envisaged by Shreve (1972).

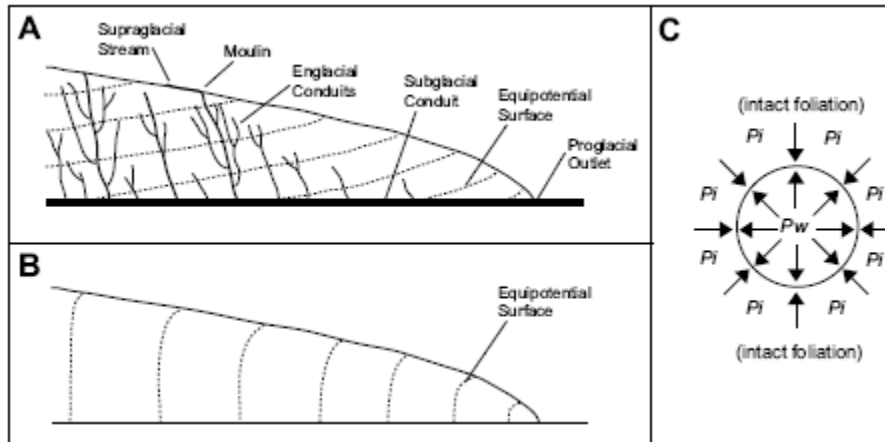


Figure 1-1. Model adapted from Shreve (1985). A) according to the Shreve model, englacial drainage systems should consist of upward branching conduits formed normal to equipotential surfaces. It is unclear where the recharge for conduits terminating in the ice is coming from, B) equipotential surfaces in a glacier if the assumption that ice pressure equals water pressure is ignored. By making the simplifying assumption that water and ice pressure are in equilibrium, the Shreve model applies work on drainage pathways in B by 'squeezing' them into the equipotential contours in A and C) if the Shreve model is correct, conduit cross-sections should be circular and be framed by intact glacier ice as indicated by continuous foliation.

CHAPTER 2

STRUCTURAL CONTROL OF ENGLACIAL DRAINAGE SYSTEMS IN HIMALAYAN DEBRIS-COVERED GLACIERS

The surfaces of stagnant or slow-moving debris-covered glaciers have frequently been compared to limestone karst, due to the presence of features such as swallow-holes, sinkholes, caves and springs (e.g. Clayton, 1964; Kruger, 1994; Kirkbride, 1995; Benn and Evans, 1998). The extent to which the karst analogy extends to the subsurface, however, has rarely been investigated (e.g. Pulina, 1984; Pulina and Rehak, 1991). In limestone karst, speleogenesis occurs where sufficient subsurface water flow exists to remove dissolved rock and keep undersaturated water in contact with the soluble walls. This is possible only where a preexisting, interconnected network of openings, such as bedding planes or joints, links recharge and discharge zones (Palmer, 1991). The location and morphology of limestone caves, therefore, is a function of both secondary permeability and hydraulic gradient. While quantitative theories of englacial drainage evolution have traditionally assumed that ice permeability is uniform and isotropic (e.g. Shreve, 1972), recent work has emphasized the role of active crevasses in guiding water flow. Fountain and Walder (1998) argued that near-surface drainages along the bottom of crevasses could incise into the ice to depths where cryostatic pressures cause closure of the upper reaches of the canyon, creating trapped, sub-horizontal tubular conduits. Alternatively, Fountain and others (2005) have proposed that high hydrostatic pressures in water-filled crevasses could allow fractures to propagate to significant depths and perhaps to the bed. Although there is good evidence that both of these mechanisms operate within glaciers (Robin, 1974; van der Veen, 1998; Pohjola, 1994; Vatne, 2001; Vatne and Refsnes, 2003; Fountain and others, 2005), their applicability to glacier karst evolution is unknown. Both mechanisms

require significant tensile stresses at the surface to initiate and maintain open crevasses, conditions that are unlikely to be widespread on stagnant or near stagnant debris-covered glaciers. Englacial drainage systems within such glaciers, therefore, must either be inherited from some earlier phase of ice flow, or form by some other mechanism.

Understanding the character of drainage systems within debris-covered glaciers, and the factors that control their development, is important because subsurface processes clearly influence surface hydrology and topographic evolution (Kirkbride, 1995; Benn and others, 2001). Information on subsurface drainage systems in such glaciers, however, is difficult to obtain with commonly-used glacial-hydrological techniques. Dye-tracing (*e.g.* Hooke and others, 1988) and geophysical sensing of subsurface voids (*e.g.* Arcone and Yankielun, 2000; Stuart and others, 2003), for example, are impractical due to the ubiquity of coarse, inhomogeneous debris, combined with complex and irregular terrain and problems of access. In limestone karst, the most comprehensive and reliable data on subsurface features are obtained through direct exploration. Therefore, we have adapted speleological techniques to gain access to englacial conduit systems in debris-covered glaciers, and to make detailed maps of their planform and morphology. The results allow the controls on conduit development and evolution to be identified, and provide new insights into the relationship between englacial speleogenesis and glacier surface evolution. We show that the karst analogy is more than skin-deep, and that the principles of karst hydrology provide a powerful framework for understanding the initiation and evolution of drainage within debris-covered glaciers.

Study Area and Methods

This study presents observations of englacial conduits within four debris-covered valley glaciers in the Khumbu Himal, Nepal (Figure 2-1). Detailed conduit surveys were conducted in the Ngozumpa Glacier in the upper Dudh Kosi catchment, and the Khumbu Glacier, which drains the southern side of Sagarmatha (Mount Everest). Conduit systems were also explored and photographed in the Ama Dablam and Lhotse glaciers in the adjacent Imja catchment. All four glaciers have extensive covers of supraglacial debris, which form almost continuous layers 1-2 m thick over most of their ablation zones, except where interrupted by steep ice faces or holes in the glacier surface. Strongly negative mass balance in recent decades has resulted in significant ice surface lowering (downwasting) and stagnation (Kadota and others, 2000; Hands, 2004). The tendency of the lower parts of debris-covered glaciers to stagnate is a consequence of the insulating effect of surface debris, which tends to thicken downglacier and thus progressively reduces ablation rates (Inoue, 1977; Benn and Lehmkuhl, 2000; Nicholson, 2004). Therefore, during periods of negative mass balance, surface lowering rates in debris-covered areas increase upglacier, so that downwasting results in a reduction of both surface gradient and ice flow speed. Uneven patterns of ablation have produced highly irregular surfaces on all of the debris-covered glaciers in the study area, with mounds and ridges up to 50 m high separating topographic lows, many of which contain supraglacial lakes or evidence of their former presence.

Supraglacial lake formation on Himalayan glaciers is encouraged by low surface gradients and irregular topography, and where glacier termini are impounded by large lateral-terminal moraines. In a study of supraglacial lake distribution in Bhutan, Reynolds (2000) found that lake formation only occurs where glacier surface gradients

are $\leq 2^\circ$, a result that was confirmed by Wiseman (2002) for the Khumbu Himal. In the early stages of downwasting, lakes tend to be ephemeral and drain after a few years while they are still relatively small (Benn and others, 2001; Hands, 2005). In contrast, when downwasting is more advanced, supraglacial lakes can attain very large dimensions and pose significant outburst flood hazards to communities located downstream (Richardson and Reynolds, 2000; Quincey and others, 2005). Whether supraglacial lakes develop into major hazards depends to a large degree on their relationship with the englacial drainage system (Benn and others, 2001). Where supraglacial lakes are perched above base level, as determined by the elevation of the outflow stream over the terminal moraine, they can persist only if the lake floor remains unconnected with englacial drainage pathways. If a connection is made, partial or complete lake drainage will follow, thus limiting lake size and lifespan. Where lakes develop at the level of the terminal moraine, however, the englacial drainage mechanism cannot operate. Lake growth can continue unchecked until the moraine dam is breached, and catastrophic drainage may ensue.

In addition to limiting the lifetime of 'perched' supraglacial lakes, englacial conduits can also play a role in the formation of new lakes. Subsidence resulting from conduit roof collapse creates new depressions on glacier surfaces, where ponds can form. Because melting is inhibited beneath thick debris covers, ablation is strongly focused around lake margins where bare ice is exposed and calving is promoted by thermal undercutting (Sakai and others, 1998; Benn and others, 2001). By preconditioning lake formation and drainage, therefore, englacial hydrological systems exert a strong control on mean ablation rates and glacier response to climate change.

To obtain detailed information on subsurface drainage, and to determine the principal controls on their location and morphology, conduits were entered and mapped using a combination of speleological and mountaineering techniques. In all, nine englacial conduits were investigated at elevations of 4,900 m to 5,300 m, making them the highest surveyed caves in the world. Conduit surveys were conducted in November and December, 2005, when surface meltwater production was small. Surveys were conducted by measuring the distance, azimuth and inclination between successive marked stations using fiberglass tape and a Brunton Sightmaster compass and inclinometer. The largest survey closure error was 0.72 m in the horizontal and 0.16 m in the vertical over a surveyed distance of 554 m. Maximum horizontal errors in the surveyed conduit planforms are therefore 0.12%. Passage cross-section dimensions were measured at every survey station. Scaled sketches of all passages in plan, profile and cross section were rendered *in situ*, including details of passage morphology and the structure of the surrounding ice. Survey data were reduced using the COMPASS[®] program. In addition, samples of debris from structures exposed in conduit walls were collected for grain size analysis (wet sieving) and bulk permeability analysis of normally consolidated material (Darcy apparatus).

Results: Englacial Conduit Morphology

Ngozumpa Glacier

Four conduit systems on the Ngozumpa Glacier were mapped in detail. All were located c. 7 km from the glacier terminus at altitudes of 4,900 - 4,950 m, and represent all of the enterable openings discovered in the central part of the glacier (Figure 2-2). The sites were located upglacier of lake basins 7092 and 7093 (discussed in detail by Benn and others, 2001), both of which contained only small residual ponds in 2005.

Conduit NG01 was located at the southern margin of a closed basin near the eastern flank of the glacier, NG02 and NG03 were at the northern and southern margins, respectively, of a similar closed basin near the glacier centreline, and NG04 was located beneath a ridge near the western flank.

NG01(27° 57' 52"N, 86° 42' 02"E): This conduit formed a singular, meandering passage graded to the level of a frozen supraglacial pond, from where entry was made (Figure 2-3). The entrance area consisted of a wide alcove behind an ice-cored talus mound at the base of an ice cliff, and was heavily modified by collapse. Extending from the rear of the alcove was a flat-floored semi-elliptical passage 0.5 m high and 12.5 m wide, the floor of which consisted of stagnant or near-stagnant water overlain by a skin of ice c. 2 cm thick (Figure 2-4a). At a distance of 120 m from the entrance, the ice skin became too thin to support the weight of the survey team, and further progress was made by wading (Figure 2-4b). This demonstrated that water depth was typically 30 - 60 cm and that the channel floor was sub-horizontal with a cover of sandy bedforms. Passage ceiling height gradually increased from 0.5 m near the entrance, to 1 m at a distance of 80 m and 4 m at 130 m, where passage morphology changed from a low, wide semi-elliptical cross-section to a more complex form with an elliptical upper section separated from a lower A-shaped section by a narrow neck (A6, A7: Figure 3). At deeper penetration, much of the passage had a canyon-like morphology with a series of notches and shelves at the lateral margins (Figure 2-4b). At the top of the canyon, the ceiling narrowed to a narrow slot, terminating in a band of coarse, sandy diamict, interpreted as a debris-filled crevasse trace. This cropped out almost continuously along the inner passage, except for two sections where the ceiling ice exposed intermittent

debris and had a complex, irregular form with numerous blind holes and re-entrants (A9, A12, Figure 2-3). Sub-horizontal debris bands cropped out around A5, and parts of the conduit margins nearer the entrance, although passage dimensions were too restricted to allow systematic exploration. The conduit was followed for a total surveyed distance of 364.3 m until it terminated in a ramp of bouldery debris. Light was visible at the top of the boulder slope, although the window was too small to permit exit.

Interpretation: The low relief of the floor profile of the passage is indicative of grading to local base level, determined by the elevation of the supraglacial pond at the passage terminus. Conduit flow was in an up-glacier direction. The wide, low semi-elliptical form of the outer passage is at least partly structurally controlled, as indicated by the presence of sub-horizontal debris bands, but it may also reflect preferential melting of the passage margins under non-full flow conditions (*cf.* Hooke and others, 1990; Hooke, 2005). Variations in the height of the passage ceiling between A6 and A9 (Figure 2-3) suggest that this part of the conduit had an up-and-down long profile, indicative of formation under phreatic conditions. The canyon-like morphology upflow of A13 suggest vadose channel incision guided by the debris-filled crevasse trace. Passage morphology at A7 indicates that the transition from vadose to phreatic flow migrated downstream as incision proceeded.

NG02 (27° 57' 58"N, 86° 41' 50"E): This conduit system consisted of a highly sinuous, multi-level passage that was accessed at its lower end, at the northern margin of a closed basin near the centreline of the glacier (Figures 2-2 & 2-5). Several small ponds occupied the floor of the basin, and patches of laminated sand and silt indicated that lake level had formerly been higher. The conduit entrance had an inverted J-

shaped cross section with upper and lower levels connected by a narrow, sub-vertically oriented canyon. The lower level (Level A, Figures 2-4c & 2-5a) averaged c. 4 m in width and had a distinctive rectilinear cross section, with unpaired sub-horizontal sills and overhanging lips at the lateral margins. The floor, which consisted of a partially-frozen stream flowing toward the entrance, rose upglacier with a gradient of c. 1:110 and gradually converged with successive lateral sills. The canyon walls were inclined alternately to the right and left, rising up towards the centre of each bend. In several localities, a frozen breccia of large (> 1 m) ice blocks formed a false ceiling near the top of the canyon (e.g. A19, A20, Figure 2-5a). In places, blocks were buckled and shortened in the cross-passage direction, indicating narrowing of the passage following emplacement of the ice breccia.

The upper part of the system consisted of a single passage (Level B) near the entrance, plus additional higher levels at more than 25 m penetration (Figure 2-5b, c). Level B had a sub-horizontal floor, into which the canyon linking to Level A had been incised. In several places, the floor of Level B was littered with angular ice blocks, and similar blocks formed an ice breccia choking parts of the canyon (e.g. B6, B10, Figure 5b). The walls of the Level B passage were highly variable in form, commonly with one or more ledges between the main floor and the uppermost level (Level C). At the end nearest the entrance, Level C formed a shelf c. 5 m above the Level B, but then formed an independent passage with a low, broad and irregular cross profile. Debris bands cropped out in the walls of Level C throughout its length, either at the lateral margins of the passage or in the roof. Mostly, the debris bands formed planar structures which cut across, and in places displaced, the primary foliation, and are interpreted as crevasse

traces (Figure 2-4d). In one part of the ceiling, an approximately tubular debris-filled structure was exposed, consisting of cross-bedded sands with a sparse ice matrix. This is interpreted as a relict, infilled englacial conduit.

At a distance of 150 m from the entrance, Level B narrowed to a partially-collapsed constriction, then opened out into a large, flat-floored passage c. 16 m wide and 8 m high, informally named the Soccer Field (Figures 2-4e & 2-5b). Levels A and C also connected with the Soccer Field: Level A as an incised channel running first parallel to and then across the passage axis, and Level C as a laterally elongated hole near the passage roof at its eastern end. The roof of the Soccer Field was highly irregular and traversed by numerous cross-cutting planar debris bands, one of which could be traced for c. 20 m along the centre of the roof, parallel to the passage axis. The ice on either side of the debris bands displayed discordant foliation, and the whole mass is interpreted as formerly heavily crevassed ice which had been partially reconstituted. Passage enlargement at this locality was achieved partly by the collapse of ice blocks from the roof, as evidenced by numerous large, angular fragments littering the floor. The frequency of fallen blocks was greatest at the eastern end of the Soccer Field, where they allowed access to a short, poorly developed vadose in-feeder passage (Figure 2-5c). This extended in a northwesterly direction for c. 20 m before terminating in a boulder choke. A single, steeply dipping debris-filled crevasse trace was observed in the passage walls and ceiling at this locality. Light was observed at two points along the crevasse trace, but the dimensions of the portals were too small to permit exit.

Two constricted passages led out of the northwestern end of the Soccer Field. One extended from the innermost end of Level A, and was almost entirely choked by

fallen ice blocks, whereas the other was graded to Level B and consisted of a 1.5 m high and 10 m wide slot between a floor of fallen blocks and a roof composed of semi-detached flakes (A31, Figure 2-5b). Both passages opened dramatically into a large, ice-block floored room 49.1 m long by 15.2 m wide, informally named the Reptile Room (Figures 2-4f & 2-5b). The roof was highly irregular, consisting of numerous partially isolated ice blocks bounded by cross-cutting linear debris bands, indicating that, like the Soccer Field, the room had been enlarged by collapse of structurally weakened ice forming the roof. Small, partially collapsed passages extended from the rear of the Reptile Room, but were too small to permit entry.

Three sediment samples were collected from the crevasse trace exposed in the ceiling of Level C (sample locations are shown in Figure 2-3). Samples NG02-a, b and c are poorly- to moderately sorted, clast-rich sandy diamicts, and have bulk permeability values of 1.2×10^{-4} , 7.0×10^{-5} and $4.5 \times 10^{-4} \text{ m s}^{-1}$, respectively.

Interpretation: The uppermost levels of NG02 display very strong associations with debris-filled crevasse traces, indicating that initial conduit location was structurally controlled (Figure 2-4d). For the most part, passage morphology reflects predominantly fluvial processes, but the two largest sections, the Soccer Field and Reptile Room, formed by a combination of fluvial erosion and gravitational collapse of the passage roofs. Collapse at these localities appears to reflect numerous structural weaknesses in the overlying ice. For the most part, ice blocks do not appear to have been fluvially transported following collapse. The ice breccias in Level A, however, appear to have been floated in at the surface of water filling the passage below, then were frozen into place before the waters receded. Buckling of ice blocks in some of the breccias is

interpreted as evidence of closure of the passage in response to stresses in the surrounding ice.

Accumulations of refrozen meltwater at numerous locations in and below Level C indicate that during the previous ablation season water entered the system via multiple surface feeders, although many of these features probably post-date passage formation. Steps in the lateral margins of the passage walls between Levels C and B are discontinuous and could mark varying rates of lateral channel migration during incision. In contrast, the flat floor of Level B is continuous throughout its length, compatible with channel grading to a stable base level, most likely a supraglacial lake occupying the closed basin at the downstream end of the passage. The dramatic change in morphology from flat-floored passage to sub-vertical canyon below Level B is interpreted as the result of a sudden drop of base level associated with partial drainage of the supraglacial lake. Initial base level drop was c. 8 m, followed by intermittent minor falls recorded by the sills at the margins of the Level A passage. This interpretation is compatible with evidence from satellite images of the glacier. An Aster scene from November 2002 shows a lake occupying the basin south of NG02 (A. Luckman, personal communication), which had almost entirely drained by February 2005 (Figure 2-2). We conclude, therefore that channel incision below Level B probably occurred following lake drainage in either 2003 or 2004.

The glacier surface above NG02 was mapped to determine whether subsurface features had any surface expression. Surface crevasse traces were found at several points above Level C, and around the entrance of the vadose feeder at the eastern end

of the Soccer Field. Neither the Soccer Field nor the Reptile Room was associated with any subsidence at the surface.

NG03 (27° 57' 55"N, 86° 41' 51"E): This conduit system consisted of a highly sinuous, multi-level passage beginning 97 m SSE of NG02, at the southern margin of the same closed basin (Figures 2-2 and 2-6). The entrance area was in a north-facing ice cliff, and took the form of a large alcove following the line of a prominent debris-filled crevasse trace with a dip angle of 45-50° (Figures 2-6a and 2-7a). Two major passages extended in a southwesterly direction from the back of the alcove, separated by a band of ice which was traversed by innumerable anastomosing tubes and blind holes, with a morphology akin to Swiss cheese. Most tubes were a few cm in diameter, but one was sufficiently large to permit entry (Figure 2-7b) and was found to penetrate through to the lowermost passage (Level A). For several tens of metres, the ice ceiling of this lower passage was perforated by anastomosing tubes and holes (A4-A5, Figure 2-6a).

The initial 50 m of Level A consisted of a wide tunnel with a low ceiling, with a series of unpaired steps and notches at the lateral margins (A3, A3b, Figure 2-6a). At a penetration of 50 m, ceiling height increased to c. 5 m, and passage morphology changed to a canyon with inclined walls, the slope direction of which varied systematically with passage bends, in a similar, but more dramatic, way to that observed in NG02. In cross section, the canyon walls commonly formed a series of steps, consisting of wider sections with gently sloping floors and ceilings separated by narrower steeper sections. The upper extremity of the canyon tapered to very small dimensions, above which it opened out into a larger upper passage (Level B). At some localities, the canyon had a false ceiling formed of a frozen breccia of sub-rounded ice

blocks (e.g. A11, A13, Figure 2-6a). After a total surveyed distance of 253.3 m, Level A terminated in a choke of boulders similar to the glacier surface debris (A20a, Figure 2-6a). A small window to the outside was observed at the top of the boulder choke, but was too small to permit exit.

The entrance to Level B was the upper passage leading from the entrance alcove, mentioned above. Near the entrance, the passage passed behind some ice pillars, then terminated at a steep drop down to Level A (shown in Figure 6A for clarity). On the other side of a gap, Level B continued as a sinuous passage, with a morphology that contrasted dramatically with that of Level A (Figure 2-6b). Passage cross sections were generally cylindrical in form, and in some places consisted of a single circular passage with a narrow canyon incised into the floor linking down to Level A below (B12, Figure 2-6b; Figure 2-7c). More usually, the morphology was more complex, with two or more passages stacked one above the other, linked by narrower slots (e.g. B4, B6, B9, Figure 2-6b). A debris-filled crevasse trace was exposed along the entire length of the ceiling of Level B, mirroring the planform of the passage to a remarkable degree (Figure 2-7d). Bends in the Level B passage were commonly substantially offset from those in Level A below. As a result, the floor of Level B was undermined in several places, and some large rooms recorded localized collapse. Large numbers of sub-rounded ice blocks occur on the floor of Level B, locally forming false floors of loosely ice-cemented breccia and littering the entrance alcove. There was a clear upper limit to the blocks, and they did not occur on the ledges bounding the uppermost conduit levels.

Higher-level passages were observed at one locality. The first was a narrow infeeder that entered from the NW, near the E end of the gap in Level B (C2, C3, Figure

2-6b). Near the lower end of this infeasurer, an elliptical conduit branched off from then rejoined the roof of Level B (Figure 2-6a; C1, Figure 2-6b). This conduit, which appeared to be genetically unrelated to the infeasurer, was 0.6 to 1.4 m in diameter, and by-passed a bend in Level B before trending parallel to it following the same crevasse trace.

Interpretation: The elliptical form of the upper passages in Level B indicates that the earliest phase of flow recorded in NG03 was under phreatic conditions. The relationship between this part of the conduit and the crevasse trace clearly demonstrates that the conduit did not form by incision from the surface. Had the conduit originated as a channel downcutting through the overlying ice, and was subsequently isolated from the surface by creep closure, evidence of this history should be preserved in the ice above and around the conduit. However, extensive outcrops of debris-filled crevasse traces in the ceiling of the conduit and around the entrance alcove show that the conduit did not cut down through these structures, but formed at its present position relative to the surrounding ice, leaving the adjacent structures intact. We conclude that the conduit was initiated *in situ* when a debris-filled crevasse trace provided an efficient pathway between areas with a large difference in hydraulic potential. For most of its length, the conduit is singular, although the short parallel offshoot (Level C) may record main conduit bypass. Once established as a large and efficient conduit, the Level B passage would have determined the maximum level of the supraglacial lake at its upstream end.

The 'keyhole' morphology of the upper parts of NG03, consisting of a circular upper passage with an incised slot in the floor, is identical in form to limestone cave

cross-sections recording phreatic to vadose flow transitions. In this case, incision of the floor was probably associated with propagation of head loss upstream from the passage exit. Although the initial conduit planform was controlled by the trend of the parent structure, during incision it appears to have become independent of structural control, and shifted location due to lateral migration of meanders.

There is some evidence for catastrophic drainage of the former supraglacial lake at the upstream end of NG03. First, the large numbers of large, sub-rounded ice blocks occurring up to the floor of Level B are compatible with substantial discharges through the conduit system. The ice blocks probably originated as icebergs in the lake, but some may represent mechanical erosion of the conduit walls. In either case, large numbers of blocks were floated in at the surface of the transporting waters and then froze together to form perched ice breccias. Second, the anastomosing tube networks between Levels A and B are morphologically similar in many respects to those generated by forcible injection of flood waters into bedding planes in limestone caves (White, 1988), and it is possible that they record the breakthrough of water from below lake level into low-pressure tunnels.

NG04 (27° 57' 53"N, 86° 41' 45"E): The entrance of this passage was located near the northern end of a linear trough close to the crestline of a ridge near the west flank of the glacier (Figure 2-2). The passage consisted of a narrow, non-meandering, unbranching slot extending to the southwest, which followed the line of a crevasse trace, the upper part of which was choked with bouldery debris. In cross section, most of the passage was bell-shaped, with a flat floor, but the innermost parts were spindle-shaped with a 1.25 m wide central section and narrow, tapering upper and lower

canyons. The passage “floor” decreased in elevation in the downglacier direction, and at a surveyed distance of 37.2 m and a depth of 19.7 m, became too constricted to permit further navigation.

