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POLYTHERMAL GLACIER HYDROLOGY: A REVIEW

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[1] The manner by which meltwater drains through a glacier is critical to ice dynamics, runoff characteristics, and water quality. However, much of the contemporary knowledge relating to glacier hydrology has been based upon, and conditioned by, understanding gleaned from temperate valley glaciers. Globally, a significant proportion of glaciers and ice sheets exhibit nontemperate thermal regimes. The recent, growing concern over the future response of polar glaciers and ice sheets to forecasts of a warming climate and lengthening summer melt season necessitates recognition of the hydrological processes in these nontemperate ice masses. It is therefore timely to present an accessible review of the scientific progress in glacial hydrology where nontemperate conditions are dominant. This review provides an appraisal of the glaciological literature from nontemperate glaciers, examining supraglacial, englacial, and subglacial environments in sequence and their role in hydrological processes within glacierized catchments. In particular, the variability and complexity in glacier thermal regimes are discussed, illustrating how a unified model of drainage architecture is likely to remain elusive due to structural controls on the presence of water. Cold ice near glacier surfaces may reduce meltwater flux into the glacier interior, but observations suggest that the transient thermal layer of near surface ice holds a hydrological role as a depth-limited aquifer. Englacial flowpaths may arise from the deep incision of supraglacial streams or the propagation of hydrofractures, forms which are readily able to handle varied meltwater discharge or act as locations for water storage, and result in spatially discrete delivery of water to the subglacial environment. The influence of such drainage routes on seasonal meltwater release is explored, with reference to summer season upwellings and winter icing formation. Moreover, clear analogies emerge between nontemperate valley glacier and ice sheet hydrology, the discussion of which indicates how persistent reassessment of our conceptualization of glacier drainage systems is required. There is a clear emphasis that continued, integrated endeavors focused on process glaciology at nontemperate glaciers are a scientific imperative to augmenting the existing body of research centered on ice mass hydrology.

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1. INTRODUCTION

[2] Glaciers represent environmental freshwater assets, and much contemporary glaciological research has focused attention upon the polar regions, where climate warming is projected to be greatest [*Cattle and Crossley*, 1995; *Christensen et al.*, 2007; *Kattsov et al.*, 2005] and where ice mass losses are most significant for global water resources and contemporary and predicted sea level change [*Abdalati*, 2006; *Alley et al.*, 2005a; *Allison et al.*, 2009; *Cazenave and Llovel*, 2010; *Radić and Hock*, 2011; *Raper and Braithwaite*, 2005; *Rignot et al.*, 2011]. With the consensus of forecasts reported by the Intergovernmental Panel on Climate Change (IPCC) and Arctic Climate Impact Assessment (ACIA) [*Christensen et al.*, 2007; *Kattsov et al.*, 2005] suggesting longer summer seasons in polar latitudes, the potential for increased volumes of glacial runoff is clear, as is broadly expressed in increasingly negative *surface mass balance* trends and declining *accumulation area ratios* in the Arctic over the past 50 years [e.g., *Dyurgerov and Meier*, 2005; *Hanna et al.*, 2008] and the *dynamic thinning* of the Greenland and Antarctic ice sheet margins in recent decades [e.g., *Pritchard et al.*, 2009; *Rignot and Thomas*, 2002]. (For the definitions of key glaciological terms that appear italicized

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throughout this review, readers are directed to the glossary included after the main text and to *Fountain and Walder* [1998].)