Interpretation: NG04 is interpreted as an undeformed longitudinal crevasse trace which was exploited and modified by meltwater. The absence of meandering suggests low discharge and the small size (c. 10^2 m^2) of its surface drainage basin is consistent with that interpretation. The entrance zone is open to the atmosphere, and it is possible that this part of the passage evolved from a channel at the base of an open crevasse. In the inner part of the passage, however, the presence of coarse, poorly sorted debris infilling the upper parts of the crevasse trace indicate that the conduit did not form at the surface and melt its way downward, but formed at a deeper level, by-passing the upper parts of the crevasse trace.

Khumbu Glacier

KH01 (27° 58' 52" N, 86° 50' 07" E, 5,200 m): This conduit was located at the southern margin of a large closed depression near the western flank of the glacier, 630 m ENE of Gorak Shep (Figures 2-8 & 2-9). Three conduit entrances were present, one above the other, each of which coincided with the outcrop of a prominent sub-horizontal debris-filled crevasse trace (Figure 2-8a). A cascade of silt-rich waterfall ice descended from the middle entrance to the lower one, indicating drainage of meltwater between passages during the previous melt season. A relict incised surface channel extended a short distance upglacier from the lower entrance, which appears also to have contributed water to KH01. Only the lowermost entrance was accessed and surveyed.

The entrance zone of the lower conduit was 5 - 15 m high and 4 - 5 m wide (A1, Figure 9a). The outermost 17 m of the floor sloped downward at 23° and was partially

composed of fallen ice blocks cemented by refrozen meltwater. The ceiling and walls in the entrance zone displayed large numbers of semi-detached flakes of deformed ice up to 0.5 m thick and 3 m long (Figure 2-8b). The highest part of the canyon tapered into a debris-filled crevasse trace, which could be traced almost continuously along the entire length of the passage. Cross section dimensions diminished with distance from the entrance, and at a penetration of 30 m the passage closed down into a tight squeeze 0.4 m wide and 0.3 m high (A6, Figure 2-9a). For 24 m beyond this point, passage dimensions remained very small, but then increased dramatically in size, forming an 18 m high, 3.5 m wide canyon (Figure 2-8c; A8, A10, Figure 2-9a). In plan view, the canyon exhibited a highly sinuous form, at one point passing within 0.5 m of closing in on itself. In some places, light was visible high above the canyon floor, and extensive covers of refrozen meltwater coated the walls from the unreachable upper portal to the floor. The canyon floor consistently dropped in elevation, with low gradient sections interspersed with a series of frozen waterfalls 0.5 m high. Passage width showed considerable variation, from as wide as 7.0 m to narrow squeezes between bulging walls. Passage height also varied considerably (Figure 2-9b). The deepest part of the passage consisted of a room 27 m long, 4.5 m wide and 6.5 m high, which terminated in a plug of frozen meltwater. The walls and ceiling of the large room were decorated with semi-detached ice flakes similar to those observed in the entrance zone, and ice blocks and coarse rock debris covered the floor. Despite the length of the passage, the innermost end is in close proximity to the entrance.

Interpretation: The almost continuous outcrop of a debris-filled crevasse trace in the roof demonstrates that the highly sinuous form of KH01 reflects structural control. Its

planform suggests that the crevasse trace has been folded during glacier flow, forming a tortuous line of relatively high permeability which was subsequently exploited by meltwater. As is the case for NG03, the relationship between KH01 and the host structure shows that the conduit did not incise downward from the surface, but formed *in situ* by exploiting subsurface sections of the crevasse trace. The large variations in passage width and height (Figure are interpreted as the result of modification by ice creep (locally narrowing the passage and reducing ceiling height), vadose intrusion of surface water (locally widening the passage and increasing ceiling height), and ceiling collapse.

The abundance of semi-detached flakes on the walls of the entrance zone and innermost room is a very striking feature of KH01. Such flakes were not observed in any other passages we examined, so probably reflect local factors. There was no obvious relationship between flake location and any structural features, as was the case for the highly irregular roof morphologies in NG02. As already noted, the entrance zone and innermost room of KH01 are in close proximity, so it is possible that the spalling of flakes from the conduit walls reflects current or past stress conditions in this part of the glacier. Whatever the cause, flake formation and collapse is clearly a locally-important mechanism of passage enlargement.

KH02 (27° 58' 12" N, 86° 49' 45" E, 4,900 m): This relatively short conduit was located in a large closed depression near the centreline of the Khumbu Glacier c. 100 m downglacier of its confluence with the Kangri Glacier (Figure 2-9). With a bell-shaped cross-section nearly 35 m across, KH02 had by far the largest sustained passage cross-sectional area of all caves investigated during this study (Figure 2-8d). The single,

straight passage maintained consistent dimensions and plunged at 45° in a down-glacier direction for a distance of 180 m before terminating in a plug of frozen meltwater. The passage floor consisted entirely of coarse bouldery debris similar to that on the glacier surface. In its innermost part, sub-vertical debris-filled crevasse traces could be seen in the passage ceiling, but closer to the entrance no obvious structural features were apparent. Throughout, the passage walls and ceiling exhibited a striking array of scallop forms.

Interpretation: The presence of crevasse traces in the inner part of the passage suggest that the location of KH02 was determined by the secondary permeability of the glacier, similar to the other mapped conduits. Subsequent to its formation, however, the passage appears to have undergone major enlargement by non-fluvial processes, obscuring the relationship between the passage and any pre-existing structures. There was no evidence of recent ice collapse events that could have enlarged the passage, although such events may have occurred earlier in its evolution. We attribute the scalloping of the walls to the action of warm air currents, so passage enlargement is at least partly the result of turbulent heat transfer from the atmosphere.

Ama Dablam and Lhotse Glaciers

Conduit systems were also entered on the Ama Dablam and Lhotse Glaciers (Figure 1) in mid-November 2005, but were not surveyed because of melting conditions and extreme instability of the conduit roofs. Observations made at these localities, however, usefully supplement the more detailed results described above.

Ama Dablam Glacier: Three major englacial conduits were discovered close to the eastern flank of the Ama Dablam Glacier, only one of which (AD01) was deemed safe enough to enter. This conduit had two entrances in a northwest-facing ice cliff at the

margin of a relict supraglacial lake basin ($27^{\circ} 53' 03''$ N, $86^{\circ} 52' 52''$ E, 5,000 m, Figure 2-10a). The left-hand, upper entrance (AD01A) had an elongated kidney-shaped cross-section with a major axis of c. 15 m inclined at c. 45° . The entrance led into a meandering passage, the ceiling of which was centered on an inclined debris-filled crevasse trace. After approximately 100 m, the passage turned 90° to the right, where further progress was halted by deep water and overhanging walls. The right-hand, lower entrance (AD01B) was c. 12 m high and had an inverted J-shaped cross section, with a sub-horizontal upper level perched above a inclined lower canyon (Figure 2-10b). An inclined debris-filled crevasse trace cropped out in the ceiling of the upper level. The passage floor consisted of a series of thin horizontal ice layers perched above voids, marking successively lower water levels. Water was observed to flow beneath a thin ice skin c. 1.5 m below the lowest of these false floors. The fragile nature of these features prevented exploration of the deeper parts of the passage, but teams in AD01B and AD01A were able to establish communication, demonstrating that both formed part of a single conduit.

Interpretation: The close association between conduit location and crevasse traces provides another example of tortuous englacial drainage controlled by secondary permeability. In addition, the presence of poorly-sorted debris in crevasse traces in the passage ceilings demonstrates that the conduits did not form by closure of a canyon melted down from the surface, but formed *in situ* by the exploitation of permeable weaknesses within the ice.

Lhotse Glacier: Several conduits were observed in close proximity around the margins and floor of a basin near the north-western side of the Lhotse Glacier ($27^{\circ} 54'$

48° N, 86° 54' 09" E, 5300 m). Extensive laminated sand and silt deposits on the basin floor indicate that it was formerly occupied by a supraglacial pond. At the lowest part of the basin floor was a 0.5 m wide hole (LH01) leading to a short, downward sloping passage then a vertical shaft, at the bottom of which flowing water could be heard.. Approximately 20 m to the south-west of LH01, and c. 2 m higher, was a crater-like depression in the basin floor (LH02), with a diameter of c. 10 m. Rills in adjacent silt deposits indicated that water had formerly drained into the bottom of the depression, but boulders choking the floor prevented further investigation.

Three abandoned conduit portals were exposed in the walls of an 8 m high ice cliff at the northern margin of the basin, each of which formerly delivered water to the basin (Figure 2-10c). The left-most portal (LH03) consisted of a narrow, non-meandering canyon c. 1.5 m wide and 8 m high, the upper half of which was infilled with coarse, poorly-sorted debris. Approximately 2 m to the right were two circular portals, each c. 2 m in diameter, one c. 2 m above the other. The lower portal (LH04A) led into a short, gently ascending passage, then a vertical 3 m step led up to a narrow, higher level meandering canyon. Discontinuous sub-horizontal benches were developed on both margins of this canyon, graded to the level of the upper portal (LH04B, Figure 2-10d). A vertical debris-filled crevasse trace occurred in the ice dividing the two portals, and along the entire length of the upper ceiling.

The floor of LH04A continued to rise with distance from the entrance, and passage dimensions progressively diminished. At the farthest accessible point, c. 50 m from the entrance, a series of small holes connected with the glacier surface above and a deep pit appeared in the floor, at the bottom of which running water could be heard.

Interpretation: Structural control on passage location is again clearly evident. The main interest of this locality, however, is that passage morphology clearly shows the effects of intermittent lowering of base level. LH04B emerges at a similar elevation to the highest occurrence of laminated deposits, and appears to have been graded to the maximum level of the former pond. LH04A is graded to a second, lower lake elevation, and we infer that it was formed - and the upper level abandoned - when the pond partially drained through the conical depression (LH02). Significantly, lowering of the water efflux point did not occur by progressive incision, but by switching from LH04B to a new, lower-level outlet (LH04A, Figure 2-10c, d). We conclude that this lower level outlet was created when a potential gradient was developed through a permeable crevasse fill linking the upper-level passage floor and the newly-exposed base of the ice cliff, and the resulting water flow melted a new conduit section. Once established, LH04A evolved by headward incision, but passage adjustment to the lowered lake level was interrupted by abandonment of the conduit. A second conduit formed in the adjacent crevasse trace (LH03), and the lake drained completely (through LH01). In turn, LH03 was abandoned when the feeder water accessed deeper weaknesses, and by-passed the older conduits.

Englacial Conduit Formation

All conduits examined in this study exhibited close association with planar, debris-filled structures. Similar structures are very common in ice exposures on all four glaciers visited in this study, and are interpreted as the infills of former crevasses which have been advected downglacier by ice flow. In the upper part of the ablation zone of debris-covered glaciers, coarse, poorly to moderately well-sorted supraglacial debris commonly falls into open crevasses. When crevasses close as they are advected down-

glacier, the granular fills form planar structures within relatively clean ice, which may be subsequently rotated or structurally deformed. With measured K values in the 10^{-4} to 10^{-5} m s^{-1} range, granular fills in relict crevasse traces provide high hydraulic conductivity pathways through otherwise effectively impermeable glacier ice, which serve as inception horizons for conduit development. Conduits observed in this study cut across foliation and bubbly-ice veins without deviation, and we conclude that these features generally lack sufficient permeability to significantly influence englacial water flow at the macroscopic scale.

Although the downcutting-and-closure model of englacial conduit genesis proposed by Fountain and Walder (1998) may apply locally on Himalayan debris-covered glaciers, it cannot explain the origin of most of the conduits observed in this study. The relationship between conduit location and well-exposed intact crevasse traces (e.g. Figures 2-7a, 2-8a & 2-10c) clearly demonstrates that the conduits formed *in situ* relative to the surrounding ice, and did not cut downward from the surface. We conclude that conduits are initiated when gradients in hydraulic potential across permeable, granular horizons are sufficient to drive water through the medium, and grow by a combination of wall melting and evacuation of the debris fill (Figure 2-11). This situation is analogous to speleogenesis in soluble limestone, where water flow and passage enlargement is possible only where a preexisting, interconnected network of openings links recharge and discharge zones. In debris-filled crevasse traces, pore spaces between grains could provide such an interconnected network, allowing sufficient through-flow of water to initiate a positive feedback between passage enlargement and discharge, leading to conduit formation.

In the case of the network of anastomosing tubes observed in NG03 (Figure 2-7b), the relationship between passage formation and the parent structures appears to be slightly more complex. The anastomosing network lies in the same plane as major debris-filled crevasse traces exposed nearby, but the majority of individual tubes are developed in clean ice. We speculate that the tubes may have evolved from an interconnected system of microfractures that bridged relatively narrow zones of clean ice between debris-filled crevasse traces. That such an anastomosing networks are not widespread suggests that the growth of macro-scale tubes from microscopic fractures requires special circumstances, such as very high hydraulic gradients, and/or influx of relatively warm water. High hydraulic gradients could have existed between the bottom of the supraglacial lake at the upstream end of NG03 and vadose conduits within the surrounding ice. Parts of the walls of NG01 also display complex systems of holes, and it is possible that these areas also acted as relatively low-conductivity zones during initial passage formation.

Passage Evolution

All of the observed conduits exhibited clear evidence of incision along most or all of their length, in the form of canyon-like passage morphologies or slots cut in conduit floors. Such passage morphologies are very common in limestone caves, and represent passage adjustment to local base level under vadose flow conditions, whereby hydraulic potential is determined solely by passage elevation, and downcutting proceeds until hydraulic gradients are eliminated (Palmer, 1991). NG01 provides a very clear example of this process, where the passage morphology records progressive grading to local base level, in this case the supraglacial pond at the efflux point (Figure 2-3). Supraglacial lakes, however, provide inherently unstable base levels, and base-level

lowering consequent on lake drainage will trigger renewed downcutting, as demonstrated by the incised floor of Level B in NG02 (Figure 2-4d). While local topography determines local base level for most of the conduit systems mapped in this study, local drainage basins can connect with deeper englacial drainage pathways if water finds a hydrologically-efficient connection through the intervening ice (Benn and others, 2001). When such a connection is made, the effective base level for the system drops and shifts location, forcing renewed adjustment of passage morphology.

In detail, passage morphology reflects the interaction between structure, base level changes, and hydrodynamic factors. In some cases, steep-sided canyon-type passages clearly reflect incision through a vertically-oriented crevasse trace (*e.g.* NG04, KH01, LH04), but more commonly the lower, incised portions of conduits show increasing independence from parent structures (*e.g.* NG02, NG03). In many of the observed conduits, incision was accompanied by lateral channel migration and the superimposition of short-wavelength meanders on the larger-scale planform. The development of meanders in supraglacial streams, like those in alluvial streams, requires supercritical flow (Knighton, 1972, 1981, 1985), and we infer that these forms reflect peak flow conditions. Meander migration during conduit incision has, in several cases, undermined higher parts of the system, leading to collapses and the development of large rooms (*e.g.* NG03). Collapses were also observed in areas of structurally weak ice, *i.e.* where ice is intersected by numerous debris-filled crevasse traces (NG02). This type of ice apparently originates in heavily crevassed areas, possibly icefalls in the upper ablation zones. A combination of fluvial erosion and gravitational collapses has the potential to open up large voids within the glaciers, which

ultimately may initiate surface subsidence such as that described by Benn and others (2001).

All of the observed conduits were within a few tens of metres of the glacier surface, so creep closure of the tunnel walls is unlikely to be a major determinant on channel morphology. Locally, however, there is clear evidence for creep closure, in the form of deformed ice breccias between conduit walls. If cumulative strain rates exceed wall melting, creep will result in closure of the upper, older parts of englacial canyons, as postulated by Fountain and Walder (1998).

Conduit Formation and Surface Topography

Five of the conduits explored in this study intercepted the glacier surface at both ends (NG01, NG02, NG03, AD01 and LH04). Although some of the conduit entrances may have been exposed by ice surface lowering, others have clearly been created after the glacier surface attained more or less its present form. The formation of conduits between two adjacent basins is readily understood using a karst-type model of speleogenesis. Large gradients in potential can occur between surface basins on debris-covered glaciers if they are at different elevations or contain lakes with different surface levels. Where a debris-filled crevasse trace traverses the intervening ice, ideal conditions are created for the development of drainage pathways. Supraglacial lakes will tend to drive water flow outward, towards regions of lower potential. Seen in this way, perched supraglacial lakes may, in some cases, be the agents of their own destruction and could drain along conduits that they helped to create, rather than by simply intercepting pre-existing conduits as proposed by Benn and others (2001).

On the highly irregular surfaces of Himalayan debris-covered glaciers, topographic hollows define a complex system of potential wells, which act as base level for local

drainage evolution. Where hydraulically-efficient pathways towards such wells follow a structurally deformed crevasse trace or a system of intersecting traces, they may be extremely tortuous and trend across- or even up-glacier for considerable distances, regardless of local ice-surface slope. Thus, gradients in ice pressure exert no discernable influence on englacial water flow paths at the macroscopic scale, due to the extremely low permeability of intact glacier ice and its highly anisotropic structure.

It is clear, though, that conduits do not only develop between adjacent basins. In the case of the conduits explored in the Lhotse Glacier, evidence was found for drainage re-routing beneath a former lake basin, by a process of by-pass. In that case, water was able to exploit a new, deeper pathway following lake drainage and base level fall, presumably because steep hydraulic gradients and hydrologically efficient pathways coincided where they had not done so before. Clear evidence for this process in a subaerial setting was provided by the conduits exposed in the basin margins (LH04: Figure 2-10c, d). Meltwater routing beneath lake basins has also been described by Benn and others (2001). Between Lake 7093 and NG04 near the western margin of the Ngozumpa Glacier (Figure 2-2), a large moulin formerly existed, at the bottom of which rushing water could be heard. This water passed beneath Lakes 7093 and 7092 at a time when both lakes had high levels and were apparently unconnected with the englacial drainage system (both lakes had almost completely drained in 2005). The only possible outlet for this water was a portal near the western end of Spillway Lake near the glacier terminus (Figure 2-2), indicating an englacial transport distance of at least 5.6 km. Although little is known about long-distance englacial water transport within debris-covered glaciers, it is likely that marginal locations such as this will provide

hydraulically efficient pathways because infilled crevasse traces are likely to be particularly abundant near former shear margins (*cf.* Stenborg, 1968, 1973), and efficient water flow will likely occur along the interface between the glacier and the flanking moraine.

The flow paths of shallow englacial conduits on the four debris-covered glaciers investigated are determined by pre-existing lines of high hydraulic conductivity rather than cryostatically determined equipotential gradients. Debris-filled crevasse traces serve as speleogenetic inception horizons. The direction of water flow is governed by the direction of least-resistive hydraulic potential and not in the direction of steepest hydraulic gradient.

Once initiated, conduits evolve by grading to local base level. Grading is accomplished by headward retreat and downcutting, producing a canyon-like passage morphology. Many of the conduits showed evidence of phreatic to vadose transitions, in the form of upper levels with circular or elliptical cross sections and canyon-shaped lower levels, indicating that throughout most of their existence passages were at atmospheric pressures.

Channel migration during incision and collapse of structurally weak ice roofs can create large voids within the glaciers. Such voids may develop into centres of surface subsidence if the overlying ice is thin.

Local base level for many conduits is provided by supraglacial lakes occupying basins on the glacier surface. Lake drainage results in sudden drop in base level, triggering rapid conduit adjustment.

Conduits at different elevations within the ice can have very different base levels, and conduits may pass beneath water bodies to discharge at lower, more distant outlets. The ultimate base level is provided by the elevation of the outlet stream passing over the terminal moraine.

By locally elevating hydraulic potential in basin floors, supraglacial lakes may help to drive water out through debris-filled crevasse traces to areas of lower potential, either adjacent basins or toward low-pressure conduits beneath the basin floor. Supraglacial lakes may, in some cases, facilitate their own destruction, rather than simply intercepting existing conduits during growth.

The mechanisms of englacial conduit formation are analogous in all important respects to speleogenesis in soluble limestones. In both cases, cave inception relies on a continuous hydraulic connection between areas of different potential, which is then exploited and enlarged by water flow. The *in situ* formation and evolution of subsurface drainage networks in stagnant or near-stagnant debris-covered glaciers is closely similar to that in limestone, and relationships between subsurface hydrology and surface topographic evolution are also closely similar.

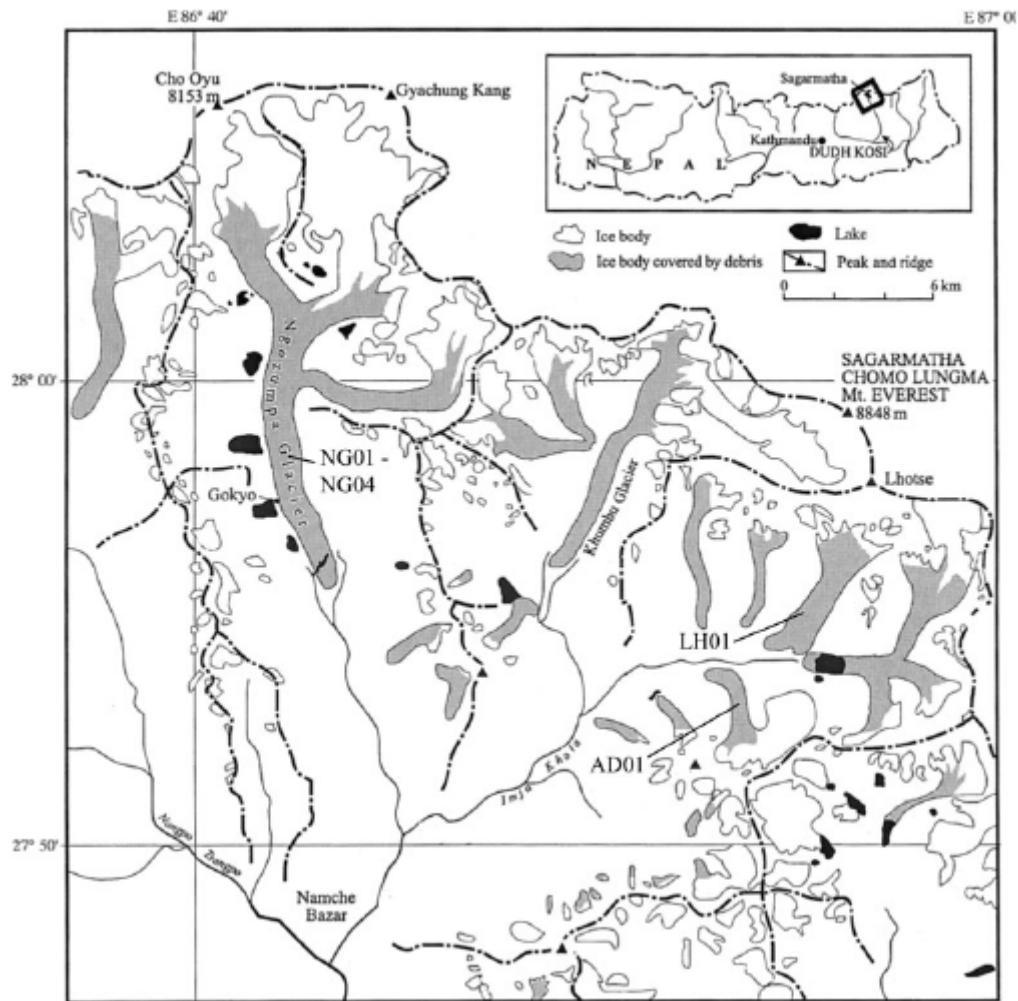


Figure 2-1. Khumbu Himal, Nepal, showing locations of englacial conduits explored in November-December 2005.

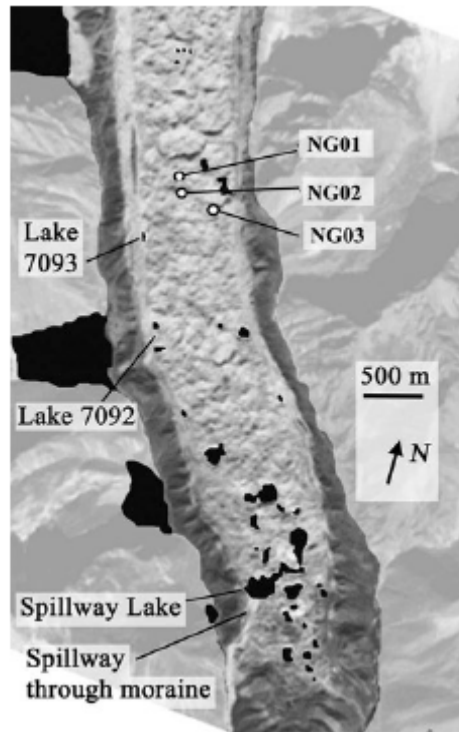


Figure 2-2. The lower part of the Ngozumpa Glacier, showing the position of cave entrances and other locations mentioned in the text. The base layer is an ASTER image from 15 December 2005.

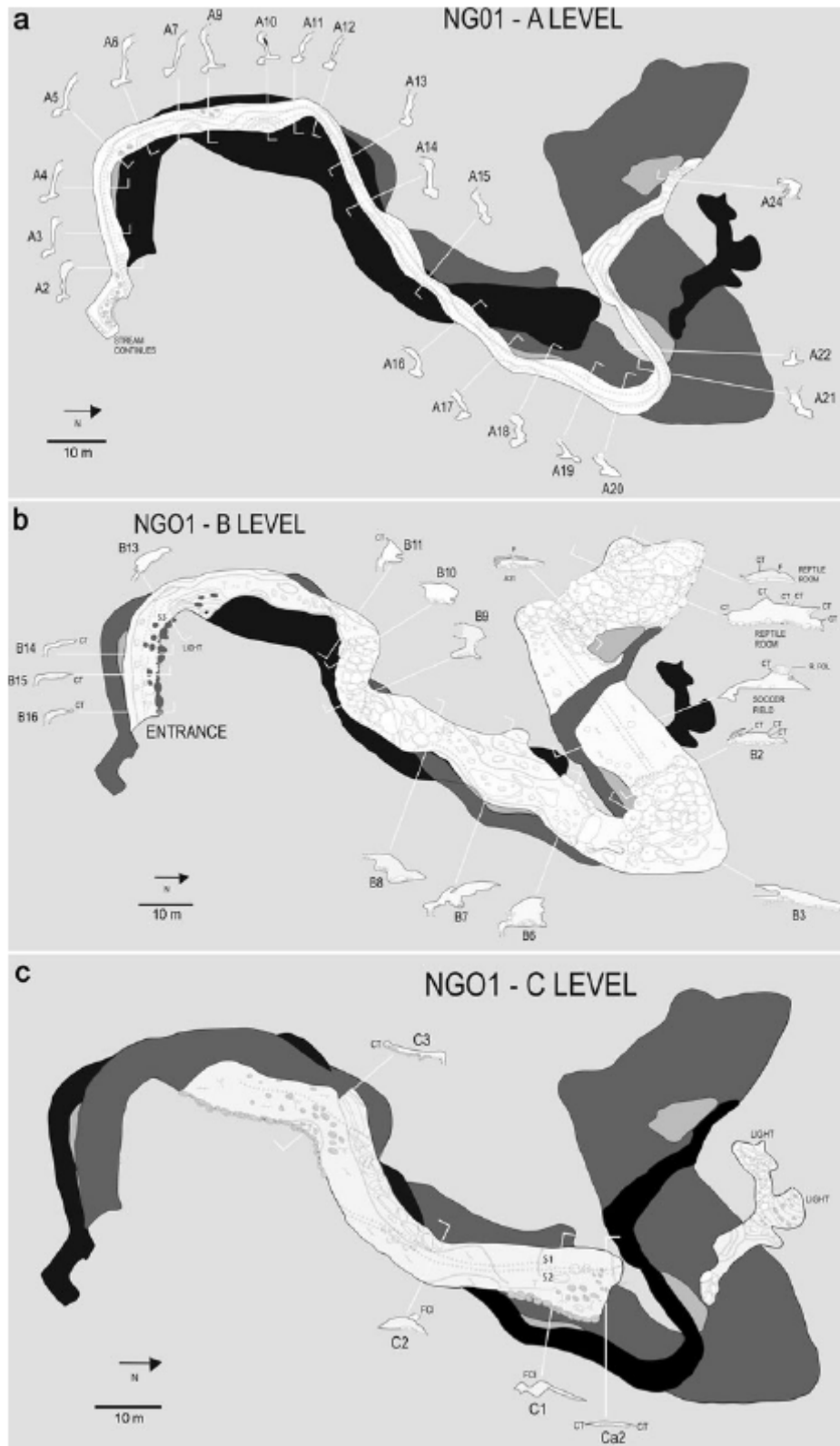


Figure 2-3. Map and passage cross-sections for NGO1. A) lower level, B) lower level and C) upper level. Overall water-flow direction was towards the south.

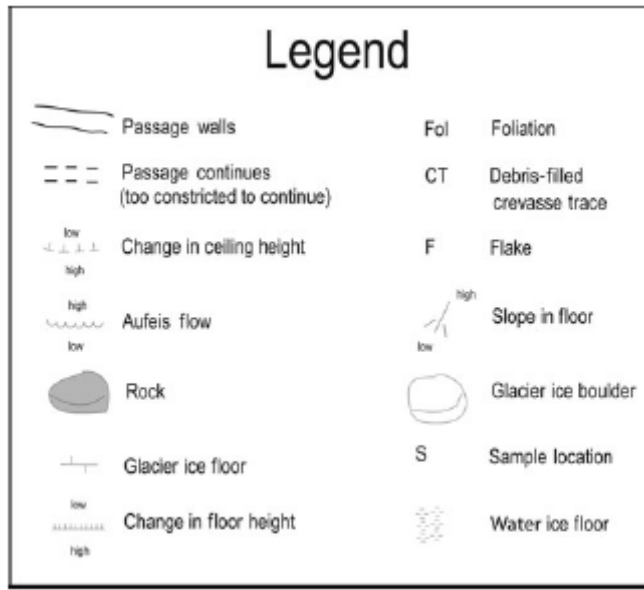


Figure 2-4. Legend for conduit survey maps.

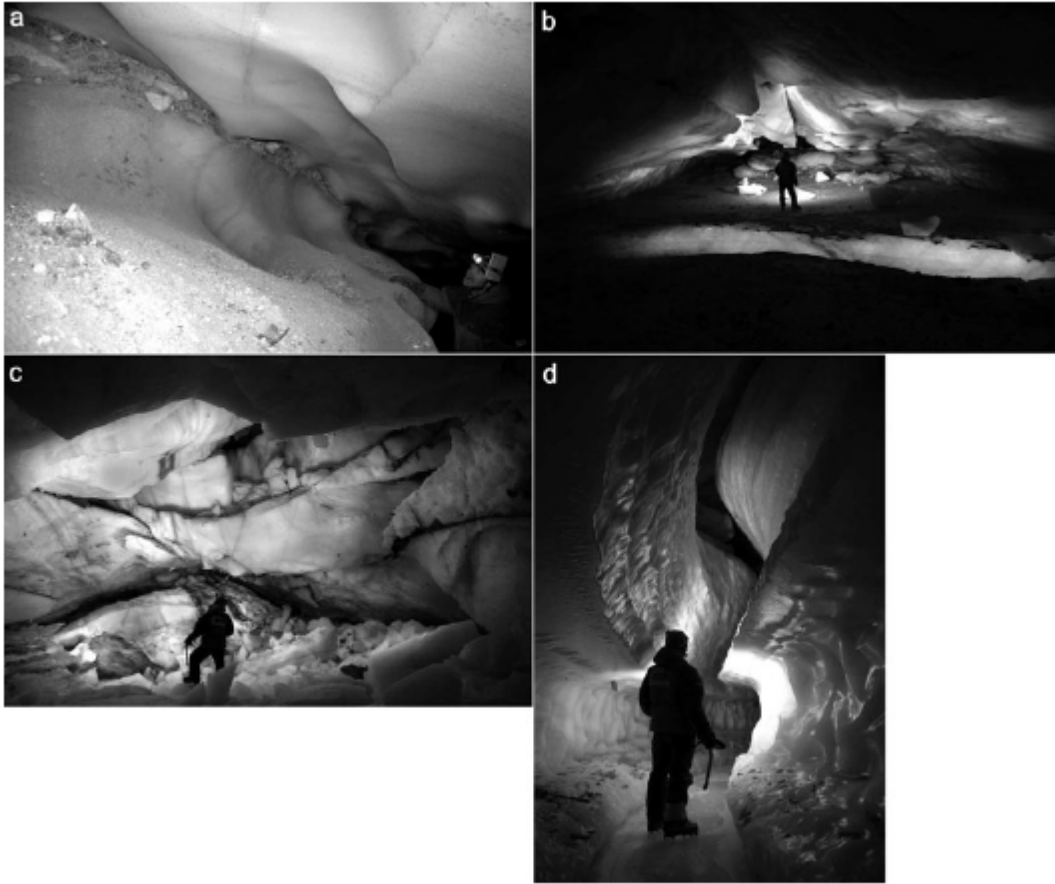


Figure 2-5. Photographs of NGO1, Nogo zumpa Glacier. A) debris band cropping out along the uppermost part of level C, B) the 'Soccer Field', level B. Note the level A canyon incised into the flat floor (running left to right), the fractured ice in the ceiling and the fallen blocks in the farthest part of the picture, C) the 'Reptile Room', level B. Blocks fallen from heavily fractured ceiling litter the floor and D) canyon passage morphology, near station A11.

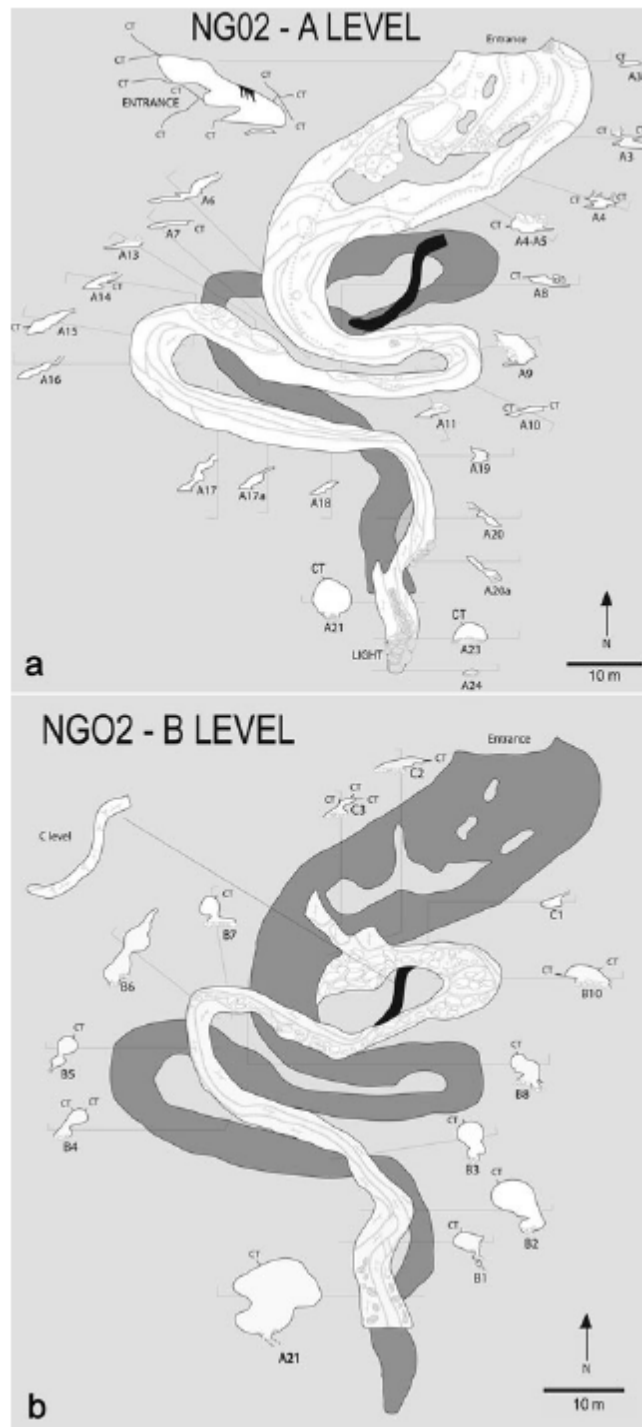


Figure 2-6. Map and passage cross-sections for NGO2. A) lower level and B) upper level. Overall water-flow direction was towards the north.



Figure 2-7. Photographs of NGO2 and NGO3, Ngozumpa Glacier. A) entrance alcove, NGO2, B) Part of the network of anastomosing holes, above station A4, NGO2, C) circular conduit with incised floor, indicating phreatic-vadose transition, near station B12, NGO2, D) crevasse trace in the ceiling of level B (station B9). Note close association between bends in conduit and crevasse trace, NGO2, E) entrance area of NGO3 and F) the inner passage of NGO3, near station A10.

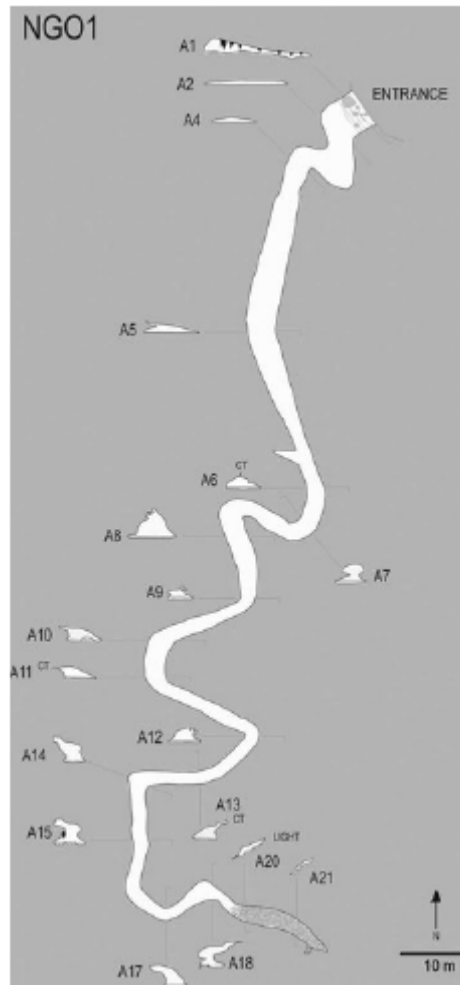


Figure 2-8. Passage plan-form and cross-sections, NGO3. Overall waterflow direction was towards the north.



Figure 2-9. Photographs of conduits on Ama Dablam and Lhotse Glaciers. A) the two entrances of AD01. Both entrances are linked by a sinuous conduit. Ama Dablam (6812m a.s.l) is in the background, B) canyon passage morphology in AD01B, C) portals LH03 (left) and LH04A (lower right) and LH04B (upper right) viewed from outside and D) portals LH04A (lower) and LH04B (upper), viewed from inside.

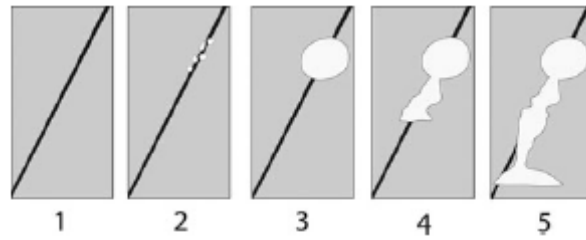


Figure 2-10. Conceptual model of conduit evolution. 1) Debris-filled crevasse trace (aligned perpendicular to the page), 2) water flow creates networks of proto-conduits, 3) elliptical phreatic tube, 4) vadose incision of conduit floor towards local base level and 5) incision become independent of parent structure and passage floor stabilizes at base level.

CHAPTER 3

A CUT AND CLOSURE ORIGIN FOR ENGLACIAL CONDUITS IN UNCREVASSED REGIONS OF POLYTHERMAL GLACIERS

The realization that surface-derived meltwater can modulate basal motion of high latitude glaciers and ice sheets has led to increased interest in mechanisms of englacial drainage through polythermal glaciers (Boon and Sharp, 2003; Zwally, and others, 2002; Copland and others, 2003). How englacial drainage systems form and evolve, and under what circumstances they can reach glacier beds, are important but poorly understood issues. The 'classical' model of englacial drainage development (Shreve, 1972) is based on the idea that arborescent systems of conduits evolve from intergranular veins by exploiting the primary permeability of ice, analogous to Darcian flow in a homogeneous, isotropic medium. Individual conduits are predicted to trend normal to equipotential surfaces determined by elevation, ice overburden pressure and conduit radius. In cold ice, however, the water-filled vein networks invoked to act as conduit nuclei are absent, so this model cannot explain the evolution of drainage networks that traverse cold regions of glaciers (Hodgkins, 1997, Stuart and others, 2003). Moreover, some reported characteristics of englacial drainages in polythermal glaciers are incompatible with the 'classical' model. For example, passages have been observed to dip more steeply (Holmlund, 1988; Pulina and Rehak, 1991; Pulina, 1984) or less steeply (Pulina, 1984; Arcone and Yankielun, 2000; Vatne, 2001, 2003; Stuart and others, 2003; Gulley and Benn, 2007) than predicted, or to have a fracture-like geometry (Fountain and others, 2005).

A number of alternative models have been proposed for the development of englacial drainages. Fountain and Walder (1998) proposed that conduits can form by incision of free-surface streams flowing along the bottom of crevasses. According to this

model, ice creep will pinch off the crevasse above the incising stream and the conduit may, over many-melt seasons, reach the glacier bed. Vatne (2001) described an englacial conduit in Austre Brøggerbreen, Svalbard, which appears to have formed by incision of a stream channel followed by creep closure. The role of surface crevasses in controlling the location of this conduit, however, is not known. It has long been supposed that stress conditions in water-filled crevasses could allow fractures to propagate to the bed, initiating steep surface-to-bed moulins (e.g. Robin, 1974; Rothlisberger and Lang, 1987). Recently, theoretical analyses (Alley and others., 2005, Van der Veen, 2007), observations of surface drainage events (Boon and Sharp, 2003) and direct observations (Benn and others, submitted) indicate that this mechanism can rapidly route surface meltwater through cold ice, thus explaining the close association between surface melt events and accelerated basal motion. Finally, Gulley and Benn (2007) have shown that near-surface conduits can form along permeable englacial debris bands connecting areas with different hydraulic potential.

These multiple theories of englacial formation have been developed primarily through interpretations of proxy data, such as dye-tracing and geophysical measurements, or theoretical considerations, thus limiting understanding of the physical processes controlling their formation and distribution. Studies of conduit morphologies in limestone have greatly facilitated interpretation of proxy data and refinement of numerical models (White, 1988) thus morphological studies of englacial conduits should also facilitate glaciohydrological modeling efforts. However, detailed maps of englacial conduits are virtually non-existent. Consequently, we have created detailed, three dimensional maps of englacial conduits in cold ice in Longyear Glacier (Longyearbreen)

Svalbard, Norway and the Khumbu Glacier, Nepal. These glaciers were chosen as representative of uncrevassed high-Arctic glaciers and high altitude debris-covered glaciers, respectively, and were studied as part of a systematic survey of englacial conduits in a range of glaciological settings.

The conduits were entered and surveyed using standard speleological techniques modified for glacier caves (Dasher, 1994; Gulley and Benn, 2007). Maps and scale drawings of passages were rendered *in situ*, in plan, profile and cross-section view and include observations of glaciostructural and stratigraphic features exposed in passage walls, allowing the origin and evolution of the conduits to be reconstructed in detail. In addition, repeat visits to the caves allowed aspects of their evolution to be directly monitored. At Longyearbreen, rates of passage closure were determined during winter 2006-2007 using pairs of stakes drilled into opposite tunnel walls, and floor incision rates in an adjacent subaerial reach of the drainage system were measured during summer 2006. A detailed account of measurement methods is given by Müller (2007).

Longyearbreen, Svalbard

General Description

Longyearbreen (78° 11' N, 15° 31' E) is ~5 km long, and flows northward from a cirque accumulation basin (Figure 3-1). Radio-echo soundings and direct temperature measurements indicate that the glacier is predominantly below the pressure melting point, except for a temperate surface layer in its upper reaches associated with refreezing of summer meltwater within the snowpack (Etzel Müller and others, 2000; Humlum and others, 2005). It is therefore a polythermal glacier of Type B according to the classification of Blatter and Hutter (1991). The glacier appears to be everywhere

frozen to its bed, and the observed ice flow velocities of 1 to 4 m yr⁻¹ are entirely attributable to ice creep (Etzelmüller and others, 2000). The uncrevassed glacier surface is mostly debris-free, except for inactive ice-cored moraines in the terminal zone and along both flanks of the glacier. Mass balance measurements conducted from 1977-1992 showed that Longyearbreen had an average mean annual specific mass balance of -0.55 m yr⁻¹ at that time (Hagen and others, 2003), and observations on nearby glaciers suggest that the mass balance has probably been even more negative in recent years with net mass loss occurring over the entire glacier surface (Neumann, 2006). During the ablation season, numerous meltwater channels develop on the surface of Longyearbreen, with the largest forming in the hollows between the main body of the glacier and the ice-cored lateral moraines. The lateral drainage systems on both sides of the glacier have supraglacial and englacial reaches.

The focus of this study is the larger and better-developed western lateral drainage system. In late August 2006, the upper part of the system consisted of an incised supraglacial channel which was intermittently roofed by old snow. Below an elevation of 486 m, the snow roof became more continuous, and could be traced as a meandering line across the glacier surface for 262 m until it became buried by debris cover at the foot of the ice-cored moraine (Figure 3-2A). A similar meandering snow roof, with occasional holes, was also visible adjacent to the ice-cored moraine at an elevation of c. 423 m. Meltwater re-emerged from the glacier at a resurgence below the ice-cored moraine at an elevation of 357 m, and thereafter followed an incised supraglacial channel to the glacier terminus. The englacial reach has been explored on several occasions during the past decade. Repeated surveys by Hansen (2001)

between 1998 and 2001 showed that the englacial part of the system consisted of a meandering passage with long gently-sloping sections interspersed with short steep steps. Passage morphology changed from year to year as a result of vertical incision and horizontal meander migration. Vertical incision rates were in the region of $1.2 - 5 \text{ m yr}^{-1}$, higher than net glacier surface ablation rates of $\sim 1 \text{ m yr}^{-1}$ at this location. Between 2001 and 2003, part of the passage floor was incised into the glacier bed, which consisted of angular scree-like regolith $\sim 35 \text{ m}$ below the glacier surface (Figure 2B). *In situ* plant remains recovered from the ice-bed interface show that no basal motion has occurred at this site since the glacier advanced over it c. 1100 years ago (Humlum and others, 2005). Hansen (2001) concluded that the englacial/subglacial part of the western drainage system had evolved from a deeply-incised supraglacial channel, following creep closure of higher, abandoned levels. To investigate the origin of the western drainage system in more detail, we conducted studies of supraglacial and englacial reaches during summer 2006 and winter 2006-07. Studies of the supraglacial parts of the system focused on short-term rates and patterns of incision, which are difficult to measure in englacial parts of the system for safety reasons. Work in the englacial parts of the system focused on longer-term patterns of channel incision and migration, and tunnel closure mechanisms. Rates of tunnel wall closure over the period September 2006 to January 2007 were determined from repeated measurements of pairs of stakes emplaced in the passage walls.

The Lower Supraglacial Channel

The channel consisted of gently-sloping sections interspersed with nickpoints and had an overall gradient of 0.13. Rates of channel incision and lateral migration, and concurrent measurements of ablation of the ice surface on either side of the channel,

were determined at three cross sections in a 12 m channel reach at an elevation of 332 m between JD 209 – 251 (July 28 to September 8) (Table 3-1; Müller, 2007).

Measurements began when the snowline retreated past the site and ended at the end of the ablation season.

We recorded large spatiotemporal variations in incision rates. Well-above-average incision rates recorded at Sites 1 and 2 occurred when a nickpoint migrated through the cross sections. Nickpoint migration rate was 35 to 47 cm day⁻¹ while channel floor incision rates of 4.1 cm day⁻¹ were measured at Site 3. Total incision of the channel floor averaged 4.05 m compared with glacier surface lowering of 1.01 m, so that the channel floor was ~3 m deeper relative to the glacier surface at the end of the period than at the beginning.

Englacial Passages

The englacial portion of the western drainage system was entered at two localities (LYR1, LYR2; Figure 3-1) in October 2006. The upper reach (LYR1) was surveyed downglacier for 332 m. The passage was followed down a series of nickpoints (Figure 2C) for an additional ~600-700 m until further progress was blocked by deep pools of water. The passage was observed to continue, however. The lower reach (LYR2) was surveyed for 570 m upglacier and 59.3 m downglacier. Including meander bypasses, 745 m of passage were surveyed in LYR2. Upglacier progress was terminated by a deep pool of water, probably a few meters short of LYR1. Downstream, LYR2 terminated in a frozen pool with no above-water continuation.

Both reaches were meandering canyons with sinuosities of 2.17 (LYR1) and 2.11 (LYR2) (Figure 3-3A, B: a legend for these and all subsequent maps is provided in Figure 3-4). Unless otherwise noted, all measurements of gradient and sinuosity apply

to the entire surveyed reach of conduits as measured along the thalweg. Meander cutoffs occurred in several places, sometimes perched several meters above the main passage. Average passage gradient was 0.03 in both cases, but most of the elevation change is attributed to nickpoints.

LYR1 has a canyon morphology with sub-parallel walls, which are either vertical or tilted up towards the centre of meander bends where they taper into snow or neve. Fresh snow drifts or icings partially or completely blocked the canyon at several locations (e.g. between A16 and A17, Fig 3A).

Passage morphology in LYR2 is more variable, and can be classified into four main types: (1) *plugged canyons*; (2) *sutured canyons*; (3) *horizontal slots*; and (4) *tubular passages*. The *plugged canyon* morphology (Fig 2D) is similar to that in LYR1. Plugged canyons are defined as relatively narrow, tall passages roofed by snow or aufeis infill. In LYR2 this occurs mainly downglacier of the entrance (F5 – A2, Figure 3C) where the passage reaches a maximum of 11 meters in height at the downstream end of a series of nickpoints.