[3] Polar ice masses, rather than being impermeable, are interlaced with drainage systems, comprising surface (*supraglacial*), internal (*englacial*), and basal (*subglacial*) hydraulic components. The role and activity of each of these hydraulic subsystems is variable both in space and time, but their coupling has profound consequences for ice dynamics: Research conducted upon Arctic valley glaciers has illustrated that glacier hydrology and ice motion are intricately linked as a result of the role of meltwater accessing the glacier sole and reducing basal traction [e.g., *Andreason*, 1985; *Bingham et al.*, 2006; *Copland et al.*, 2003; *Jansson*, 1995; *Müller and Iken*, 1973; *Rippin et al.*, 2005b]. Such a linkage has been suggested to drive *glacier surge cycles* in higher latitudes [e.g., *Björnsson*, 1998; *Harrison et al.*, 1994; *Kamb*, 1987; *Lingle and Fatland*, 2003; *Merrand and Hallet*, 1996; *Murray et al.*, 2000] and more recently has been invoked to explain dynamics at ice sheet margins [*Bartholomew et al.*, 2010; *Palmer et al.*, 2011; *Pritchard et al.*, 2005; *Shepherd et al.*, 2009; *Sundal et al.*, 2011; *van de Wal et al.*, 2008; *Zwally et al.*, 2002]. The coupling between polar glacier hydrology and dynamics has the potential to accelerate ice losses via a positive feedback whereby the rate of mass transfer to lower elevations is increased. Moreover, a glacier's hydrological architecture affects the timing and volume of meltwater release [*Haeberli*, 1983; *Jansson et al.*, 2003] and the quality of waters delivered to downstream terrestrial and/or aquatic environments [*Alley et al.*, 1997; *Brown*, 2002]. Yet despite this comprehension of key issues, considerable uncertainty surrounds the nature of meltwater propagation through glaciers [*Fountain and Walder*, 1998; *Lemke et al.*, 2007], which is one of a number of ambiguities surrounding glacial processes. Consequently, the uncertainty in our understanding of the internal drainage structures and hydraulic properties of differing glaciers is important and compromises our ability to model future polar ice mass response to climate forcing. Thus, the importance of fostering a better understanding of glacier hydrology in the above contexts has never been clearer than in the present era of research motivated by climate change.

[4] In the past, researchers have sought to simplify the range of thermal conditions that can occur in the glacial cryosphere, proposing that *temperate glaciers* lie at one end of a "spectrum" with *cold glaciers* at the other [*Boulton*, 1972; *Hodson*, 1994], with the remainder of the nontemperate thermal regimes, notably, characteristic of polar environments, encompassed by the term *polythermal glaciers* [*Blatter and Hutter*, 1991; *Fowler*, 1984]. Critically, the hydrological configuration of a glacier is significantly influenced by its thermal structure, and yet the differences between the hydrology of temperate and nontemperate glaciers has remained largely underappreciated. This is surprising given that the majority of Earth's grounded ice is found in the polar regions, exhibiting nontemperate thermal structures, and with emerging research problems including, first, the ability of surface meltwater to penetrate to the bed of the Greenland Ice Sheet [*Catania et al.*, 2008; *Das et al.*, 2008]; second, the potential impact of glacial runoff along the Antarctic Peninsula upon the marine ecosystem [*Dierssen et al.*, 2002] and regional glacier extents [*Cook et al.*, 2005]; and last, the establishment of episodic hydrological connections between subglacial lakes beneath continental Antarctica [e.g., *Fricker et al.*, 2007; *Gray et al.*, 2005; *Wingham et al.*, 2006]. The complications that persist in advancing the conceptualization of glacier hydrology include (1) a literature that has been dominated by models and reviews developed for temperate alpine glaciers [e.g., *Fountain and Walder*, 1998; *Hooke*, 1989; *Hubbard and Nienow*, 1997; *Richards et al.*, 1996; *Röthlisberger and Lang*, 1987], (2) the longevity of the notion that cold ice inhibits water flow through glaciers [*Paterson*, 1994], a perspective that recent work is beginning to dispel, and (3) the diversity and complexity of polythermal glacier thermal structure, which through geophysical techniques have become increasingly recognized.