The term *sutured canyon* is introduced to refer to canyons with walls brought into contact by ice flow. They do not have infillings of snow or aufeis and the passage walls taper into an angular ceiling. Sutured canyons are common upglacier of A29 (Fig 3-3C). Areas of contact (or *sutures*) can occur partway up canyons, where bulges in opposing walls are brought together, or form the passage roof or floor (e.g. Fig 3-3C, stations: A29, A30, A31, A48; Fig 3-5A). In several places sutures separate upper and lower levels of the passage (e.g. between A40 and A50 in Fig 3-3C) and have probably isolated numerous inaccessible voids within the glacier. Creep closure has also folded

ice floors and deformed icicles at several locations. Passage closure rates were determined at A29 (Fig 3-3C,D) (12.8 m ice overburden) by measuring the change in distance between three pairs of stakes emplaced in opposite walls between September 20 2006 and January 18 2007. During this period, the walls converged by 4.2 to 4.8 cm, an average rate of 0.35 to 0.40 mm day⁻¹. Similar measurements made at sections A22 (Figure 3-3C,D) (8.0 m ice overburden) and F3 (9.1 m ice overburden) indicated stable passage dimensions over the same period.

Horizontal slots are wide with very low roofs, and occur in several places upstream of A32 (Figure 3-5B). They are commonly separated from upper canyon passage segments by sutures. Horizontal slots tend to become increasingly lower as winter progresses through a combination of aufeis accumulation and creep closure.

The *tubular passage* sections in LYR2 are roughly elliptical in cross-section, but tend to be somewhat irregular. They occur only in a restricted area, a few tens of metres upstream from the entrance (A4 - A7, Figure 3-3A). Morphologically, this passage type is a hybrid between the circular passage cross-sections that form under pipe-full (phreatic) conditions and the incised canyon morphology of vadose streams (Figure 3-5C; Gulley and Benn, 2007; Ford and Williams, 2007).

Although other workers had previously reported observing the glacier bed in past manifestations of LYR2, we did not reach the bed during 2006-07. In the vicinity of the former subglacial reach, the floor was only 15 m below the surface and about 15 m above the glacier bed. This means that the same conduit was at least 15 m higher in the ice than it was in 2003 (Humlum et al., 2005).

Conduit blockage and discharge rerouting occurred at the lower end of LYR2 during our study. In October 2006, LYR2 terminated downstream in a frozen pool. Soundings taken through the ice revealed the pool was 8 metres deep. This pool was not present in the winter of 2005-06, at which time a nickpoint led to a lower passage which rapidly diminished to a low horizontal slot.

In July 2007, LYR2 backflooded and discharged from the roof of F5 to a supraglacial channel (Figure 3-5D). During this event there was no discharge issuing from the LYR3 resurgence.

Interpretation

The incised canyon morphology, stepped long profile, and snow and neve plugged roofs indicate LYR1 originated as an incised supraglacial channel and became isolated from the surface. Conduit deepening occurs each summer because incision rates outpace surface ablation. Incision is dominated by nickpoint migration. Nickpoint migration is faster than downcutting because viscous dissipation provides the energy for melting and is directly proportional to discharge and channel slope, (Fountain and Walder, 1998). More steeply sloping sections of the channel floor, therefore, will incise more rapidly than gently sloping reaches. Large nickpoints will tend to migrate more rapidly than small ones, leading to 'capture' and the rapid evolution of sloping reaches into series of waterfalls and plunge pools (cf. Knighton, 1985; Piccini et al., 2000).

Passage roof closure results from three processes: plugging by snow and ice, burial by debris derived from the ice-cored moraine, and creep closure at greater depths. Plugging occurs when snow bridges fail to completely melt during the ablation season. In the following autumn the wet snow bridges refreeze, increasing their density and encouraging their survival in subsequent years. Debris flows from the ice-cored

moraine can cover the snow bridges, further encouraging their survival by reducing surface melt rates. Sutured canyons are interpreted as incised passages in which upper levels have closed by ice creep. Measured wall convergence rates show that ice creep can be an important mechanism of passage closure at depths greater than ~12 m. Low, wide passages (horizontal slots) are created in the deepest parts of the system by a combination of lateral channel migration and the closure of relict higher levels by creep and secondary ice accumulation.

The tubular passage section between sutured and plugged canyon sections in LYR1 appears to reflect the re-enlargement of a reach that had been previously reduced by creep closure. Constricted conduit cross-sections encourage transient phreatic (pipe full) conditions during periods of high meltwater discharge, and we infer that consequent passage enlargement creates hybrids of phreatic and vadose cross-sections. (cf. Gulley and Benn, 2007; Ford and Williams, 1989). Transient phreatic conditions are a common hydrological condition in limestone karst and are called epiphreatic, or flood phreatic (White, 1988).

Winter closure of passages by ice build-up and/or ice creep forces conduits to find an alternate outlet – most commonly a breach in the passage ceiling. Blockage and drainage re-routing occurred at the lower end of LYR2 during winter 2006-7. The englacial channel became blocked when the deep horizontal slot observed in the 2005-6 winter season was occluded by waterfall ice originating from a hole in the ceiling. Meltwater filled the lower part of LYR2 up to the level of the lowest outflow point onto the glacier surface (Figure 3-3C - between A1 and F3), 14 m above the level of the frozen pool. From this point, the stream followed the glacier surface, bypassing ~50 m

of former englacial passage. This resurgence (Figure 3-3C, between A1 and F3) created a new local base level for the conduit upstream, which would have flooded back to A19. Flooding therefore created an epiphreatic loop from a formerly vadose passage. Previous epiphreatic conditions would explain the tubular passage morphology between the entrance and A8 (Figure 3-3C). Melting of the conduit ceiling at times of high discharge could initiate upward incision. This process, called paragenesis in limestone caves, would be assisted by bedload armoring of the conduit floor when suspended sediments are deposited due to diminished sediment transport capacity under pipe full conditions (White, 1988; Palmer, 2003). Downward incision is expected to occur at the resurgence at the same time that upward incision was occurring in the flooded conduit. Once downward incision lowered the water level in the conduit back to atmospheric levels, the conduit would incise downward into the ice as before.

Khumbu Glacier, Nepal Himalaya

General Description

Khumbu Glacier (28° 00' N, 86° 51' E) is a 16 km long, debris-covered valley glacier in the upper Dudh Kosi catchment in Nepal, terminating at c. 4,900 m above sea level. The main accumulation basin is in the Western Cwm to the south of Mount Everest, from where ice flows via the Khumbu icefall to the ablation area. Below the icefall, debris cover increases in extent and thickness towards the terminus where it is several metres thick. Measurements in shallow boreholes in the upper part of the ablation area indicate perennial cold ice in a surface layer ~16 m thick and temperate ice at greater depths (Mae, 1976). Furthermore, pronounced seasonal velocity variations in this area (Kodama and Mae, 1976, Seko and others, 1998; Nakawo and others, 1999) strongly suggest the existence of temperate basal ice beneath the icefall

and upper ablation areas. Khumbu Glacier is therefore interpreted as polythermal, of Type D in the Blatter and Hutter (1991) classification.

The ablation area of the glacier can be subdivided into two contrasting zones. The upper zone, above ~5,000 m, has an average surface gradient of around 4°, and ice velocities of 20 - 40 m yr⁻¹. In contrast, the lower zone has an average surface gradient of < 1° and is stagnant or nearly so, with velocities < 5 m yr⁻¹. Uneven ablation of the debris-covered ice has produced highly irregular surfaces in both zones. Relative relief is highest in the lower zone, with mounds and ridges up to 50 m high separating topographic lows, many of which are occupied by supraglacial ponds (Iwata and others, 1980; Nakawo and others, 1999). In the upper zone, summer meltwater drains via several incised supraglacial streams that meander across the glacier surface (Figure 3-6). These streams enter englacial conduits in the lower part of the upper zone, and drainage lower down the glacier is predominantly englacial until water re-emerges onto the glacier surface very close to the terminus (Iwata and others., 1980). Water leaves the glacier via a spillway over the terminal moraine, which acts as hydrological base level for the whole glacier system (cf. Gulley and Benn, 2007). Two englacial conduits, located at the downglacier ends of relict and active supraglacial channels, respectively, were surveyed: KH01 in November 2005, and KH02 in December 2006.

KH01

This englacial conduit was located near the western flank of the glacier (27° 59' 5" N, 86° 10' 16" E, 5083 m), and consisted of two segments separated by a bouldery hollow in the glacier surface. The upper segment (KH01A) was a large, meandering canyon, fed at its upper (SE) end by a meandering supraglacial channel originating below the Khumbu Icefall. The lower segment (KH01B) was more constricted and was

partially surveyed but not sketched, for safety reasons. KH01A formed a highly sinuous canyon (sinuosity - 3.21) with an average gradient of 0.03 (Figures 3-7A, B). Canyon walls tilted alternately to the left and right, up towards the inside of meander bends. There were no significant steps in the canyon, which was floored by a porous sub-horizontal mass of ice crystals at least 0.4 m deep, interpreted as an accumulation of frazil ice that masked the true floor morphology. Water was flowing beneath this ice accumulation, and exited at the downstream opening before disappearing beneath bouldery supraglacial debris. The canyon roof consisted of a band of partially ice-cemented debris and ice breccia perforated by windows to the glacier surface (Figure 3-8). The width of the band varied between 2 and 0.3 m.

The upstream entrance of KH01B was extremely constricted and entry was gained by squeezing through a partially refrozen ice breccia. The passage continued as a horizontal slot 0.3 to 1 m high for 40 m. Large detached ice blocks, spalling flakes and ice breccia comprised the right hand wall (Figure 3-9). At a penetration of 56 m, a short side passage extended north through the ice breccia and connected with the glacier surface. Downstream of the junction, the main passage was generally larger, and continued southwest as an inclined canyon for 120 m as the lowest-gradient (0.01) and straightest (sinuosity – 1.17) passage in our sample set. Cross-sections averaged 10 m wide and 2 to 3 m high and the canyon walls inclined steeply to the right where they terminated in a mass of broken ice blocks and frozen ice breccia. Exploration was halted by rising water in the passage.

Supraglacial Channel

The supraglacial channel upstream of KH01A is of interest because of the information it can provide concerning the origin of the englacial reaches. The channel is

a long-lived feature which is visible on satellite images spanning several years. It originates as a series of rills in a field of *penitents* (large cones of ice protruding through very thin debris-cover - see the foreground of Figure 3-9B), below the Khumbu Icefall known as the Pinnacles, and becomes increasingly entrenched once it enters the more thickly debris-covered part of the glacier, where the channel is typically 5 to 10 m deep. The channel exhibits a series of sharp meanders, and in places, a combination of lateral channel migration and vertical incision has undercut the cutback banks, creating ice overhangs which are prone to collapse. At one meander bend, the channel was bridged by a mass of ice breccia about 25 m long and 4 to 5m wide, composed of sub-rounded to angular ice blocks, rock debris and ice cement (Figures 3-9 A & 3-9B). A gap c. 2 m high existed between the base of the breccia and the channel floor, forming an 'englacial' reach of the stream bed. The breccia appears to have formed when collapsed portions of the cutback bank and farther-traveled ice blocks were rafted into place when discharge in the river was high (i.e. during the summer monsoon), and then jammed in the meander bend. Subsequently, additional ice blocks and rock debris fell onto the ice bridge from the glacier surface above. One such rockfall event was witnessed at close range by the survey team.

Interpretation

The supraglacial channel, KH01A, and KH01B are interpreted as successive stages in a continuous evolutionary sequence. Downglacier of the pinnacles, stream incision is faster than surface ablation because the ice is insulated by thick layers of debris. Rapid incision and meander migration encourage formation of overhanging cutback banks (Boyd et al., 2004). When discharge is high, floating blocks of ice can jam in tight meander bends and freeze together, to forming bridges of frozen ice

breccia. Additional debris is added to bridges during meltout and collapse of the overhanging cutback bank and will further reduce surface ablation rates over the channel.

KH01A is a later stage of the cut and fill process. The bands of ice and debris extending along the cave roof and to the glacier surface at both entrances are interpreted as the infill of a canyon incised down from the glacier surface. The former canyon cuts through a topographic high on the glacier, the top of which is 30 m above the upper entrance. This implies that differential ablation of the debris-covered ice created the current topography after channel initiation and downcutting. Debris thickness and albedo have long been recognized as important controls on the topographic evolution on the Khumbu and other Himalayan glaciers (Fushimi and others, 1980; Nicholson and Benn, 2006). Meltback and collapse of cutback banks concentrates debris over nascent cut and closure conduits. This reduces ablation locally and predisposes cut and closure canyon apexes to become new topographic highs.

A third stage of passage evolution is found in KH01B. Extensive fills of angular ice blocks reflect widespread collapse of overhanging canyon walls (Figure 3-7E). Long sections of the passage consist of very low horizontal slots, which were being narrowed further by the accumulation of aufeis on the channel floor. It is likely that creep closure is also an important process at this site, since low, wide passages will close more rapidly than circular or vertical-walled ones in response to ice overburden pressure (cf. Hooke et al., 1990).

KH02

This conduit was located at the margin of a large closed depression near the western flank of the glacier, 440 m SSW of KH01 (*27° 58' 52" N, 86° 50' 07" E, 5,050*

m). Three conduit entrances were present at different points along the outcrop of a prominent, dipping debris band (Figure 10A). A relict incised supraglacial channel extended upglacier from the lower entrance.

The lowermost passage was surveyed in late November 2005 (Figure 3-11A, B). The entrance sloped downwards at 23° and was partially composed of fallen ice blocks cemented by aufeis. The highest observable parts of the passage tapered upward and met at a partially debris-filled seam in the roof, which could be traced intermittently along the entire length of the passage. Passage dimensions diminished with distance from the entrance, and, at a penetration of 30 m, closed down into a tight squeeze 0.4 m wide and 0.3m high (A6), and for 24 m beyond this point, passage dimensions remained very small. Passage morphology in this narrow section consisted of a roughly tubular upper level with a narrow, partially occluded incised slot in the floor. Beyond the narrow section, passage dimensions increased dramatically, opening out into a canyon 18 m high and 3.5 m wide (Figure 3-10B). In some places, light was visible high above the canyon floor, and extensive covers of aufeis coated the walls below these holes. The canyon was highly sinuous (sinuosity of 6.29) and at one point came within 0.5 m of closing on itself. The canyon floor consistently dropped in elevation with an average gradient of 0.11, with lower gradient sections interspersed with a series of frozen waterfalls 0.5 m high. Passage width showed considerable variation, ranging from 0.5 m to 7.0 m. The deepest part of the passage consisted of a room 27 m long, 4.5 m wide and 6.5 m high, which terminated in a frozen-over pool. Fallen ice blocks and coarse rock debris covered the floor. Despite the length of the passage, the innermost end is in close proximity to the entrance.

When revisited in December 2006, KH02 was inaccessible because the entrance was submerged below a frozen pond (Figure 3-10C). From the 2005 survey data, we estimate that the volume of stored water (which was at least partially refrozen in December 2006) was c. 5,300 m³.

KH02: Interpretation

KH02 shares many characteristics with KH01, and we conclude that it formed in a similar way. The inclined debris band above the entrance and the partially debris-filled seam running along the canyon roof are interpreted as a canyon suture marking the former position of an incising stream channel. Closure was accomplished by a combination of infilling by supraglacial debris and ice breccia and creep closure which isolated the canyon from the glacier surface. Although most of the cave exhibited forms indicative of vadose downcutting, the narrow section between A4 and A6 (Figure 11) appears to have experienced a phreatic phase. Because a canyon suture is clearly visible in the roof of the narrow section (Figure 3-11D), we conclude that this part of the cave underwent epiphreatic modification following an earlier phase of vadose incision and partial closure.

The condition of the cave in 2006 suggests that the system may have reached the end of its evolutionary cycle. It appears likely that the blockage at the lower end of the cave in 2005 did not reopen during the 2006 melt season, and that meltwater filled up the cave, flooding it back as far as the entrance. In the following winter season, this accumulated water began to freeze, increasing the integrity of the blockage.

The evidence from Longyearbreen and Khumbu Glacier allows a generic model of conduit formation by 'cut and closure' to be proposed (Figure 3-12):

(1) For supraglacial stream channels to evolve into englacial conduits by a process of cut and closure, channel incision must be faster than surface ablation. This occurs where debris cover or low air temperatures inhibit surface melting. The energy driving incision comes from viscous dissipation, so incision rates are directly proportional to discharge and stream slope (Marston, 1983; Fountain and Walder, 1998). Because surface ablation rates must be low, large catchment areas are required to maintain high discharges. Extensive supraglacial drainage networks can only form in largely uncrevassed regions of glaciers, otherwise streams are captured and routed to the subsurface (Stenborg, 1968, 1973; Boon and Sharp, 2003). Conduit sinuosity will be higher on debris-covered glaciers than on clean glaciers for equivalent stream discharges due to sediment transport (Boyd et al., 2004).

(2) Channel incision occurs most rapidly by upstream nickpoint migration, which may be an order of magnitude faster than floor downcutting.

(3) Drifting snow or jams of rafted ice blocks initiate closure of the upper reaches of canyons. Partial melting and refreezing or burial by supraglacial debris increase the integrity of the nascent bridges. Snow, debris and aufeis can add to blockages at deeper levels where gaps persist in the canyon roof. Ice breccias initiated by internal collapses (autochthonous breccias) or rafting of non-local (allochthonous) blocks can also block deeper parts of englacial canyons.

(4) With increasing depth, canyon walls converge by ice creep. Creep closure can also occur in response to horizontal compressive stresses at low overburden pressures. At shallow depths upstream, roof plugs may contain snow, ice or debris infillings and transition to clean canyon sutures at depth downstream.

(5) If closure processes reduce passage dimensions to the extent that pipe-full flow occurs at periods of high discharge (epiphreatic conditions), or the passage becomes blocked to backflow and discharge at a breach in the ceiling, epiphreatic conditions can transform canyons or horizontal slots into tubular morphologies.

(6) Vadose incision terminates at a glacier's hydrological base level, which may be determined by the elevation of impermeable terminal moraines or the lip of an overdeepening.

(7) Channel floors can cut down to the glacier bed and form subglacial drainage pathways even beneath cold-based glaciers. Where basal ice is cold, such pathways will not communicate hydrologically with the glacier bed on either side of the channel and presence of subglacial water will not influence glacier velocity. In contrast, if cut-and-closure conduits reach unfrozen parts of the beds of polythermal glaciers, water could spill out of the conduit through a subglacial film, permeable subglacial sediments or some other form of distributed drainage system.

(8) Conduits can become blocked by creep closure at depth or by build-up of secondary ice. Conduit blockages can cause the stream course to be re-routed to higher levels, determined by the elevation of the next possible outlet. If no outlet is available, the conduit will fill completely with water and a new supraglacial diversion channel will develop. Observations over several years in Longyearbreen indicate that conduit downcutting and upward re-routing is a cyclic process.

Upstream migration of nickpoints can create vertical sinkpoints at the upglacier end of cut-and-closure conduits, mimicking the morphology of 'tectonic' moulins formed by hydrologically-driven overdeepening of crevasses (Stenborg, 1968). Cut-and-closure

conduit sinkpoints can be easily distinguished from tectonic moulins when there is no snow cover by their sinuous canyon sutures extending downglacier (Figure 3-2A).

We found no evidence that conduits were relict Shreve-type drainage systems inherited from past thermal regimes (Hodgkins, 1997). Inherited Shreve-type conduits would be expected to have multiple tributaries with circular passage cross-sections and plunge steeply to the glacier bed. Instead, we documented long stretches of unbranching, sub-horizontal, canyon passages that, for the most part, did not reach the glacier bed. These crucial observations are incompatible with the predictions of 'classical' englacial drainage theory. The documented long stretches of horizontal passage additionally contradict the Stuart and others (2003) hypothesis that conduits in cold ice are relict drainage features inherited from a former thermal regime.

The observations presented in this paper confirm that englacial conduits can develop by the incision and closure of supraglacial streams, as proposed by Fountain and Walder (1998) and Vatne (2001). However, our results show that the conditions required for conduit initiation, the mechanisms of closure, and the ways in which conduits evolve through time, differ in several respects from previous models and thus constrain their occurrence to predictable locations.

First, the primary condition for conduit initiation and survival is that incision rates exceed glacier surface ablation rates. Pre-existing crevasses are not required to initiate conduit formation. In the Fountain and Walder model, incising conduits are given a 'head start' by initiating at the bottom of a crevasse. However, such a conduit will only persist if incision rates outpace surface ablation, so a pre-existing crevasse is neither a necessary nor sufficient condition for conduit inception. Moreover, while small

discharges are unlikely to generate sufficiently large incision rates, large discharges into crevasses might be expected to encourage deep fracture propagation, routing meltwater vertically towards the bed (Boon and Sharp, 2003; Alley and others., 2005; Van der Veen, 2007; Benn and others, in review). Although shallow or relict crevasses can guide surface drainage patterns (Marston, 1983), crevasses are not expected to be important as nucleation sites for cut-and-closure englacial conduits. Indeed, conduits of this type are much more likely to form on uncrevassed parts of glaciers, due to the requirements that stream contributing areas are large and that discharge is not diverted vertically through fractures.

Second, in all the examples we have studied, accumulation of snow or ice has played a major role in roof closure, and ice creep became important only in later stages of conduit development. Like incision rates, roof blockage processes also appear to be climatically controlled. Snow and ice bridges are most likely to survive the following melt season if surface ablation rates are small. Additionally, effective bridging will only occur where canyons are narrow. Where melt season temperatures are high, supraglacial channels tend to have high width to depth ratios due to melt-back of the channel walls. Wide channel reaches will discourage the formation of ice jams and snow bridges, and their survival through subsequent years. Thus, even where supraglacial stream discharges are large enough to cut deep canyons, cut-and-closure type conduits are unlikely to form.

Third, conduits will not necessarily incise monotonically towards the bed, and drainage can be re-routed to higher levels within the glacier following the blockage of deeper passages. Major blockages occur where closure processes outpace

enlargement processes, and may occur in response to roof collapses, creep closure, or freezing of stored water. Because the chance of winter-season blockage increases with depth, we expect that perennial 'cut and closure' conduits are unlikely to persist below thick ice. There is evidence on Longyearbreen for multiple episodes of downward incision, conduit blockage and upward re-routing. Incision from the surface to the bed takes many years, whereas upward re-routing to the surface can be accomplished in a single season; at this locality, therefore, 'cut and closure' conduits will be located at the bed for a relatively small proportion of their life-cycle.

Observations elsewhere in Svalbard indicate that 'cut and closure' conduits are very widespread. We and other researchers have observed over 20 such conduits on 14 glaciers (Anne Hormes, pers. comm.) and we conclude that 'cut and closure' is the dominant process for forming low-gradient englacial conduits on uncrevassed regions of polythermal glaciers.

These findings establish important boundary conditions for models of polythermal glaciers and the Greenland Ice Sheet. If, as we suggest, cut-and-closure conduits are the only conduits likely to form in unstressed cold ice, they are unlikely to have any influence on the dynamic response of the Greenland Ice Sheet. Conduits will incise to an ice depth sufficient to close conduits in the winter when recharge is absent. The conduit will either become re-established at a higher elevation or, if no other outlet is available, be abandoned. In either case, the conduit will be unlikely to reach the bed in thick ice. And in thin ice, cut-and-closure conduit formation is too slow to explain seasonal velocity transients at the onset of the melt season on land-terminating glaciers (Copland et al., 2003). Because cut-and-closure conduits cannot incise below base

level, they cannot reach the bed of tidewater glaciers and should not cause their observed seasonal speedups (Vieli et al., 2004).

We conclude that the only viable hypothesis for surface melt to access glacier beds through thick ice is hydrofracturing (e.g. Alley et al., 2005; Van der Veen, 2007; Benn et al., submitted), which requires large discharges of water and stressed ice. Applied to the Greenland Ice Sheet, this means that rapid subglacial recharge is most likely to occur in areas of accelerating flow.

Table 3-1. Lowering rates in the Longyearbreen supraglacial channel and ablation rates for the adjacent glacier surface.