[5] To date, only *Hodgkins* [1997] has provided a summary of hydrological observations from nontemperate glaciers, although his review was restricted to research conducted in Svalbard. However, a much wider literature base exists thanks to the occurrence of nontemperate glaciers in other regions, and has been updated by more recent investigations. Here we provide a synthesis of these studies to better inform the growing research attention that is being focused upon nontemperate ice masses in polar regions. This review aims to outline both recent and arguably underreported, pioneering developments in the field of polythermal glacier hydrology across the Arctic. The diversity of thermal regimes is discussed in detail, providing a context for the subsequent sections that sequentially explore temperate ice hydrology and the supraglacial and englacial components of a nontemperate glacier's drainage system; we also highlight observations of subglacial, seasonal, and proglacial hydrology. Throughout, we place emphasis upon polar valley glaciers, but for completeness include a short section to consider the parallels between the existing knowledge of nontemperate glacier and ice sheet hydrologies; we note that the latter has yet to be subjected to the integrated studies of hydrological processes described here.

[6] Thorough reviews exist elsewhere for the hydrology of seasonal snow and *firn* [e.g., *Fountain*, 1996; *Male and Gray*, 1981; *Schneider*, 2000; *Singh and Singh*, 2001] and temperate glaciers [*Fountain and Walder*, 1998; *Hooke*, 1989; *Röthlisberger and Lang*, 1987]; therefore these topics are only touched upon where appropriate. Similarly, rather than presenting a comprehensive review on subglacial hydrology, an extensive topic is its own right, we tend toward observations gleaned at nontemperate valley glaciers, but for more detailed consideration of subglacial processes readers are directed to more general reviews such as *Hubbard and Nienow* [1997] or the more detailed overview presented by *Clarke* [2005].

2. COMPLEX THERMAL REGIMES AND NOMENCLATURE

[7] The thermal structure of a glacier is ultimately dependent on its heat balance, which is controlled by intrinsic and

extrinsic factors affecting processes of conduction, advection, and latent heat transfer [*Paterson*, 1994]. These processes occur on a variety of time scales, but as the thermal response time for large, nontemperate ice masses is very long compared to the rate at which many environmental (extrinsic) factors change, the thermal structure of a polar ice mass has a significant memory that is largely inherited. Consequently, rapid heat exchanges across a glacier's key interfaces (i.e., surface to atmosphere or substrate to base) are important, but so too is the current and historical (paleo) burial of thermal conditions and heat within accumulating and metamorphosing snow and firn. The combination of influences upon a glacier's thermal regime results in a marked diversity of temperature structures, resulting in a similar variation in the hydraulic properties of ice masses. Therefore, before considering hydrology in detail, it is necessary to appreciate the array of glacial thermal structures that dictate important constraints on water drainage processes.

[8] Polythermal glaciers are defined as ice masses displaying a perennial concurrence of *temperate* and *cold* ice [*Blatter and Hutter*, 1991; *Fowler and Larson*, 1978]. Temperate ice is at the *pressure melting point* (PMP) exhibiting coexisting water and ice and an *interstitial* liquid content up to ~9% by volume [*Pettersson et al.*, 2004]. Contrastingly, cold ice is below PMP and exhibits negligible interstitial water. Where cold ice characterizes the glacier-bed interface, the high adhesive strength of cold ice ensures the glacier is effectively frozen to the rock substrate. Polythermal glacier regimes are a common, characteristic response in regions with protracted subfreezing winter air temperatures where steep, annual near-surface temperature gradients result in the net conduction of heat away from the glacier ice and a deeper penetration of the winter cold wave [e.g., *Björnsson et al.*, 1996; *Blatter and Hutter*, 1991]. Detailed mathematical approaches to modeling the temperature distribution within nontemperate ice masses are given by *Hutter* [1982] and *Aschwanden and Blatter* [2009].

[9] The complexity of thermal regimes hidden within the umbrella term "polythermal" was recognized by *Hutter et al.* [1988] and *Blatter and Hutter* [1991]. Although the diminishing interstitial water content transition between cold and temperate glacier ice is gradual [*Blatter*, 1987; *Moorman and Michel*, 2000a], is difficult to distinguish [cf. *Björnsson et al.*, 1996; *Brown et al.*, 2009], or may vary in space due to strain heating [*Hutter*, 1982] or over time [*Irvine-Fynn et al.*, 2006; *Jania et al.*, 2005], geophysical surveys on nontemperate glaciers have revealed variation in the location of a *cold-temperate ice transition zone* (CTZ), which has facilitated understanding of the complex controls on the spatial distribution of temperate and cold ice and recognition of the wide diversity of nontemperate regimes (Figure 1 and references therein).