	Channel lowering (cm day ⁻¹)			Surface ablation (cm day ⁻¹)		
JD 209 – 251	Mean	Max.	St. Dev.	Mean	Max.	St. Dev.
Cross section 1	10.4	32.9	9.9	2.2	5.7	1.2
Cross section 2	13.1	33.1	5.6	2.6	4.9	1.4
Cross section 3	4.1	8.7	2.5	2.2	7.2	1.5
All Sites	9.2			2.3		
JD 209 – 230						
Cross section 1	17.9	32.9	1.3	3.0	5.7	1.3
Cross section 2	14.5	33.1	7.6	3.5	4.9	1.6
Cross section 3	6.1	8.7	2.3	3.4	7.2	1.5
JD 231 – 251						
Cross section 1	2.8	2.5	0.4	1.5	2.2	0.4
Cross section 2	11.7	12.2	1.7	1.8	1.9	0.3
Cross section 3	2.3	2.7	0.7	1.2	1.3	0.1

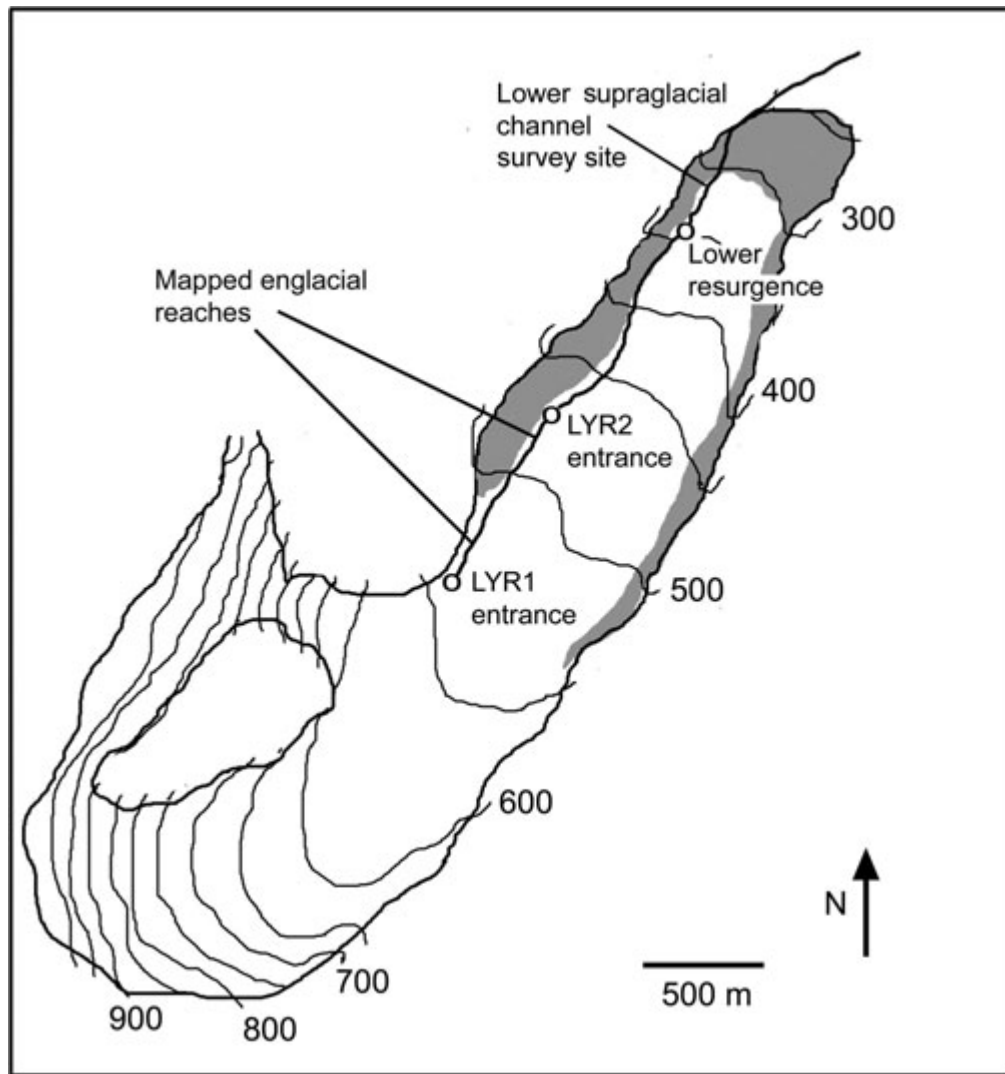


Figure 3-1. Map of Longyearbreen, showing the western drainage system and the location of cave entrances. Ice-cored moraine is indicated in gray tone.



Figure 3-2. Photos of passage morphologies. A) meandering snow plug at the surface of Longyearbreen above LYR 1, B) subglacial section of the western drainage system, March 2001. This section of the bed is located below the lower end of LYR 2, and consists of stony regolith capped by *in situ* soil and vegetation (Photo: Ole Humlum), C) nickpoint in the unsurveyed downstream section of LYR1 and D) a plug of snow, aufeis or neve was traceable along the apex of the canyon passage for the entire surveyed length of LYR1.

LYR1

Plan View

Longyearbreen, Svalbard, Norway

October, 2006

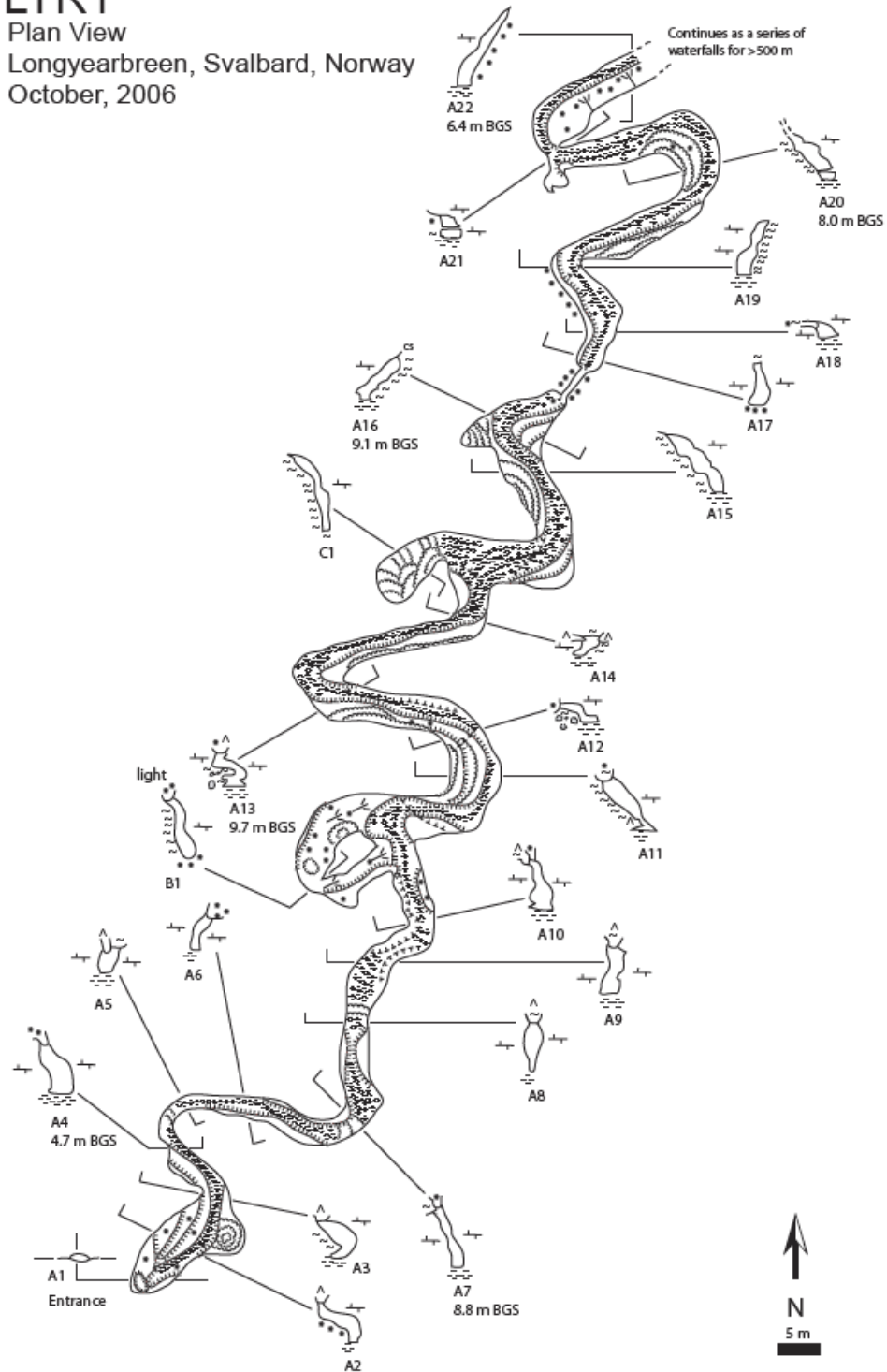


Figure 3-3(A). Maps of conduits surveyed on Longyearbreen. Depth below glacier surface (BGS) is indicated at approximately every fifth station in the plan views. LYR1 plan view.

Plan View
Longyearbreen, Svalbard, Norway
October, 2006



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Legend	
	Passage Walls
	Passage Continues (too constricted to continue)
	Change in Ceiling Height
	Aufeis Flow
	Rock
	Glacier Ice Floor
	Change in Floor Height
DB	Debris Band
~	Flowing Aufeis (Cross Section View)
*	Snow
	Cobbles, small boulders
Fol	Foliation
CT	Debris Filled Crevasse Trace
F	Flake
	Slope in Floor
	Glacier Ice Boulder
S	Sample Location
	Water Ice Floor
hoar	Hoarfrost
CrS	Crossbedded Sand
CS	Canyon Suture
AE	Nonflowing Freeze-on Aufeis

Figure 3-4. Legend for conduit survey maps.



Figure 3-5. A) canyon suture at A32 (LYR2), B) the upper reaches of a canyon have been squeezed shut to create this horizontal slot in LYR2, a low-wide passage developed from lateral incision of cutback banks in meander bends. This photograph is an upstream view of the lower level meander cutoff between stations A39 and A50 and is shown as a grey passage outline in Fig 3A, C) passage cross-sections between A3 and A7 appear to be a hybrid between phreatic (tube) and vadose (canyon) cross-sectional morphologies and likely represent transient phreatic conditions and D) in July 2007, the entrance used to access LYR2 was discharging water. Water upwelled from the entrance, flowed over the glacier surface for a short distance and discharged to a supraglacial channel which was still roofed with snow when this photograph was taken.

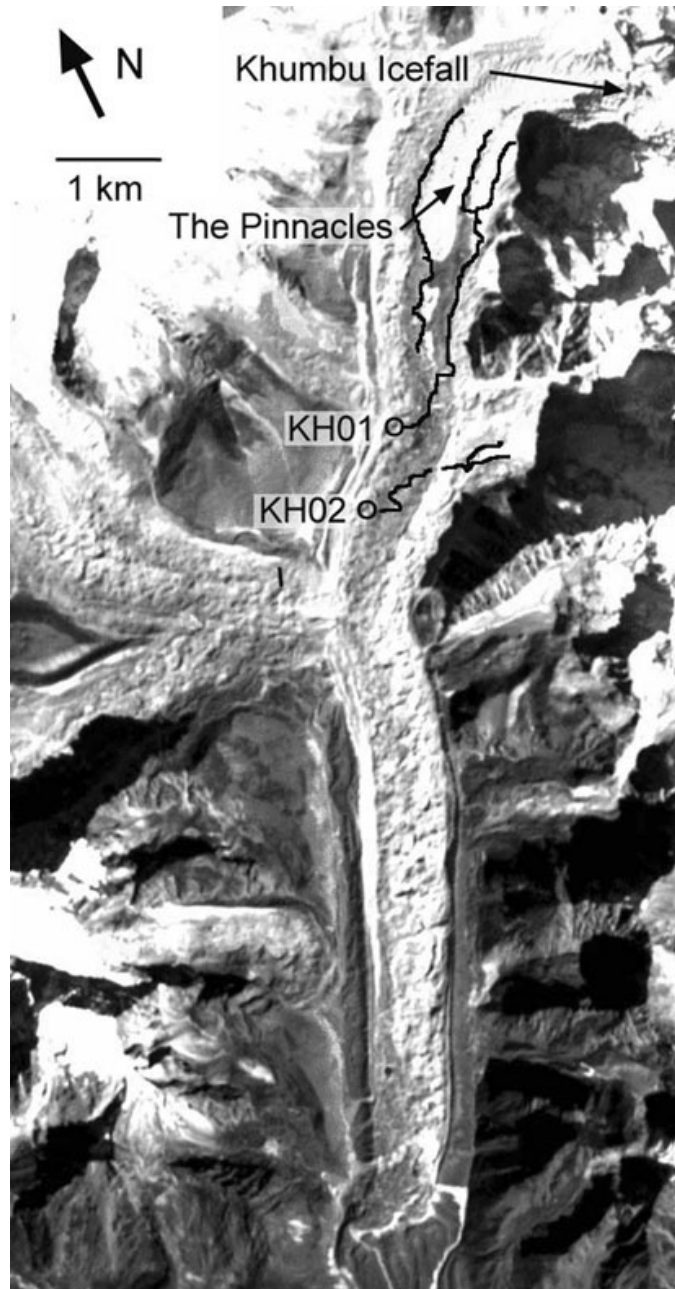


Figure 3-6. Aster image of Khumbu Glacier, December 2005, showing locations of cave entrances and supraglacial channels.



Figure 3-7. Photos of KH01. A) lower (northern) entrance to KH01A, B) the canyon walls of KH01A tapered to either a debris band with a dense ice matrix or ice breccias and C) debris-band near the northern entrance of KH01A.

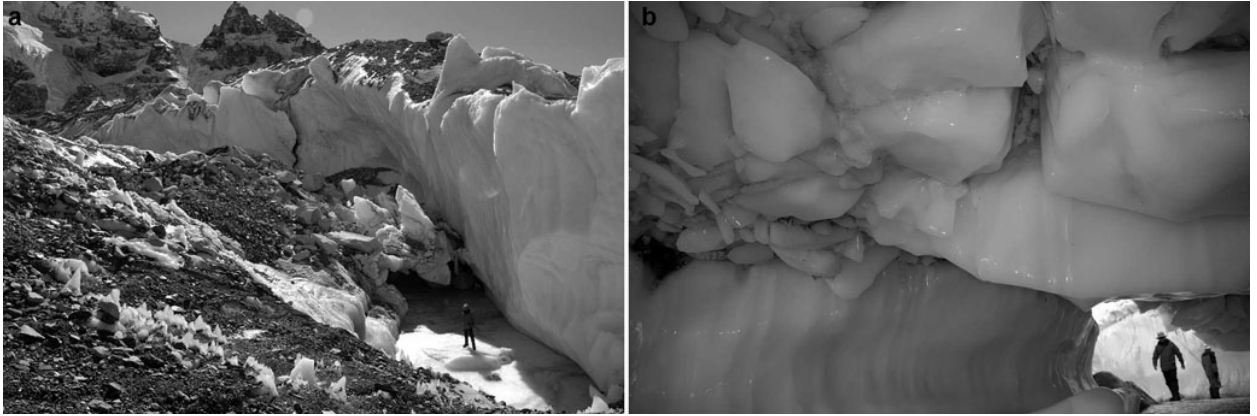


Figure 3-9. Photos of the supraglacial stream on the Khumbu Glacier. A) upstream of KH01A, ice breccia bridges the supraglacial channel and B) ice breccia roof.

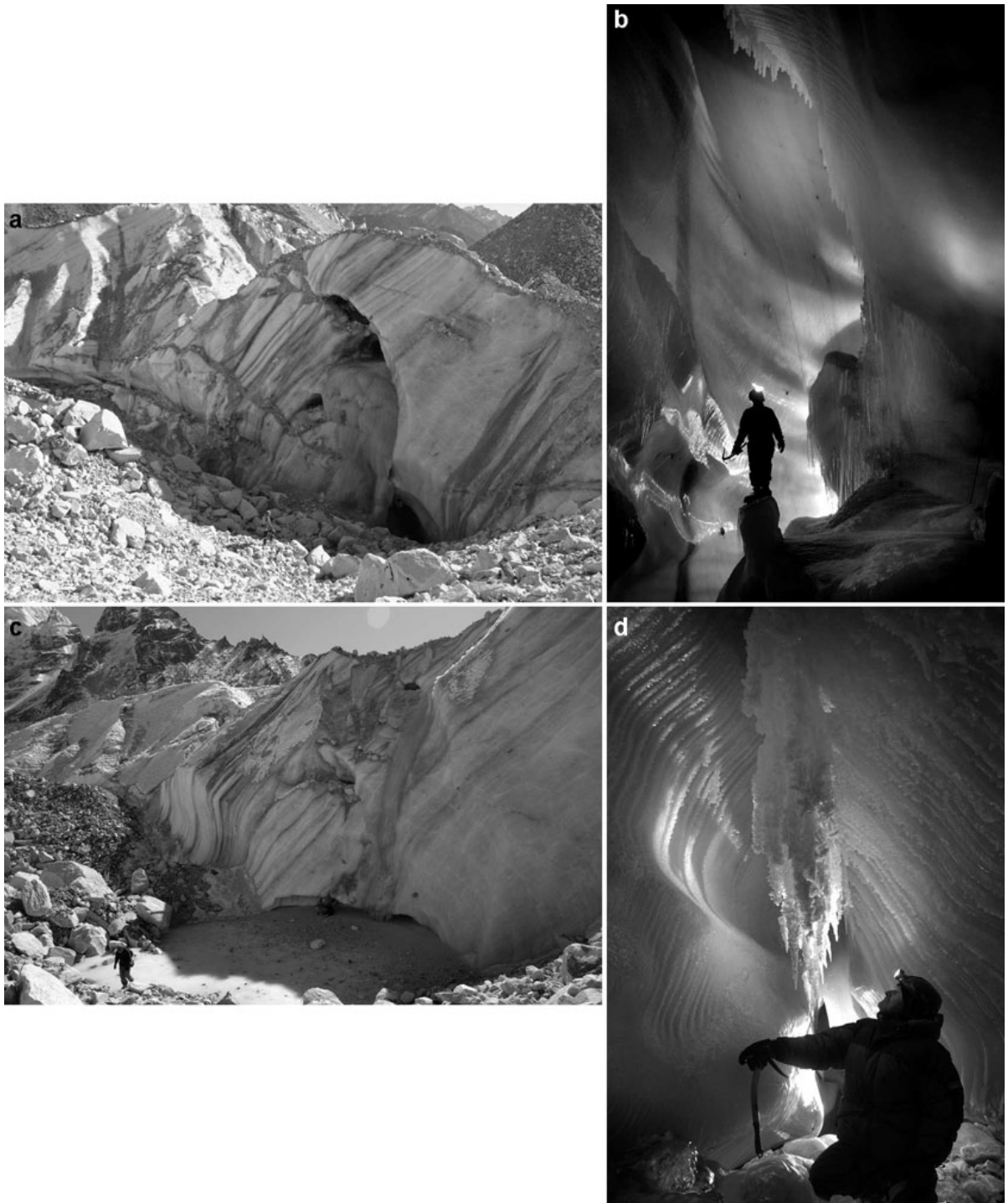


Figure 3-10. Photos from the Khumbu Glacier. A) entrance of KH02 in November 2005. Note the debris-band and isolated canyon segments partially obscured by waterfall ice, B) canyon passage in KH02 near station A13, C) the entrance of KH02 in December 2006. The figure in the photo is walking on a frozen lake, beneath which was located the entrance used in 2005 and D) roof suture in the restricted part of KH02.

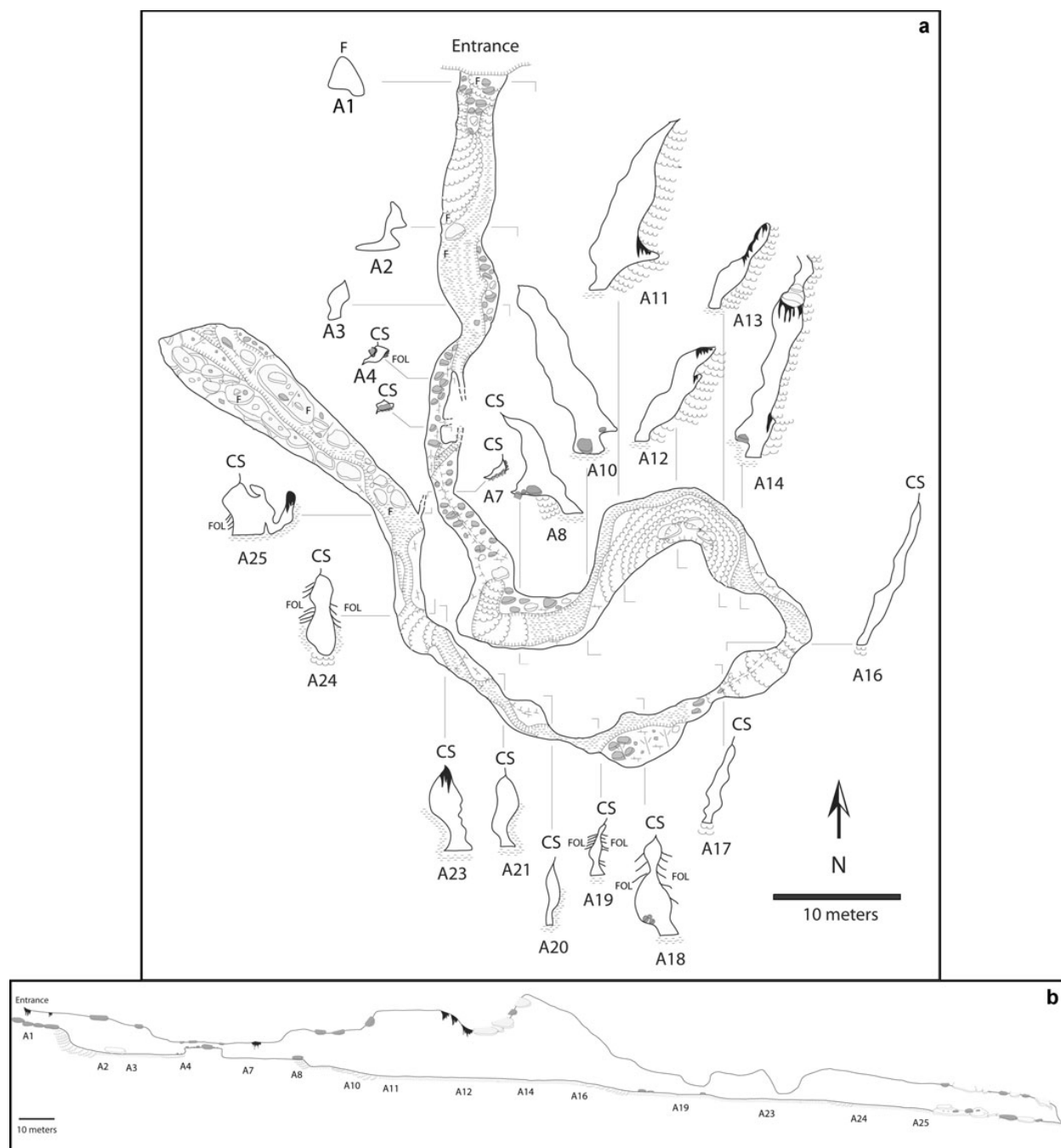


Figure 3-11. Maps of KH02. A) plan and cross-sectional views of KH02 and B) profile view of KH02.

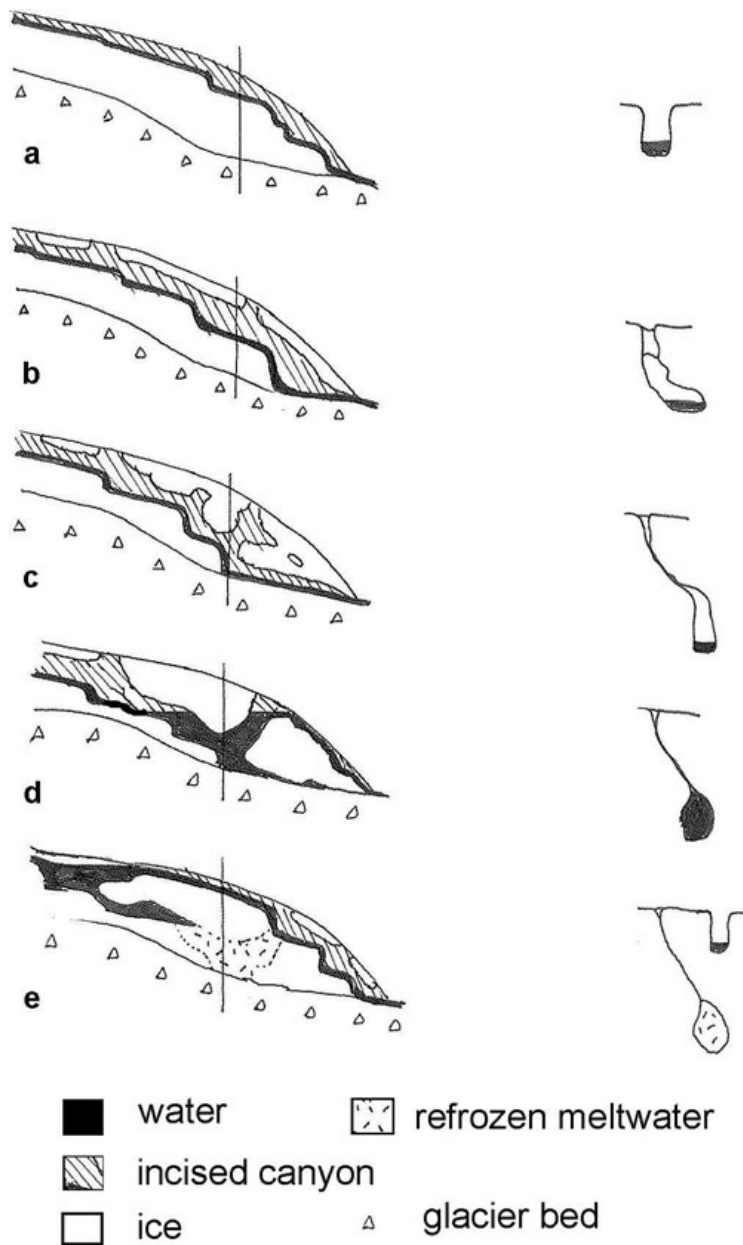


Figure 3-12. Conceptual model of conduit development by cut and closure. Left: long profile; right: cross sections. A) conduit begins as a supraglacial stream, B) if incision is faster than surface ablation, the conduit cuts deeper. Upper reaches of canyons become plugged with snow and aufeis, C) continued incision, and closure by aufeis accumulation and ice creep, D) lower levels become plugged by aufeis accumulation or creep closure, water backs up to discharge from the next available pre-existing outlet and E) winter freezing of ponded water propagates the blockage upstream, and in the following summer water finds the next lowest outlet point. Water flowing onto glacier surface incises new channel.