[10] Where ice thickness is in excess of the cold wave penetration depth or where ice overburden pressure elevates ice temperature to the PMP, a temperate ice zone may be sustained [*Bamber*, 1989; *Hagen and Sætrang*, 1991], leading researchers to suggest threshold glacier widths >430 m [*Williams*, 1970] or ice thicknesses >80 m [*Hagen et al.*, 1993; *Murray et al.*, 2000] for a seemingly common, "two-layered" polythermal regime [*Macheret*, 1990] (Figure 1e). A similar thermal structure exhibiting a basal zone of temperate ice (Figures 1e–1g) can also result from geothermal heat flux and/or heat released by internal ice deformation or strain heating [*Aschwanden and Blatter*, 2005; *Björnsson et al.*, 1996; *Blatter*, 1987; *Blatter and Hutter*, 1991; *Fahnestock et al.*, 2001; *Fowler*, 1984; *Paterson*, 1994]. Clearly, ice flow is a critical parameter in determining thermal regime. As suggested by *Hooke et al.* [1983] and *Blatter and Hutter* [1991], the advection of temperate ice from up-glacier locations may result in warm basal ice extending down-glacier (Figures 1g–1j). *Glacier hypsometry* and geometry, as contributing drivers of ice flow, may be important to the spatial pattern of down-glacier advection of temperate ice, particularly where the accumulation area ratio is high [e.g., *Etzelmüller et al.*, 1993]. The process of temperate ice advection is further complicated by the nature of the accumulation regime and leads to a characteristic contrast between the nontemperate glaciers of the continental Canadian Arctic and those located in maritime Svalbard and Scandinavia: Differences in typical winter temperatures and snow depths result in regional disparity in insulation and latent heat release during snow metamorphosis and burial (compare Figures 1c, 1e, and 1h). Indeed, the transfer of cold ice from accumulation areas can further complicate a glacier's thermal structure, even for glaciers outside the polar regions [e.g., *Eisen et al.*, 2009; *Suter et al.*, 2001], while the thermal properties of water-filled crevasses and conduits may promote anomalous temperate ice zones [e.g., *Jarvis and Clarke*, 1974] (Figure 1d). Such variability leads to an uneven distribution of temperate ice and water content within, first, the accumulation area [e.g., *Björnsson et al.*, 1996; *Jania et al.*, 1996] and, second, the ice advected down-glacier [*Pettersson et al.*, 2004]. While the schematics presented in Figure 1 present idealized glacier centerline representations, it is important to recognize that real ice masses will exhibit a spatial variability in their thermal structure: Temperate ice areas within nontemperate glaciers may be patchy [*Murray et al.*, 2000].

[11] Complexity in the thermal regime of nontemperate glaciers is also evident through time on account of its thermal response time and sensitivity to climatic variability. A predominantly cold contemporary thermal regime (Figure 1f) imposed by longer-term climatic history has been shown to exist at Laika Glacier and John Evans Glacier, Arctic Canada [*Blatter and Hutter*, 1991; *Wohlleben et al.*, 2009], and such findings attest that many nontemperate glaciers are, thermally, in disequilibrium with contemporary climates. The thinning of numerous nontemperate glaciers at high latitudes results in an increase in the extent and/or proportion of cold ice or, in some instances, completely changes the thermal regime [*Blatter and Hutter*, 1991; *Hodgkins et al.*, 1999] (Figure 2). Recently, such changes have been exemplified by Austre Brøggerbreen, Svalbard: Sustained recent retreat coupled with progressive thinning has changed the glacier to a cold-based regime [cf. *Björnsson et al.*, 1996; *Glasser and Hambrey*, 2001; *Hagen and Sætrang*, 1991; *Hodson et al.*,