CHAPTER 4

STRUCTURAL CONTROL OF ENGLACIAL CONDUITS IN THE TEMPERATE MATANUSKA GLACIER, ALASKA, USA

Englacial conduits are a critical component in glacier hydrological systems because they can rapidly convey large volumes of meltwater from glacier surfaces to glacier beds (Hooke, 1989; Fountain and Walder, 1998) where subglacial storage elevates basal water pressure and causes a transient increase in glacier velocity (Iken and Bindshadler, 1986; Willis, 1995). Despite their importance, little is known about the spatial distribution, mechanisms of formation or the longevity of englacial conduits. Conceptual models of englacial drainage have been deeply influenced by the theoretical model developed by Shreve (1972). According to this widely-cited model, englacial hydrological systems should consist of arborescent systems of conduits that evolved from intergranular veins by exploiting the primary permeability of ice, roughly analogous to Darcian flow in a homogeneous, isotropic medium. This model predicts that individual conduits should trend normal to equipotential surfaces controlled by elevation, ice overburden pressure and conduit radius. Because of its requirement that glacier ice is permeable, Shreve's theory only applies to englacial conduits in temperate glaciers because ice below the pressure melting point is impermeable.

There is no direct evidence, however, that Shreve-type conduits exist in temperate glaciers and several other theories of englacial conduit formation have been advanced. Stenborg (1968, 1969) initially proposed that englacial conduits developed from crevasses. More recently, it has been suggested that englacial conduits could form when supraglacial streams incised along crevasse bottoms and either reached the glacier bed or entered a Shreve-type englacial drainage system at depth (Fountain and Walder, 1998). Investigation of boreholes on Storglaciern, in Sweden, using downhole

cameras led Fountain et al., (2005) to conclude that englacial drainage systems are dominated by slow flow in fractures and that conduits only form under unspecified special circumstances. This theory also does not completely explain englacial conduit formation since slow flow in fractures is inconsistent with the sharply peaked hydrographs diagnostic of integrated conduit flow observed on many glaciers (e.g. Swift et al., 2005).

Recent speleological investigations of polythermal and debris-covered glaciers have demonstrated that englacial conduits can form by multiple mechanisms. These studies have shown that conduits form where high-hydraulic conductivity glacio-structural features connect discrete recharge and discharge points (Gulley and Benn, 2007; Benn and others, in press; Gulley et al., submitted). Additionally, speleological mapping of moulins in the polythermal Storgalciern, Sweden, found conduits followed the orientation of the crevasse even at depths exceeding 60m (Holmlund, 1988).

Other than direct observations of englacial conduits in primarily polythermal and cold-based glaciers (Gulley et al., in press; Benn et al., in press), most theories of englacial conduit formation have been developed by interpreting proxy data such as dye-tracing and geophysics (c.f. Willis et al., 2009). These proxy data limit understanding of the physical processes controlling their formation and distribution. Determining where, when or if each of these theories is applicable requires substantial direct observations as exemplified by studies of conduit morphologies in limestone, which have greatly facilitated interpretation of proxy data and refinement of numerical models (White, 1988; Palmer 1991; Ford and Williams, 2007; Palmer, 2008). Similar to these studies of limestone conduits, morphological studies of englacial conduits should

improve understanding of mechanisms of englacial conduit formation and glaciohydrological modeling efforts. This paper extends the speleological approach applied by Gulley and Benn (2007) in polythermal debris-covered glaciers, Gulley and others (in press) in polythermal and cold-based glaciers and Benn and others (in press), primarily in polythermal glaciers, to englacial conduits found in a temperate glacier. This paper reports the results of a three-year englacial conduit mapping effort at the Matanuska Glacier, Chugach Mountains, Alaska, USA, which includes detailed, three dimensional maps of englacial conduits and their relationship with glaciostructural features. These results are used to demonstrate substantial differences between observed englacial conduit morphologies and those predicted by the Shreve (1972) model.

Study Area and Methods

Matanuska Glacier in the Chugach Mountain Range of south-central Alaska began advancing around September 2002, impacting stagnant buried ice in the terminal zone (Baker and others, 2003; Chesley and others, 2005). Thrusting and deformation of proglacial sediments and buried ice have been reported in the near terminus region (Baker and others, 2003; Pyke et al., 2003) and folds in laminated sediments near the terminal ice margin were observed in this study. Longitudinal crevasses extend ~2 km upglacier from the terminus, immediately above which ice is less densely crevassed and a well developed supraglacial stream network is present. Upglacier-dipping transverse crevasses occur near the terminus and, during this study, were commonly associated with vents discharging supercooled water (Lawson and others, 1998; Alley and others, 1998). Several vents were found near the intersection of transverse and longitudinal crevasses during this study. The combination of these structural features leads to the

characterization of the stress field of the central crevassed glacier tongue as longitudinal compression and transverse extension. Crevasses in the area where active ice flows past slower debris-covered ice on the northeast margin (Shumway et al., 2005) and along the northwest lateral moraine are interpreted to be shear crevasses and frequently cut across both longitudinal crevasses and/or transverse crevasses. For logistical reasons, fieldwork was limited to within 3 km of the glacier terminus.

Conduits were surveyed on three expeditions between 2005 and 2007 using standard speleological techniques (Palmer, 2008) modified for glacier caves (Gulley and Benn, 2007). Distance was measured between two stations in a conduit with a Leica Disto laser distance meter. Azimuth and inclination were measured using Brunton Sightmaster and a Brunton Clinomaster respectively. Cross-section measurements were made at each station using the Disto. This process was repeated until the entire conduit had been surveyed. These data were then used to draw scaled geomorphic maps of conduits in plan, profile and cross-section views.

Two conduits were explored but not surveyed in October, 2005, nine conduits were surveyed in September, 2006, and six conduits were surveyed in October 2007. Conduits that remained open between years were revisited to document their evolution. Surveys of englacial drainage networks were conducted at the end of the ablation season when meltwater flow had largely ceased but passages remained fully open. Survey data were reduced using the COMPASS cave survey software program, from which planimetrically accurate maps were drawn. Patterns of strain in the ice were inferred at each site by crevasse pattern analysis (cf. Nye, 1952; van der Veen, 1999).

Results

A total of 15 conduits were explored and 14 were surveyed to a maximum ice depth of 65 m. Mapping ended in seven of the conduits because they were water filled and ended in eight of the conduits where constrictions were too small to pass through. All conduits showed clear structural controls and fall into two separate classes based on glacier stress patterns and conduit morphology. Twelve conduits that formed along shear-crevasses or transverse crevasses comprise one class. These conduits were investigated along the northeastern clean-ice boundary within 2 km of the glacier terminus, where active ice shears past slower or stagnant, debris-covered ice (Shumway et al., 2005). The remaining three conduits (DP 1-3) were formed at the upglacier limit of three adjacent longitudinal crevasses and are placed into the second group. The second class of conduits was located in a basin (DP in Figure 4-1; site names used in this paper, e.g. 'DP', are abbreviations of field designations) at the confluence of three supraglacial streams about 2.1 km from the glacier terminus. The locations of all conduit entrances are shown in Figure 4-1. No additional conduit entrances large enough for human entry were found in the serac zone between the DP basin and the terminus despite extensive searches during all three field seasons.

Transverse Crevasse and Shear Zone Conduits

Entrances for this class of conduits formed where shear crevasses and/or transverse crevasses intersected supraglacial streams. Conduits associated with shear crevasses had vertical shafts or enlarged fissure entrances (Fig 4-2a). Entrance shaft depths increased systematically from a few meters near the terminus to 20 m at the furthest upglacier site (SC – Figure 4-1). Six of the conduits were simple shafts or fissures following shear crevasses. Five conduits formed along shear crevasses that

intersected and exploited transverse or other crevasses at depth and one conduit (CP) formed along a transverse crevasse only (Figure 4-2c). Conduit orientations changed, often dramatically, at fracture intersections (Figure 4-3 and 4-4a,b – a legend for englacial conduit maps is provided as Fig 4-4c). Conduits that formed along intersecting shear crevasses and transverse crevasses change direction at the fracture intersection and follow the transverse crevasse obliquely to strike and dip. Conduits branching off from the shaft entrance (shaft drain) (Figures 4-4 & 4-5) displayed a classic 'key hole' cross sectional morphology indicative of a transition from phreatic to vadose flow (Gulley and Benn, 2007). One tube-shaped conduit (WD – Figure 4-1) increased in elevation in a downstream direction (Figures 4-5 & 4-6) indicating that water flowed uphill under pipe-full conditions. Conduit WD's shaft drain extended from the entrance shaft as a horizontal, phreatic tube following a shear crevasse (Figure 4-6a). Three meters from the entrance, the tube increased 1.5 m in elevation over a distance of 4.9 m before turning 63° and plunging 45° along an intersecting crevasse and terminating in a pool of water. A narrow vadose canyon with migrating nickpoints incised the middle of the tubular cross-sections (Figure 4-6) and was graded to the terminal pool where the conduit continued underwater.

The evolution of conduit CP is more complicated than that of conduit WD. In 2006, conduit CP was formed entirely along a transverse crevasse and the conduit was draining in an up-glacier direction down the dip of the crevasse. In 2005, the same location hosted an englacial conduit formed along a shear crevasse (Figure 4-7), which drained toward the clean ice margin. In 2005, this conduit was followed for 70 m down a series of progressively higher nickpoints until it terminated in a pool of water at the

bottom of 10 m deep shaft. A fracture was visible in the ceiling of the conduit for its entire length. When revisited in 2006, the conduit roof had melted completely to create a nearly linear supraglacial canyon. Conduit CP had formed in the middle of the supraglacial channel sometime in the ablation season of 2006 when the supraglacial stream incised deeply enough to expose part of an intersecting transverse crevasse. All of the stream discharge was pirated from the supraglacial channel into the englacial conduit which formed as water flowed down the dip of the transverse crevasse in an up-glacier direction. From Figure 4-2b, it can be seen that water briefly exploited the transverse crevasse before developing a free-surface stream and vadose incision created the canyon resulting in the T-shaped cross-section.

Partial exploitation of a transverse crevasse can be seen in conduit MS, whose trajectory was guided entirely by a separate, longitudinal crevasse. The conduit formed as a phreatic tube along this crevasse except for where it intersects a transverse crevasse (A4, Figure 4-8). The transverse crevasse was enlarged in the vicinity of the conduit in an up-dip direction resulting in a composite cross-section, midway between a tube and a T-shape (A4, Figure 4-8).

Compression-zone Hydrostatic Crevasse Penetration Conduits

In 2007, three conduits (DP1-3; Figure 4-1) were surveyed in a deep, supraglacial basin where three supraglacial streams sank into englacial conduits. Nickpoint migration and melt-back of the channel walls had created a basin approximately 30 m deep and the conduit entrances were located in the bottom of the basin. All three conduits were fracture-guided, low sinuosity passages plunging at angles between 30° and 40°. The conduit entrances were located on adjacent longitudinal crevasses. Conduit DP1 was actively receiving water from a supraglacial stream. Conduit DP1 (Figure 4-9) was

separated from DP2 (Figure 4-10) by a 4 m high meander cutbank at the top of which was an abandoned supraglacial channel leading into the entrance of DP2. DP3 (Figure 4-11) was located in the third longitudinal crevasse from the active stream.

Conduit DP2 was briefly explored during 2006 (Benn et al., in press), but unseasonably warm weather that rapidly melted out ice screws and high flow in the conduit stream prevented access. In 2006 the accessible portions of the conduit had vertically oriented, lenticular cross-sections with crevasse traces visible in the floor and the ceiling of the cross-sections (similar to cross-sections in DP1 and DP3). In 2007, much of the ice face in which DP2 was developed had melted out and the conduit dimensions rapidly diminished with depth and pinched out completely at 22 m below the conduit entrance (~52 m ice depth).

Conduit DP3 was formed along the longitudinal crevasse furthest from the main active stream suggesting it is the oldest conduit. It was in an inaccessible portion of the basin in 2006, however, so its developmental history remains uncertain. Regardless of when it formed, DP3 exhibited strong fracture control and plunged toward the glacier bed at a 30° angle. Crevasse traces extended from the ceiling of all conduit cross-sections and were frequently noted where the conduit floor was visible. In cross-section and profile views DP3 clearly demonstrates that some parts of the crevasse were not conducive to conduit formation. Cross-sections A4, A14, A15 and B1 (Figure 4-11) show intact “bridges” of glacier ice bifurcated by a crevasse trace (see also Figure-12a). DP3 terminated in a pool of water ~65 m below the ice surface but the conduit was observed to continue underwater.

The history of DP1 is well constrained. In 2006, all drainage entered DP2 from the main supraglacial channel via a 15 m high nickpoint, but sometime between the winter of 2006 and fall of 2007, DP1 formed in the middle of this supraglacial stream and pirated all stream discharge. Paired sills that form by vadose incision (cf Gulley et al., in press) at the entrance indicate the supraglacial stream had incised ~2.4 meters since the conduit formed under phreatic conditions. DP1 is similar to DP3 in plan, profile and cross-sectional views. Again, vertically-oriented, lenticular cross-sections taper into crevasse traces in the ceiling and floor (Figure 4-12b). Cross-sections A4, A7 and A8 (Figure 4-9; see also Figure 4-12a) show that only portions of the initial fracture were exploited. Discontinuous fractures (Figure 4-12a) suggest fracturing may have occurred in stages or in narrow swarms. DP1 was surveyed to an ice depth of ~60 m where the conduit became too narrow to continue.

All conduits mapped for this study were unbranching, followed glaciostructural features and did not lead into an accessible arborescent network of passages. Despite the hydraulic low that these conduits create within temperate ice, they did not drain systems of infeeder conduits nor did they enter a Shreve-type drainage system. Because the bed was not reached in any of the conduits, the possibility that these conduits entered a Shreve-type network at greater depth cannot be ruled out by this data set but based on observations of englacial conduits in many other glaciers located world-wide (Gulley et al., submitted) this is considered unlikely. All conduits cross-cut foliation and bubbly ice veins without deviation. These observations indicate that temperate glacier ice is not sufficiently permeable to influence englacial water flow at the macroscopic scale. Rather than the simple, circular cross-sections predicted by

classical theory (Shreve, 1972; Rothlisberger, 1972), cross-sections were oriented on fractures instead of in intact glacier ice and varied from planar to tubular where they formed under phreatic conditions and transitioned to canyons when free-surface streams developed. The differences in observed conduit morphology with Shreve theory result from the model's primary assumption of permeable ice and that it does not consider that fractures provide high hydraulic conductivity pathways through effectively impermeable glacier ice.

For several reasons, conduits in this study cannot have formed as free-surface streams that flowed along crevasse bottoms and later became isolated by creep closure of the upper crevasse walls, a mechanism Fountain and Walder (1998) proposed for englacial conduit development in temperate glaciers. The clear association of all conduits with fractures in conduit ceilings and especially floors (Figure 4-12b) precludes this possibility. If conduits had formed by incision along crevasse bottoms, canyon sutures (cf Gulley et al., in press) and not fractures, would have been present in conduit ceilings and there would be no fractures in conduit floors. Ice bridges separating upper and lower conduit segments in D1 and D3 (Figure 4-12a) additionally argue against this formation mechanism. Finally, conduit CP and an, unmapped conduit formed along an adjacent shear crevasse were crevasse-bottom streams in 2005, but when revisited in 2006, the original conduits had melted out completely. The unroofing of these englacial conduits occurred between October 2005 and September 2006, reflecting the fact that surface ablation rates outpaced downcutting rates in the ice despite the conduits having a "head start" by forming within the glacier along a crevasse bottom. The unroofing is a consequence of high ablation-season temperatures and the small contributing areas

(and hence discharges and incision rates) of surface streams in the crevasse field. Englacial conduits can form by the incision of supraglacial streams; however, the environmental conditions necessary for this mechanism of formation (termed 'cut and closure') appear to restrict them to uncrevassed regions of polythermal and debris-covered glaciers (Gulley et al., in press).

While all conduits in this study formed along fractures, there are important differences in the mechanisms of formation between those formed along shear crevasses and transverse crevasses and the conduits in the DP basin. Both shear crevasses and transverse crevasses are pre-existing fractures which, if they are kept open by continued slip during glacier motion create discrete zones of high hydraulic conductivity. Because these transverse crevasses already connect to the bed, there is no need for water to hydrostatically penetrate them from the top-down. High basal water pressures may however be important for their formation (Ensminger et al., 2001). Shear crevasses provide vertically oriented fracture pathways that readily capture supraglacial streams and provide direct access to the transverse crevasses to form conduits from the top down. Alternatively, transverse crevasses can function as englacial discharge features if subglacial water pressure is high (Ensminger et al., 2001). Transverse crevasses interpreted to be thrust faults have formed low and wide conduits discharging subglacial water in the compressive tongues of polythermal glaciers in Svalbard (Mavlyudov, 2005), thus forming englacial conduits from the bottom-up. Transverse crevasses might therefore be important for both subglacial recharge and discharge in glacier compression zones such as where glaciers, like the Matanuska, advance into an

ice-cored moraine (Baker, 2003) or at the leading edge of surge bulges in polythermal glaciers (Murray et al., 1998).

The three conduits in the DP basin formed at the up-glacier limit of longitudinal crevasses associated with the compressive glacier tongue. All three conduits clearly followed longitudinal crevasses and attained glacier depths of 65 m, which is much deeper than typical 'dry' crevasses (van der Veen, 1998). It is concluded that these conduits formed when a combination of longitudinal compressive stress and water pressure led to hydrostatic crevasse penetration (e.g., Benn et al., in review). Discontinuous segments of over-deepened fractures and intervening ice bridges suggest fracturing either occurred in swarms or in stages (Figure 4-12a).

Because of the clear association of all investigated conduits with fractures, it is proposed that englacial conduits can only form on temperate glaciers either where large amounts of supraglacial meltwater are diverted englacially along a preexisting line of high hydraulic conductivity linking recharge and discharge points or where high hydraulic conductivity zones are created by crevasse penetration in stressed ice. Englacial conduits are unlikely to form in uncrevassed parts of temperate glaciers because the surface ablation rates and supraglacial stream incision rates are too similar to allow conduit formation by the cut-and-closure mechanism (Gulley et al., in press). Indeed, no conduits were found in uncrevassed parts of the Matanuska glacier.

The hypothesis of structural control of conduits is attractive for many reasons. Hydrologically connected fractures are abundant throughout low-permeability glacier bodies (Fountain et al., 2005; Pohjola, 1994), and, where present, would create preferential flow paths to form conduits if they are connected to supraglacial recharge

sources such as lakes and streams. Structural control of englacial conduits explains conduit formation in both temperate ice as well as cold ice. If conduits require a combination of fractures, which are a function of a glacier's principal stresses, and a large source of supraglacial recharge, the locations of englacial conduits, and by extension, the general location of discrete subglacial recharge become predictable. This concept, which builds on modern karst geomorphology and hydrology, is reviewed in greater detail in Gulley et al. (submitted).

This research on conduits in the temperate Matanuska Glacier builds on past work in polythermal, debris-covered and cold-based glaciers that has systematically demonstrated the formation of englacial conduits requires preexisting lines of high hydraulic conductivity linking recharge and discharge points (Gulley and Benn, 2007; Gulley et al., in press; Benn et al., in press). The recognition that zones of high hydraulic conductivity is a prerequisite for conduit development in temperate glaciers is particularly important because classical englacial hydrological theory was initially developed specifically for temperate glaciers. No evidence of Shreve-type conduits was found within the accessible parts of any conduit explored within 3 km of the terminus of the temperate Matanuska Glacier. If intact glacier ice were sufficiently permeable to form conduits, arborescent tributary conduits should have been found feeding into the structurally controlled, low-pressure conduits. No such infeeders were discovered. Instead, all conduits followed fractures for their entire observable length. The observations presented in this paper suggest that the concept of Shreve-type drainage systems in temperate glaciers may need re-evaluation.

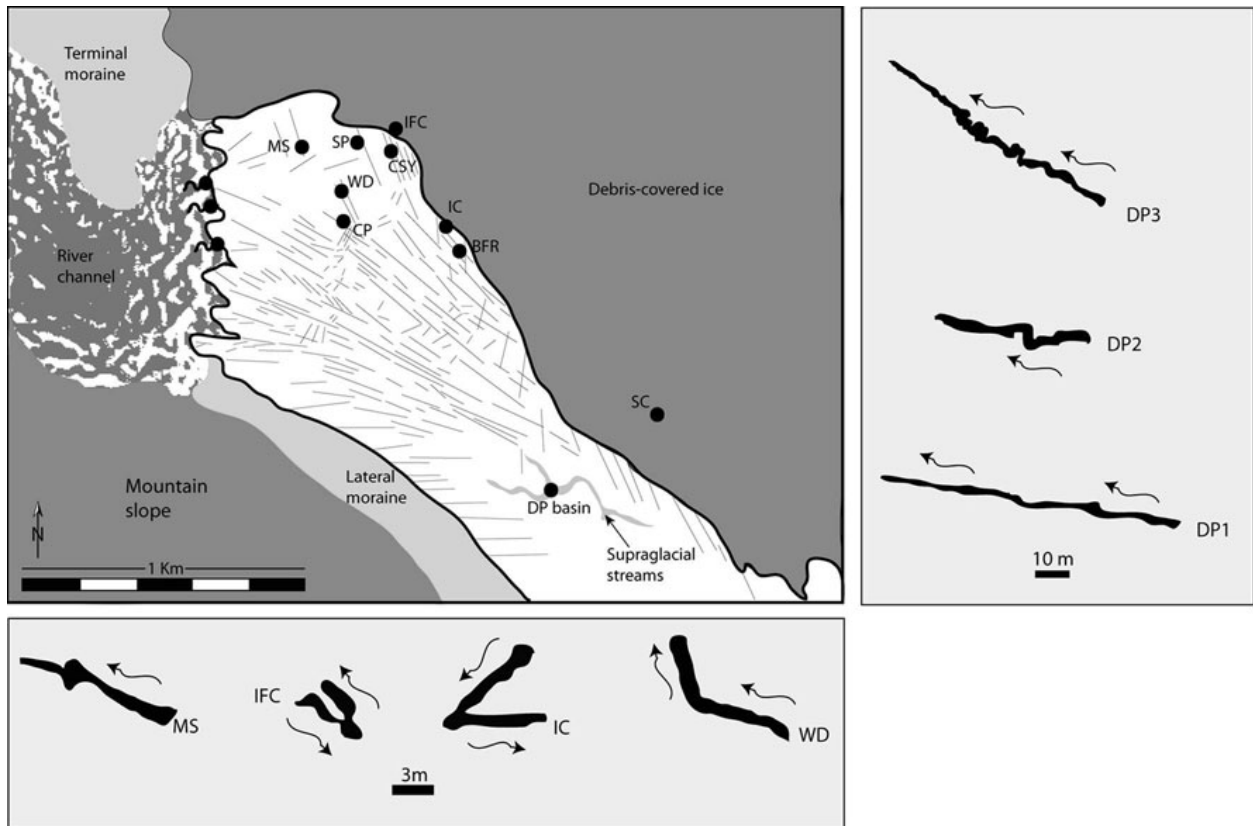


Figure 4-1. Location of englacial conduit entrances and their relationship to crevasse patterns on the Matanuska Glacier. Plan views of the conduits are included to show the general relationship between conduits and the overall fracture patterns of the terminus region. Arrows indicate the direction of water flow.

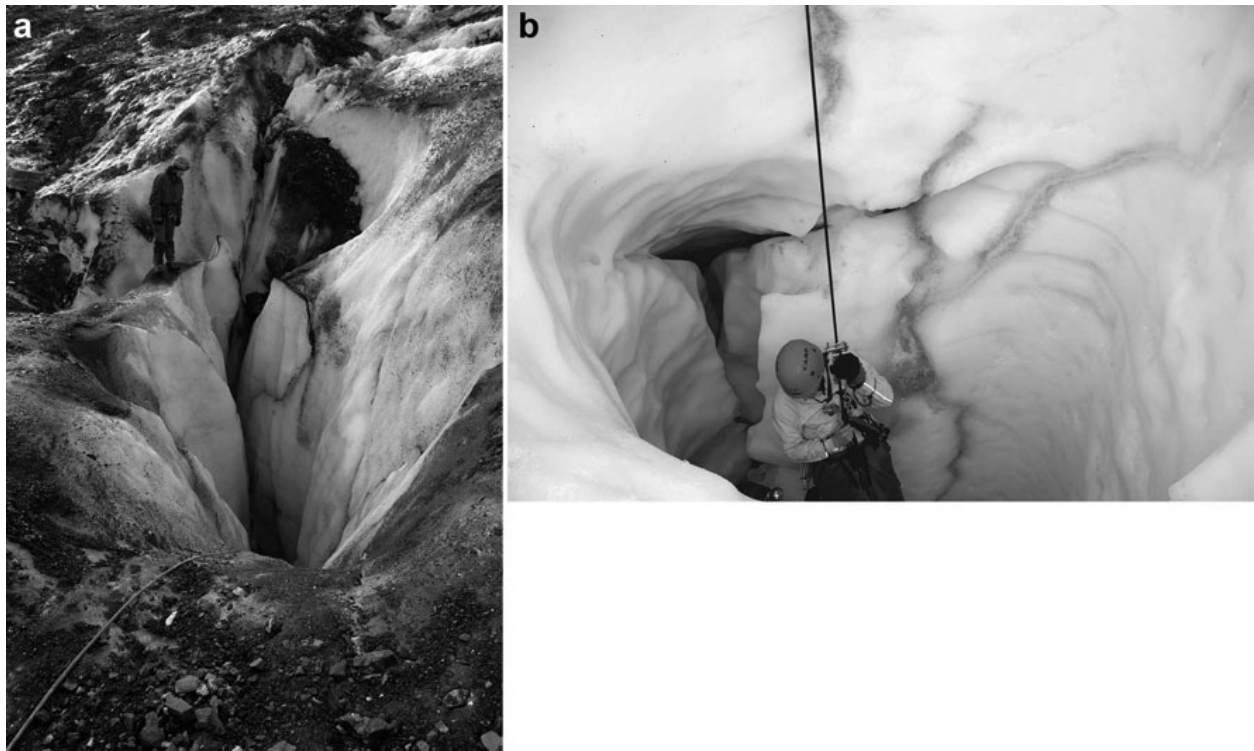


Figure 4-2. Photos of conduits guided by crevasse traces. A) the entrance to englacial conduit IC is located along a shear crevasse formed where active ice shears past dead, debris-covered ice on the Matanuska's northern margin and B) conduit CP briefly exploited this thrust fault before developing a free-surface stream and incising as a vadose canyon to create this T-shaped cross-section.

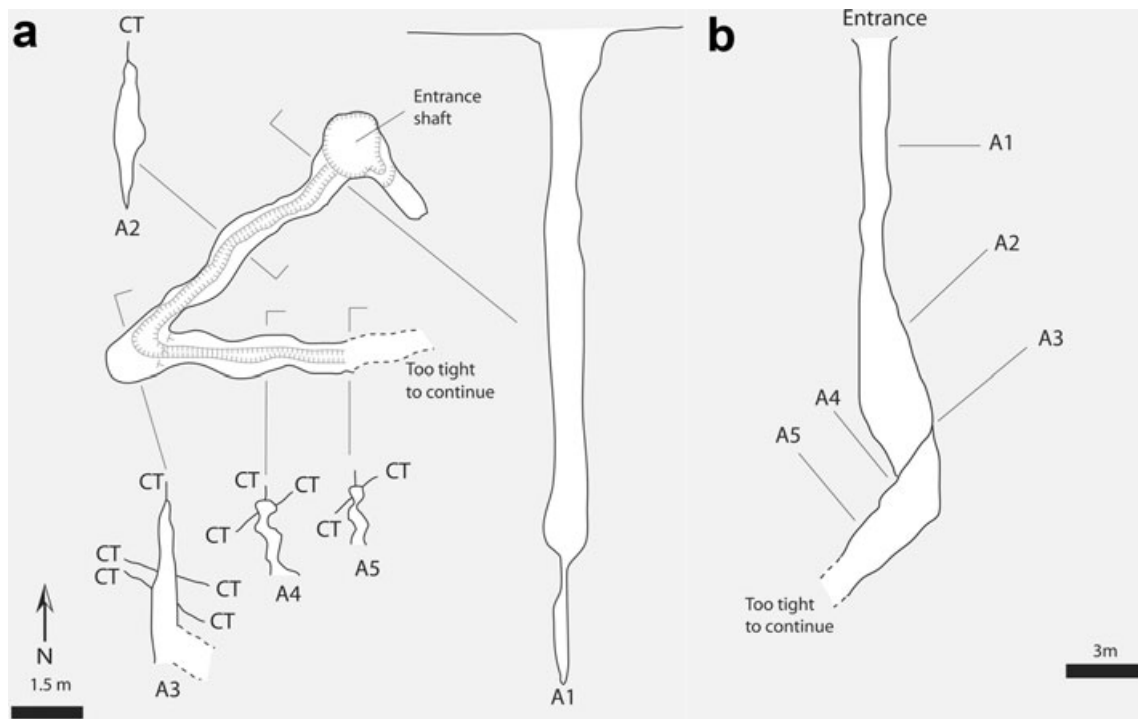


Figure 4-3. Maps of conduit IC. A) plan view of conduit IC and B) profile of conduit IC.

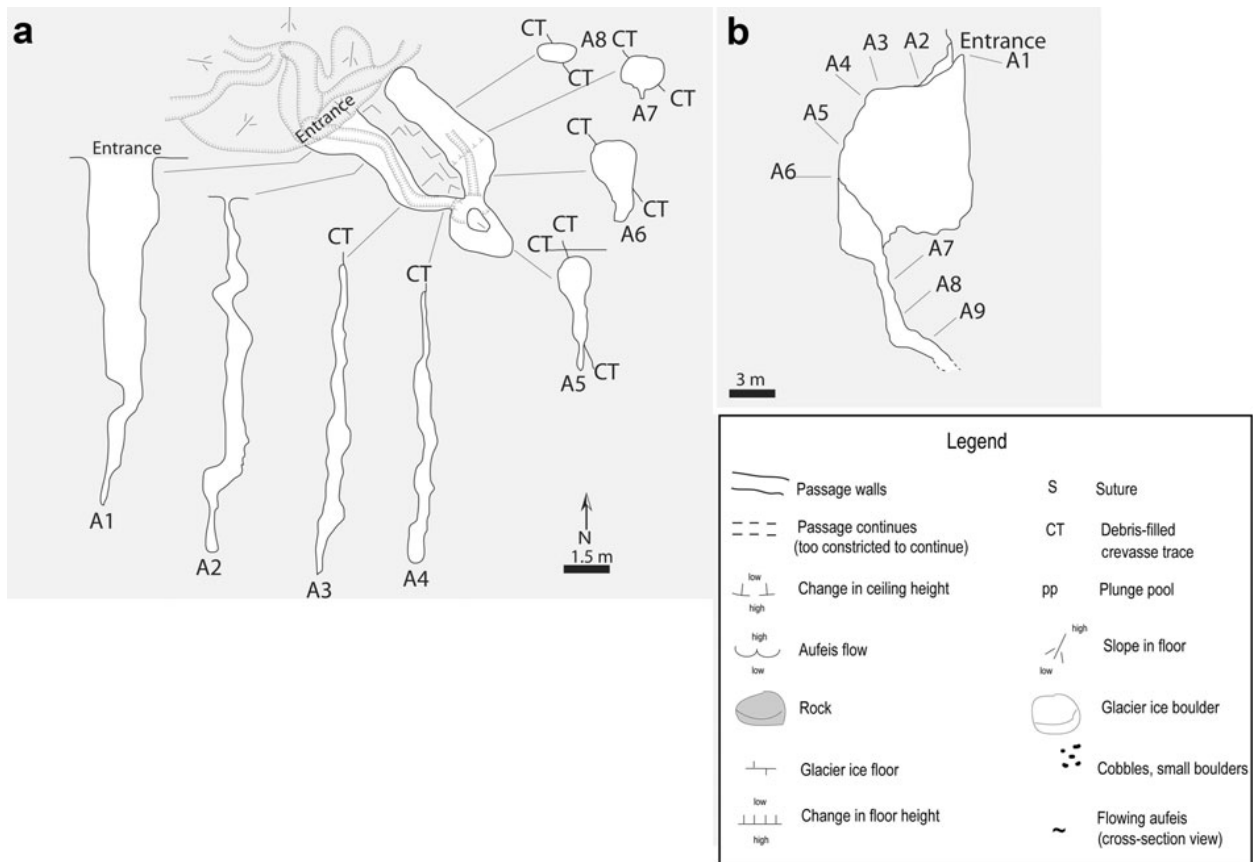


Figure 4-4. Maps of conduit IFC. A) plan view of conduit IFC, B) conduit IFC profile and C) map legend.

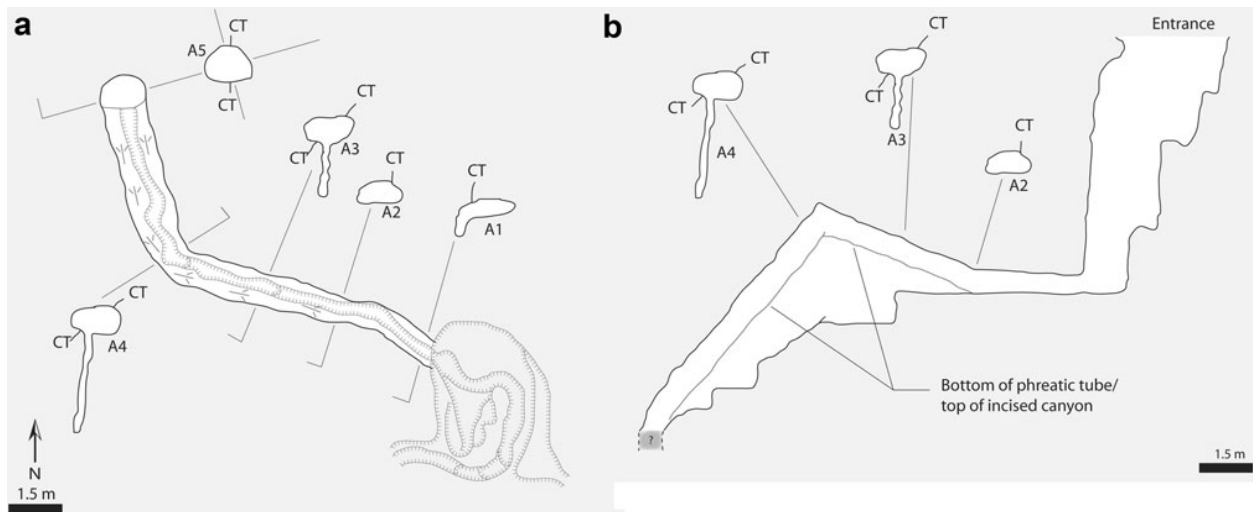


Figure 4-5. Maps of conduit WD. A) plan view of conduit WD and B) profile view of conduit WD.



Figure 4-6. Conduit WD flows uphill in a downstream direction and exhibits a classic 'keyhole' conduit cross-section indicative of a transition from phreatic to vadose flow. The association between the conduit and formative crevasse can be clearly seen in this picture.



Figure 4-7. The crevasse this conduit formed along in is visible in the conduit roof in 2005. By 2006, this conduit roof had melted out and the conduit designated as CP in Figs 1, 2, formed when this conduit incised down to a transverse crevasse.

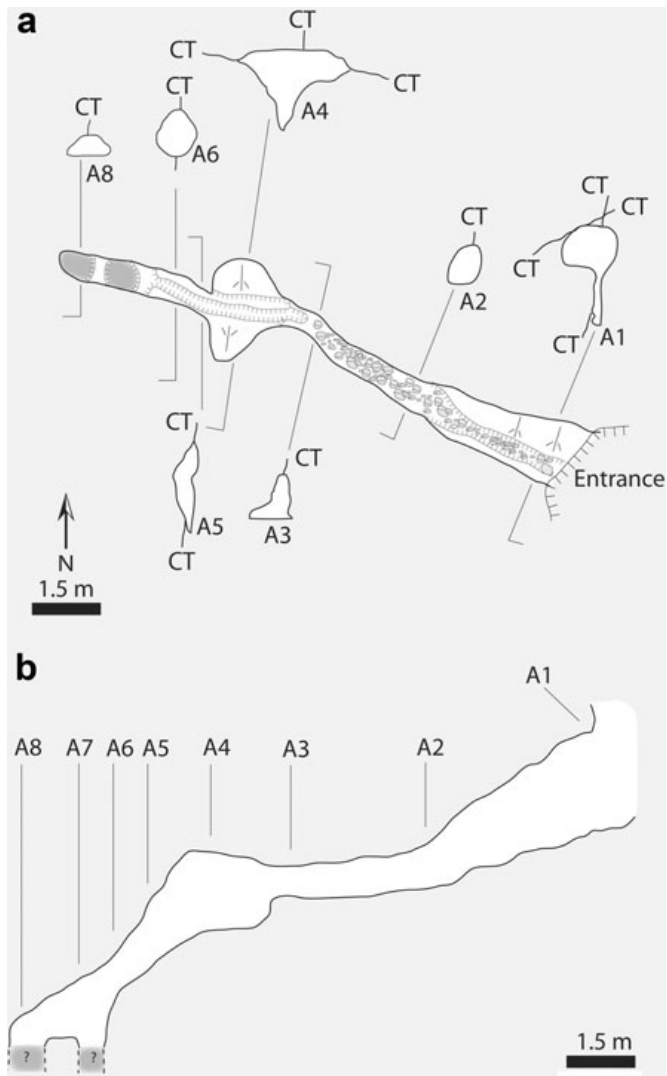


Figure 4-8. Plan and profile views of conduit MS.

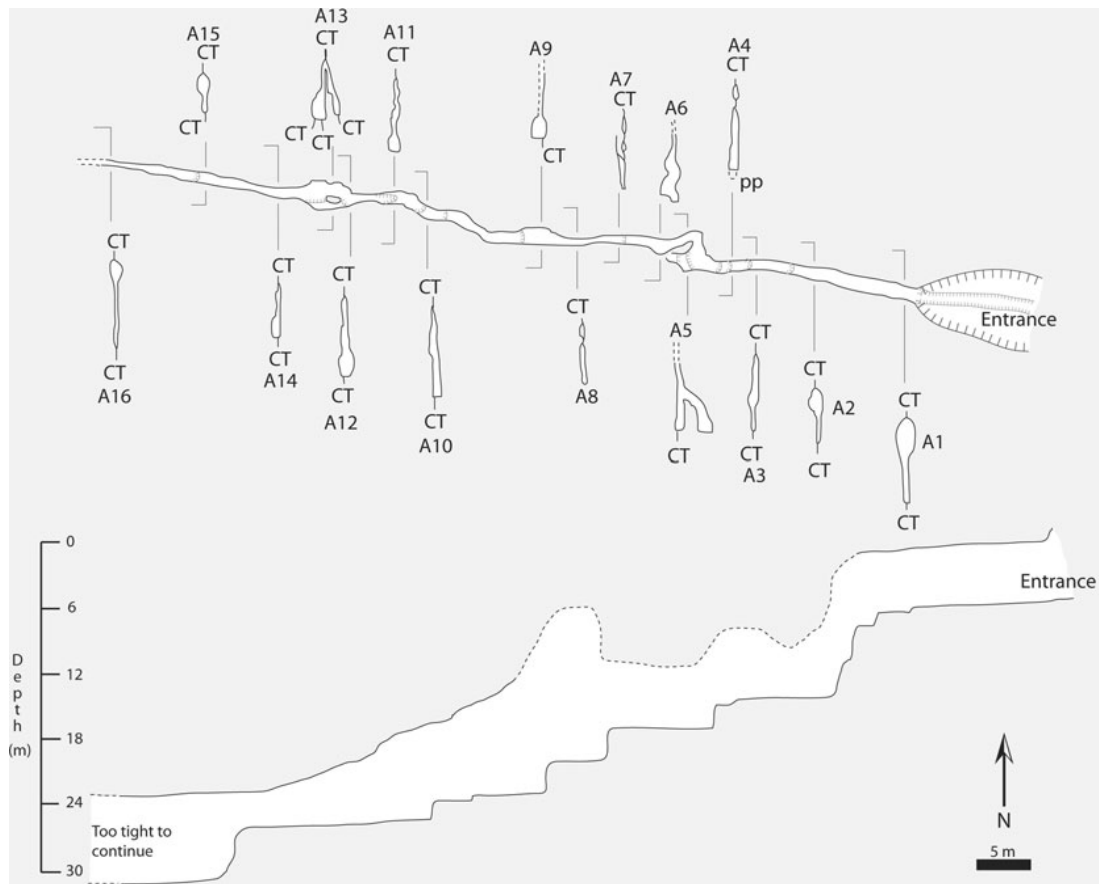


Figure 4-9. Plan and profile views of conduit DP1.

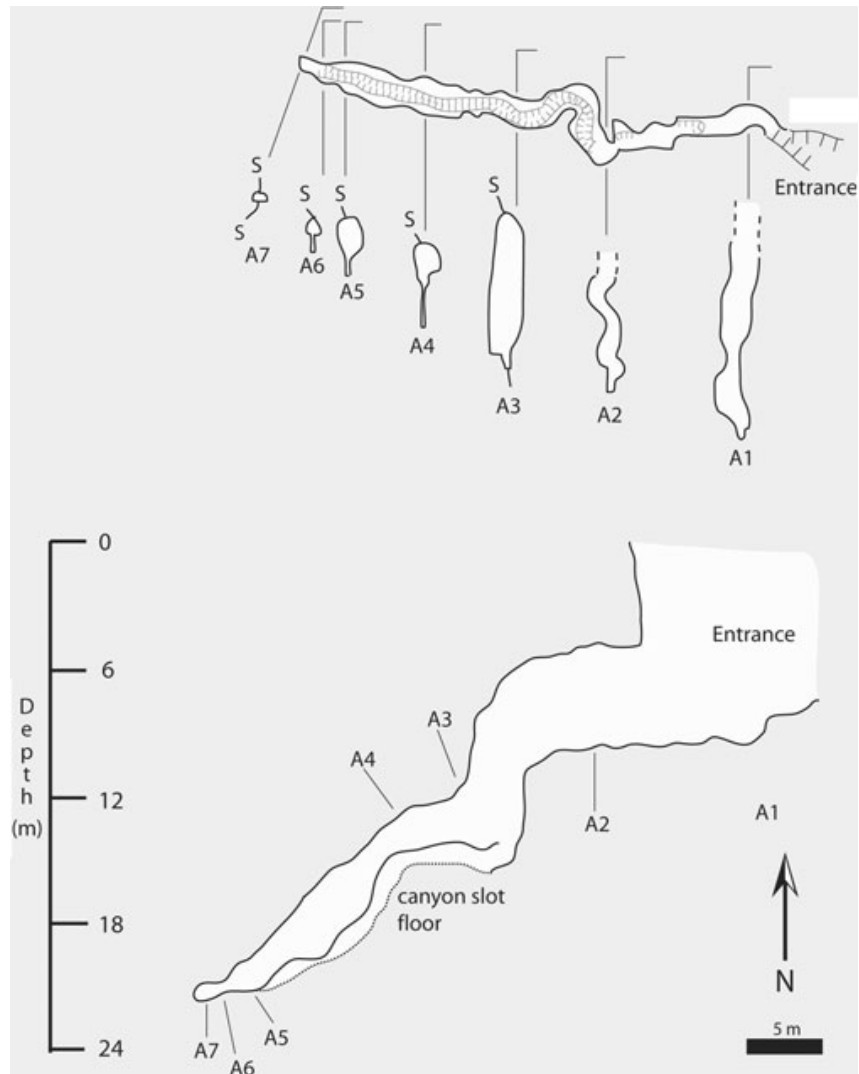


Figure 4-10. – Plan and profile views of conduit DP2.

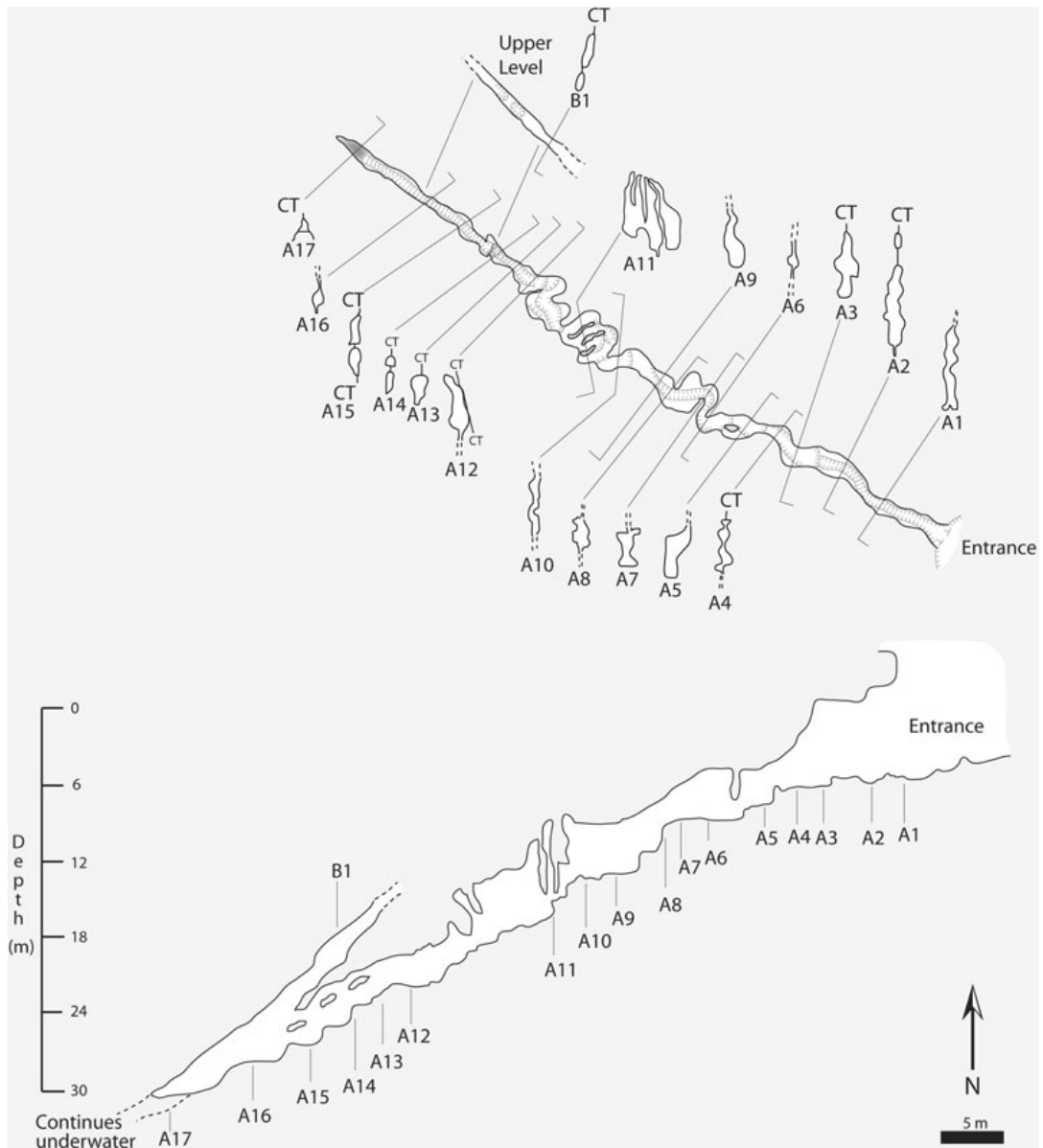


Figure 4-11. Plan and profile views of conduit DP3.

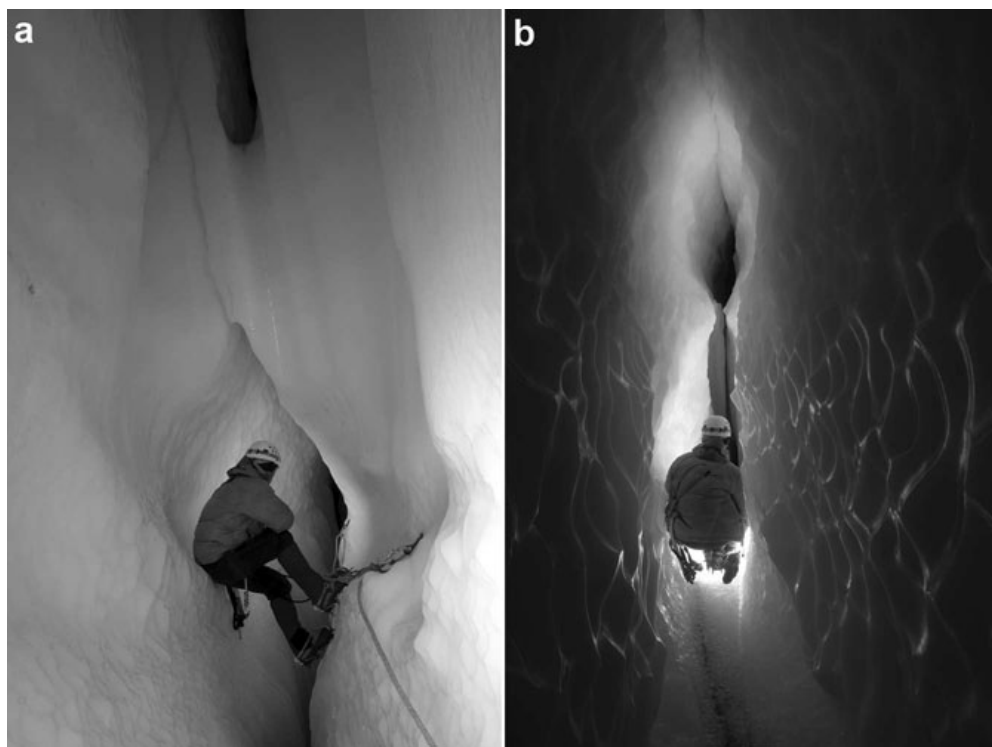


Figure 4-12. Photos of passage morphologies in DP1. A) conduit formation by hydrofracturing switched from one fracture to an adjacent fracture or a splay fracture in this portion of conduit DP1. This photograph corresponds to cross-section A7 in Fig 9 and B) the hydrofracture trace can be seen in floor and ceiling of DP1. This cross-section corresponds to cross-section A16 in Figure 9.

CHAPTER 5 SYNTHESIS AND CONCLUSIONS

Mechanisms of englacial conduit formation for which we have direct observational evidence. These are: (1) the incision and closure of supraglacial streams ('cut-and-closure'); (2) exploitation of permeable structures; and (3) hydrologically assisted fracture propagation. Generalized conduit planforms, profiles and cross-sections for each type of conduit are presented as Figure (5-1).