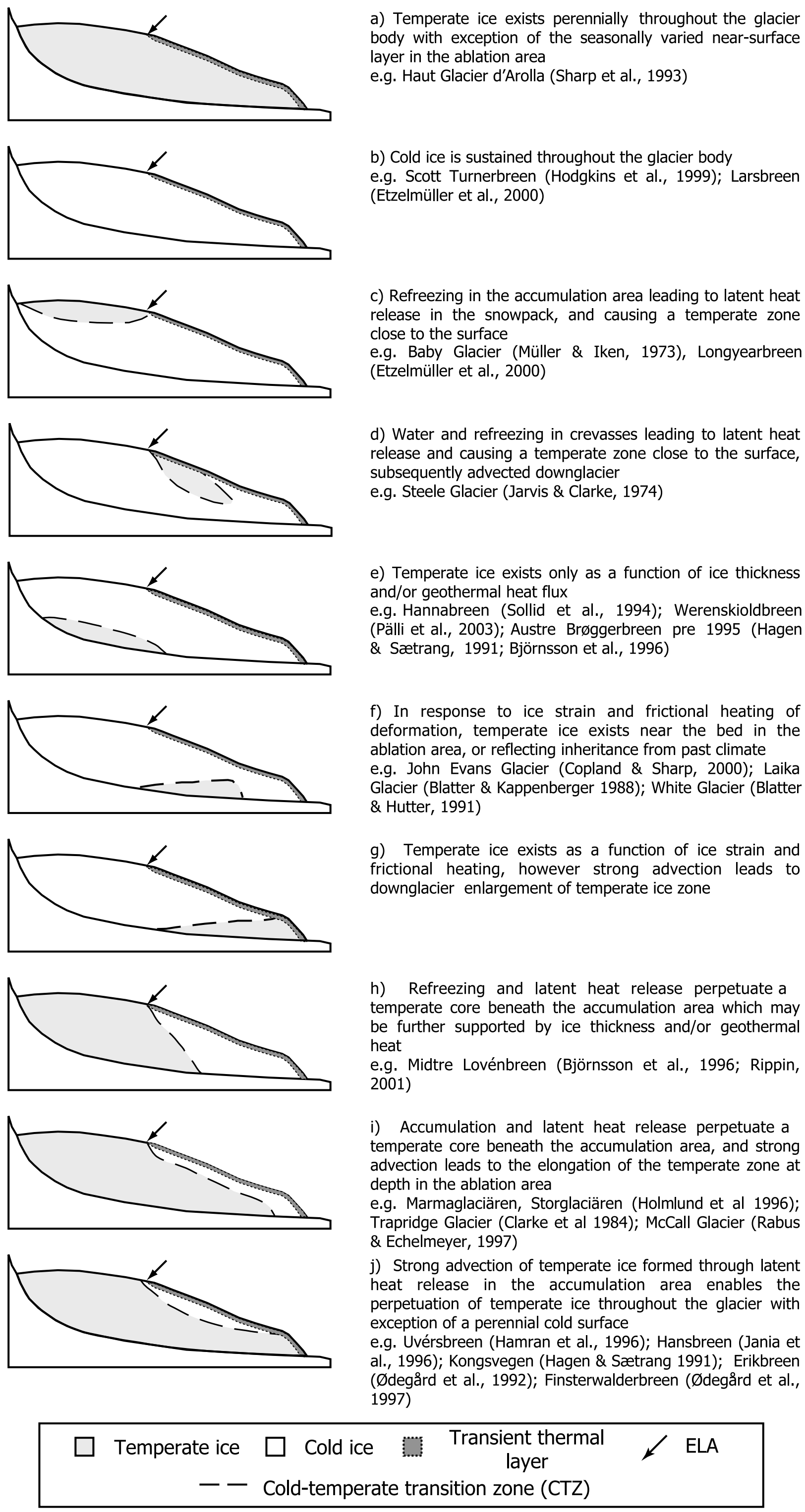

Figure 1. Illustration of the various configurations of thermal regimes documented for glaciers (developed from *Blatter and Hutter* [1991]). The equilibrium line altitude (ELA), where annual surface mass balance is zero, is indicated. With the exception of the temperate condition (Figure 1a), glaciers exhibiting any of the remaining thermal configurations may be considered "nontemperate."