Glaciers as Deformable Karst Aquifers

The englacial drainage systems we have mapped exhibit many morphological similarities with caves formed in karst aquifers. In developing a comprehensive view of englacial drainage systems, therefore, it is instructive to consider mechanisms of cave formation in soluble rock and to examine the reasons for commonalities and differences. Caves are known to form in many soluble media, including limestone, dolomite, gypsum, chalk, carbonate-cemented sandstone, soil, conglomerate, salt and even quartzite (Klimchouk et al., 2000; Ford and Williams, 2007). It has been demonstrated that, despite differences in enlargement mechanisms (i.e. dissolution, abrasion, etc.), the morphometric similarity of caves in all lithologies stems from the similar hydrogeological requirements for cave inception: a circulating solvent must selectively remove a portion of the matrix rapidly enough to produce conduits before the overburden is removed by denudation (Palmer, 1991; Huntoon, 1995).

Solutional caves form where there is sufficient subsurface water flow to remove dissolved bedrock and keep undersaturated water in contact with the soluble walls. This is possible only where a pre-existing network of integrated openings connects recharge and discharge points (Palmer, 1991). Fractures and bedding planes in rock

provide relatively direct hydraulic pathways that are orders of magnitude more efficient than the tortuous flow paths in a porous medium. Fissures, therefore, act as speleogenetic inception horizons because they have the greatest initial solvent flux (Lowe, 2000) and always win out over matrix porosity in the “race” to initiate a conduit. Even conduits formed in geologically young limestone with high matrix permeability show strong structural controls (Florea, 2006).

Water trajectories in karst conduits are therefore governed by hydraulic gradients aligned along the most conductive zones, rather than the direction of the steepest hydraulic gradient if the overall aquifer is considered as an equivalent porous medium (Ford and Ewers, 1978; Palmer, 1991). The amount of head in these features is determined by differences in elevation and the rate of recharge versus the rate of discharge in the permeable feature. If recharge exceeds discharge, head will increase. If the hydraulic capacity of the system has been reached, head will continue to increase to an elevation equivalent to the elevation of the recharge point while discharge remains constant. Because dissolution is a function of discharge, and flow is concentrated in the most hydraulically conductive zones, a positive feedback loop is established as these zones are preferentially enlarged (White, 1988). Conduit diameter enlargement increases the hydraulic capacity of the system and will draw down head throughout the system for an equivalent recharge. Given sufficient time, large cave systems will develop with ultimate passage morphologies being dictated by a combination of pre-existing geologic structure and the distribution of recharge and discharge points (Palmer, 1991; Ford and Williams, 2007).

In important respects, the formation of englacial conduits follows similar principles, although there are also major differences between speleogenesis in soluble rocks and glacier ice. As is the case for soluble rocks, the formation of a conduit within glacier ice requires a continuous and pre-existing hydraulic connection between areas of different hydraulic potential. This connection must have a high enough hydraulic conductivity to permit sufficient flow to begin the process of passage enlargement, which is by melting in response to viscous heat dissipation in the case of glacier ice. Pathways of high hydraulic conductivity in glacier ice are provided by pre-existing fractures or debris-filled structures lacking ice cement. The flow rate through a fracture in glacier ice is many orders of magnitude greater than through the ice matrix and increases with the cube of the fracture aperture (Figure 5-2). Once flow is established along fractures, a positive feedback between discharge and wall melting leads to the development of a conduit, provided sufficient recharge is available. Conduits therefore evolve from recharge point to discharge point along discrete features of secondary permeability. Because conduits form by head-driven flow in pre-existing pathways connecting recharge and discharge points, neither water flow path nor head are determined directly by ice pressure gradients. Instead, head is determined by the height of the column of water in the permeable feature minus the pressure drop (head loss, $\Delta\phi_f$) resulting from friction between the water and the walls which supports some of the weight of the overlying water:

$$C = P - \Delta\phi_f + \gamma z \quad (5)$$

(Rothlisberger and Lang, 1987). The rate of head loss (or head gradient) depends on the resistance to flow resulting from drag on the fluid which depends on conduit geometry and the amount of turbulence in the flow regime.

Ice pressure does however influence the head in conduits by two other important mechanisms. First, the ice thickness determines the maximum elevation (the elevation and pressure heads) of water in moulins. Second, ice pressure constricts conduit diameters during low-flow periods which reduces the hydraulic capacity of conduits and causes water to back up in the system during later recharge events to increase the total head upstream of the constriction. This is demonstrated with Bernoulli's equation for horizontal flow (6) and the Continuity Equation (7):

$$P_1 + \frac{1}{2}\rho v_1^2 = P_2 + \frac{1}{2}\rho v_2^2 \quad (6)$$

$$A_1 v_1 = A_2 v_2 \quad (7)$$

Where P is the hydraulic pressure in N/m^2 and A is the conduit cross-sectional area. From this relationship we can see that as the conduit cross section decreases, the downstream velocity increases and this increase in velocity is accompanied by an increase in hydraulic pressure on the upstream side of the flow constriction. This creates steep head gradients across the restriction that increase discharge, and therefore ice melting, across the flow restriction. Because of this relationship, conduits that are continuous from recharge point to discharge point should not be able to freeze shut as long as they are being recharged as has been suggested (Rothlisberger and Lang, 1987; Hooke and Pohjola, 1994; Alley et al., 1998). This is because the freeze-on constriction to flow in the downstream reaches will locally steepen the hydraulic gradient, thus increasing the velocity and conduit enlargement rate to keep at least part

of it open. This does not preclude the possibility that conduits can become blocked by squeezing shut during the winter when there is no recharge to keep them open as we discovered to be the case with several cave systems in Nepal and Svalbard (Gulley et al., in press).

Some processes involved in the formation and evolution of englacial conduits have no direct analogue in speleogenesis in rocks. First, near-surface stresses in glaciers commonly exceed the tensile strength of ice, allowing standing water in surface fractures to force its own connection from recharge to discharge point. This is not known to happen in common karst aquifers. Second, the deformation of ice in response to relatively small stress gradients means that conduits can contract or close when their water pressure is less than that of the surrounding ice. As a result, the hydraulic capacity of englacial drainage systems reduces when discharges fall, except where ice overburden and hence cryostatic pressures are small. Because of this, the head in an englacial conduit is extremely sensitive to the relative rates of creep closure and enlargement, and conduits under thick ice will have greater periods of high head than conduits under thinner ice for an equivalent amount of recharge. The efficiency of passage enlargement and closure processes in ice means that the evolution of englacial drainage systems is cyclic, and mature systems can evolve within weeks compared to the millennia required for carbonate caves where dissolution rates are more limited by kinetics.

Where englacial conduits are fed by surface meltwater, variations in recharge during the melt season will typically vary on timescales shorter than those required for passage adjustment by wall melting and ice creep. It cannot be expected, therefore, that

head in conduits will be in equilibrium with the pressure of the surrounding ice. Indeed, it is well known that water levels in moulins can fluctuate several hundred meters over the course of single day (Iken, 1972; Vielli et al., 2004; Badino and Piccini, 2002), resulting in large pressure transients in the vicinity of subglacial conduits (Harbor et al., 1997). This is a direct consequence of the controls on pressure in a conduit open at both ends and experiencing through-flow; total head is determined by the column of water upstream minus head loss due to friction along the pipe's length. Total head will increase when recharge exceeds discharge and water backs up the system and, conversely, falling water pressures will result when discharge exceeds recharge. Ice pressure therefore only controls water pressure in conduits by reducing conduit diameter (and hence hydraulic capacity).

The requirements for formation of the range of conduit types outlined in Section 4 allows a predictive model of englacial drainage, and by extension the location of discrete subglacial recharge, to be developed, based on patterns of supraglacial water supply and glacier structure and dynamics. The expected locations of generic englacial conduit systems in an idealized polythermal glacier are shown in Figure 5-3. The same types of conduits will form in the same locations by the same processes in temperate glaciers, with the exception of cut-and-closure conduits which are not known to form in temperate glaciers.

The simplest drainage systems occur on cold-based, polythermal glaciers, including small Svalbard glaciers such as Longyearbreen and Scott Turnerbreen. The lack of basal motion below such glaciers means that glacier surface velocities – and velocity gradients – are small, so that surface crevassing is limited. In consequence,

drainage of surface meltwater occurs entirely by supraglacial streams and, where incision rates outpace surface melt rates, cut-and-closure englacial conduits. Cut-and-closure conduits are particularly common along the lateral flanks of small cold-based glaciers in Svalbard, although they can occur on any part of the glacier wherever meltwater from large catchments is focused.

Where warm-based conditions permit basal motion, glacier velocity gradients, both in the cross-flow and along-flow directions, can be high enough to initiate fracturing of near-surface ice. Crevasse fields along the lateral margins and the main body of the glacier can therefore intercept surface drainage and route it to the bed at the margins (Stenborg, 1969, 1973). Where ice is thin, such as close to glacier margins, 'dry' surface crevasses may reach the bed, so that surface-to-bed drainage may be possible without the need for the high recharge rates required for hydrologically assisted fracture propagation. Where ice is thicker, hydrologically assisted fracturing appears to be the only mechanism capable of routing water to the bed rapidly and consistently. The joint requirements of stressed ice and high recharge rates mean that locations of deep moulins, should be relatively constant in both space and time, and potentially predictable. Surface-to-bed drainage is particularly likely at glacier confluences because flow patterns encourage longitudinal extension and transverse compression providing crevasses, and lateral meltstreams converge to provide a large meltwater source (Gudmundsson and Bauder, 1999; Benn et al., in press). Surface-to-bed drainage will also be favoured in areas of extending flow and high meltwater production rates (e.g. Catania et al., 2008).

The englacial drainage systems of Himalayan debris-covered glaciers also reflect patterns of water supply, structure and glacier dynamics. The lower ablation zones of such glaciers are stagnant, with low surface gradients and abundant ponded water. In contrast, the upper ablation zones have surface slopes of a few degrees and have typical surface velocities of a few tens of metres per year (Quincey et al., in press). Drainage of the upper ablation zones is dominated by supraglacial channels and cut-and-closure conduits, whereas on the lower ablation zones intermittent relocation of ponded meltwater occurs when supraglacial ponds drain through structurally controlled conduits (Gulley and Benn, 2007; Gulley et al., in press). In the intervening zone, water can penetrate into the glacier by hydrologically assisted fracturing, which is facilitated by compressive stresses generated by decelerating ice.

The co-existence of different types of englacial drainage system on a single glacier means that there may be both 'quickflow' and 'slowflow' components of englacial drainage. A possible example of this is at Storglaciären, Sweden, where Fountain et al. (2005) observed numerous water-filled fractures in the walls of boreholes. Discharge hydrographs from this glacier, however, indicate that efficient surface-to-bed drainage occurs in a high capacity conduit system (e.g. Seaberg et al., 1988; Hooke et al., 1988; Jansson, 1996). This suggests that slowflow in an interconnected fracture network is dominant form of englacial drainage in terms of spatial extent, whereas water draining via moulins is dominant in terms of the volume of water reaching the bed. Large englacial conduits are few and widely spaced, and are unlikely to be intercepted by boreholes, especially as conduit entrances are most likely to be at the edges of crevasse fields which make poor drill sites.

Connection to the bed by hydrologically-assisted fracturing delivers high hydraulic heads to discrete portions of the glacier bed. Because the locations of moulins are determined by ice dynamics and surface conditions and not gradients in ice pressure, subglacial recharge points can potentially be located in overdeepenings.

While there is likely a continuum of englacial conduit sizes, many of which are not large enough to permit direct human exploration, these conduits will have a much more limited influence on the rate and magnitude of subglacial recharge due to their substantially smaller discharges. For instance, in laminar flow, the discharge of a tube increases with the fourth power of the conduit diameter (White, 1988; Palmer, 2008). If a conduit diameter is increased by a factor of ten, the discharge increases by 10,000. We acknowledge that flow in conduits with diameters greater than 1 cm is likely to be turbulent and this approximation is an upper boundary condition.

Applied to the Greenland Ice Sheet, our model predicts that supraglacial meltwater would only be able to reach the bed by hydrologically assisted fracturing. This constrains the locations of discrete subglacial recharge to zones of accelerating ice flow and supports the hypothesis that the acceleration of inland ice is driven by longitudinal coupling (Price et al., 2008).

While we have found no evidence for a Shreve type drainage system in our study of englacial conduits, Shreve's model has been used to successfully locate a subglacial conduit on the Haut Glacier D'Arolla (Sharp et al., 1993). Other applications of Shreve theory for finding subglacial conduits have not been as successful. The subglacial conduit system on the Glacier D'Argentierre famously changed locations after an expensive hydroelectric scheme had been constructed (Hantz and Lliboutry, 1983). It

could be that Shreve theory sometimes predicts the location of subglacial conduits, but perhaps for the wrong reasons.

Summary and Conclusions

The characteristics of englacial drainage systems, as determined by direct exploration, down-borehole imaging and geophysical surveys, are inconsistent with conduit evolution from intergranular veins as proposed by Shreve (1972). Flow paths are strikingly at variance with hydraulic gradients that would be calculated from conduit elevation and ice overburden pressure. Other fundamental assumptions of the model, i.e. that ice is homogeneous and permeable, and discharge is constant, are also inconsistent with observations. Therefore the Shreve model is probably an inappropriate framework for understanding englacial drainage.

Englacial drainage systems are inferred to form by three main mechanisms: (1) cut-and-closure, or the incision of supraglacial streams followed by closure of the upper levels by snow or ice, and ice creep; (2) exploitation of pre-existing secondary permeability, in the form of debris-filled crevasse traces or open fractures; and (3) hydrologically assisted fracture propagation and hydrofracturing.

Englacial drainage systems formed by these different mechanisms exhibit systematic relationships with glacial thermal regime and local stress conditions. Cut-and-closure conduits and drainages developed along relict debris-filled structures form where glacial stresses are insufficient to initiate or maintain open fractures. Cut-and-closure conduits are common on uncrevassed regions of polythermal glaciers, and in Svalbard are most frequently encountered along the flanks of small polythermal glaciers, where surface meltwater is focused by surface topography. Cut and closure

conduits are also found to occur on relatively steeply sloping, uncrevassed surfaces of debris-covered glaciers.

Cut-and-closure conduits require a large water supply to drive incision, but require surface ablation rates to be low in order for stream incision to outpace surface lowering. Surface ablation rates are kept low on debris-covered glaciers by the insulating effects of debris-cover and on high latitude glaciers by low energy received from the atmosphere. In both cases, large supraglacial catchments are only possible where the ice is unfractured. Cut-and-closure conduits are therefore likely to be restricted to polythermal glaciers at high latitudes or altitudes, or debris-covered glaciers with relatively high surface slopes with low surface strain rates.

Conduits that exploit debris bands or debris-filled crevasse traces are generally restricted to the stagnant, low-surface-gradient tongues of debris-covered glaciers. In order for debris-bands to be permeable they need to be free of interstitial ice cements, which generally restricts the origin of the debris-features to high altitude ice falls, where melting and refreezing is limited. These permeable features are later exploited as lake basins expand and coalesce in the stagnant debris-covered glacier tongues.

We have found no evidence that supraglacial meltwater draining into this group of conduits has any influence on glacier dynamics. This is because the conduits either do not connect to the glacier bed, as is the case with some cut and closure conduits and all observed debris-filled trace conduits, or because their subglacial continuations form very narrow corridors flanked by cold ice. It is possible, however, that in some circumstances cut-and-closure conduits might establish a connection with a warm

glacier bed. We believe this possibility is small, because of the tendency of cut-and-closure conduits to become blocked when overlain by thick ice (Gulley et al., in press).

Conduits developed along pre-existing fractures or by hydrologically assisted fracture propagation occur on more dynamically active glaciers, where surface strain rates are sufficiently high to initiate and maintain fractures. The locations of moulins are determined by a combination of glacier dynamics and meltwater supply. Large moulins will form wherever sufficiently stressed ice coincides with a sufficient water supply, such as shear zones close to glacier margins, the upglacier end of areas of extending flow, and glacier confluences.

Pre-existing fractures can route water to glacier beds where ice is thin. Where ice is thick, surface-to-bed drainage is probably only possible by hydrologically assisted fracturing. Because the location of deep fractures is a function of glacier dynamics and surface melting, they should reform in similar places each melt season, and are potentially predictable.

Englacial conduits form from recharge point to discharge point, and with the exception of cut-closure conduits, form by head-driven flow in fractures and pipes. Gradients in ice pressure do not appear to influence the spatial distribution of englacial conduits because water pressure in conduits is decoupled from ice pressure. Ice pressure is only likely to influence total head by constricting conduit diameters during low-flow periods which causes water to back up in the system during later recharge events.

Moulins deliver water to glacier beds at discrete points, which may be irregularly spaced, with varying density and discharge. This is likely to result in subglacial head

distributions that are different than those modelled using Shreve theory and will be discussed in a separate paper focusing on subglacial drainage. Recharge points can be located anywhere consistent with the glacier stress field and surface water availability, including overdeepenings. Consequently, there is no reason to expect that water will bypass overdeepenings by englacial or sub-marginal routes.

Because patterns of recharge are a function of glacier dynamics, and dynamics are affected by water at the bed, an improved understanding of englacial drainage systems may have important implications for modelling coupled hydrology and ice flow.

Principle Stresses At entrance	Form by	Conduit Planiform	Conduit Profile	Diagnostic Cross-section	Glacier Crevasse Pattern
	Hydrofracture				
	Hydrofracture				
	Existing Secondary Permeability				
	Existing Secondary Permeability				
	No Secondary Permeability				

Figure 5-1. Planforms, profiles and cross-sections of conduits characteristic of each conduit formation mechanism are grouped with the glacier stress conditions in which they form. From left to right: A) extensional hydrofractures, B) compressive hydrofractures, C) exploitation of shear crevasses (in this case where a shear crevasse intersects a thrust fault), D) exploitation of permeable debris-filled crevasse traces and E) cut-and-closure. CT = crevasse trace. DCT = Debris-filled crevasse trace.

Hydraulic conductivity as a function of fracture diameter

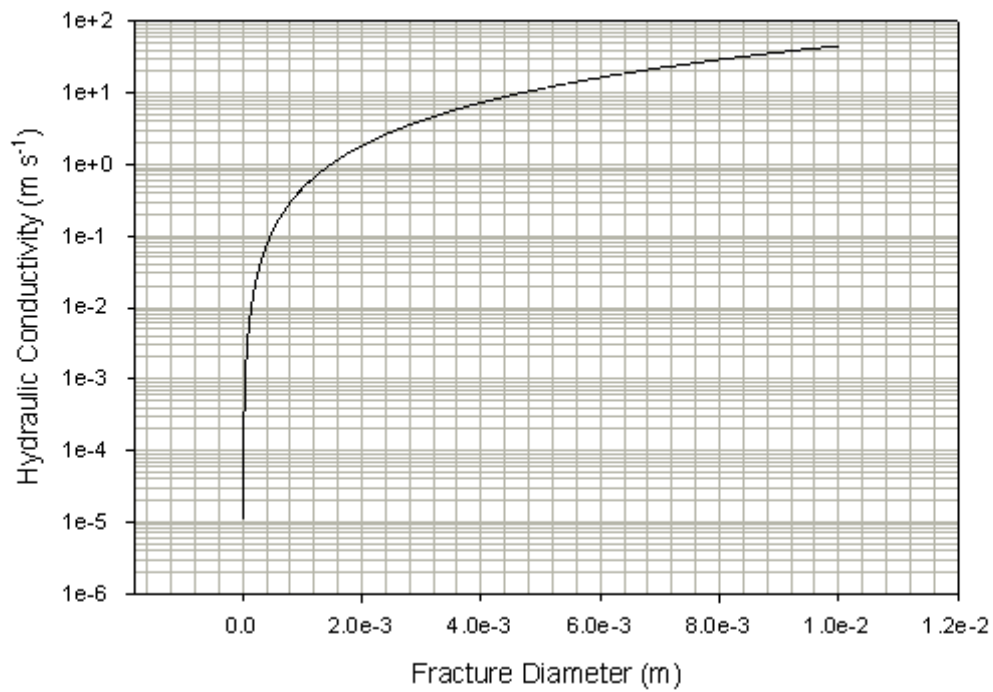


Figure 5-2. Hydraulic conductivity of fractures plotted as a function of fracture aperture, calculated using the cubic law for flow between parallel plates. Calculations assume a water temperature of 0 degrees Celsius.

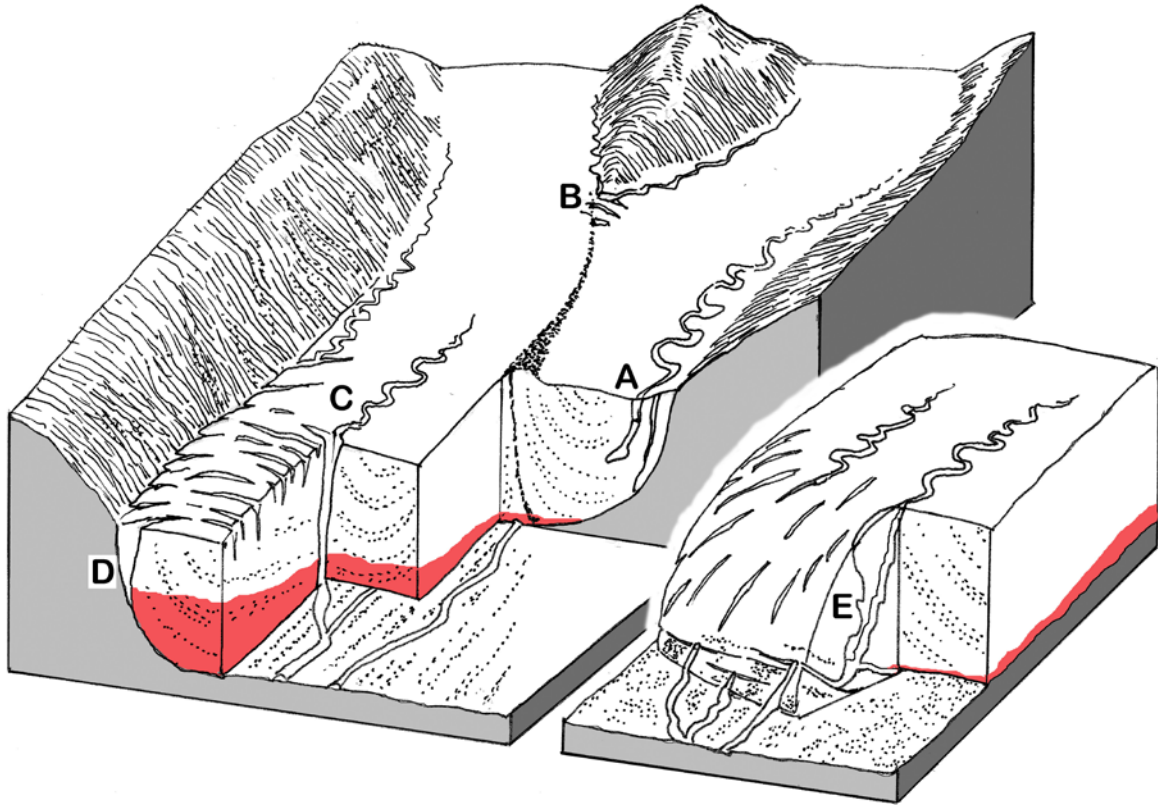


Figure 5-3. The locations of generic englacial conduits are shown in an idealized polythermal glacier. The shaded area near the bed denotes warm ice. The same types of conduits will form in the same locations by the same processes in temperate glaciers, with the exception of cut-and-closure conduits which are not known to form in temperate glaciers. A) Cut and closure conduits will form where supraglacial streams cross uncrevassed regions of polythermal glaciers. B) Extensional hydrofractures are commonly found at glacier confluences where extensional crevasses encouraged by ice convergence and acceleration form proximal to lakes that form in the topographic hollows between glaciers. These lakes are frequently fed by ice-marginal cut-and-closure conduits that are graded to the lakes. C) Extensional hydrofractures frequently form where large supraglacial streams formed on uncrevassed ice flow onto ice that is accelerating. D) Shear crevasses route englacial drainage to the bed at the margin where ice at the pressure melting point. E) Compressive hydrofracturing occurs where supraglacial streams or lakes overdeepen longitudinal crevasses, potentially reaching the bed. Thrust faults in compressive glacier tongues (which can form where polythermal glacier tongues are frozen to their bed or where temperate glacial tongues impact an ice-cored moraine) can bridge cold ice to route water from temperate glacier beds to the surface.

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BIOGRAPHICAL SKETCH

Like many graduate students, Jason Gulley was born. Soon thereafter, he attended preschool and learned that: paste could be tasty, Cheerios could be strung together with yarn to make necklaces and that if he didn't say the Pledge of Allegiance the school principal would call his mother and his mother would come to school to enthusiastically administer corporal punishment. His elementary school was co-located with a church and as such was decorated with pictures of some dude that died roughly 2,000 years ago and who, despite being from the Middle East, was rather strangely depicted as a bearded white guy. Jason generally loathed his time at William Henry Harrison High Schools and maintains that his low GPA reflects a much richer education gained from the 'other books' he preferred to the government issued ones. Jason dropped out of more colleges and universities than most high school students visit during their senior year, including (but not limited to): CBS, UC (main and Raymond Walters), Purdue and the University of New England (where he briefly majored in Aboriginal studies). He finally graduated from the prestigious Eastern Kentucky University with a BA in geology in August, 2006. Jason enrolled in a PhD program at UF in January, 2007, but spent most of his time crawling around holes in limestone and ice or underwater.