2000; *Stuart et al.*, 2003]. Stagnation Glacier, Bylot Island, Canada, has highlighted a thermal lag in which current glacier recession has slowed to approximately 5 m yr^{-1}, and yet the apparent contemporary rate of recession and thinning of the temperate ice zone (~65 m yr^{-1}) is more comparable to the rapid glacier retreat that occurred more than four decades before [*Moorman et al.*, 2004]. At other nontemperate glaciers, cold ice layer thinning of up to 1 m yr^{-1} has been

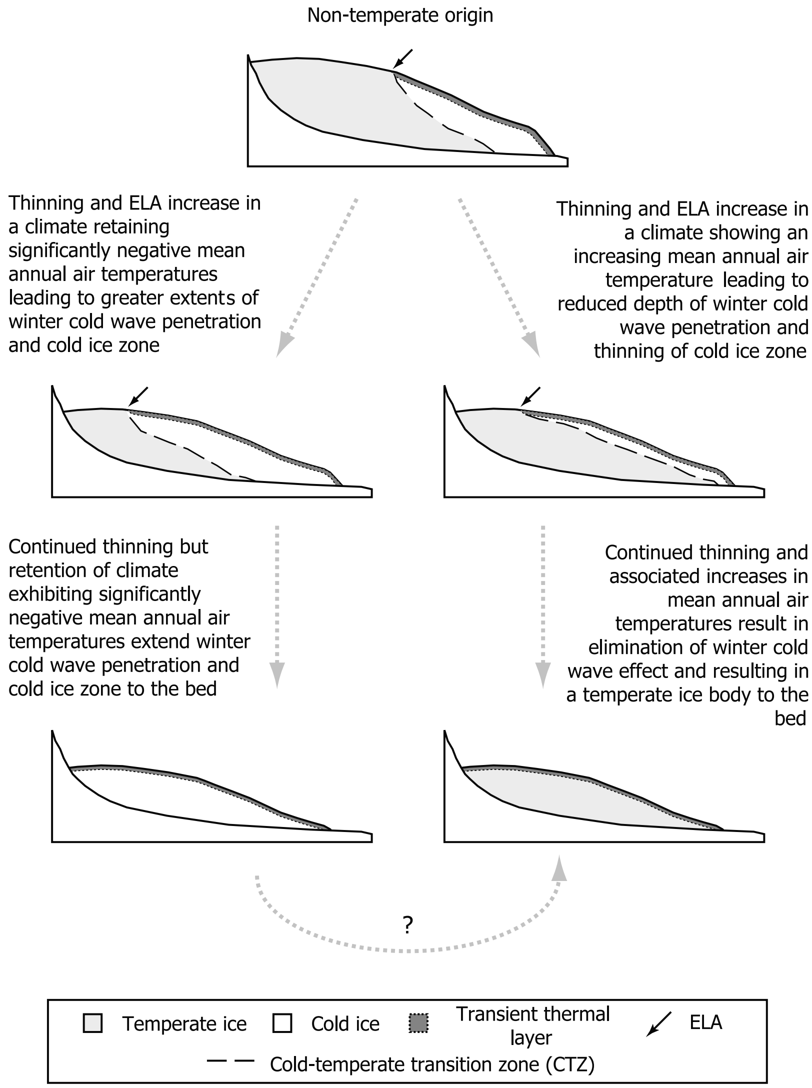

Figure 2. Schematic of the impact of glacier recession and thinning on thermal regime. It is unclear whether under continuing amelioration of climate, a cold glacier may reach a temperate state in the absence of winter cold wave penetration.

reported [e.g., *Pettersson et al.*, 2003; *Rabus and Echelmeyer*, 2002], where ice velocities may be greater than rates of basal freezing [e.g., *Moore et al.*, 2011]. Clearly, a paradox arises: Glacier thinning as a response to rising air temperatures may result in either a reduction or an increase in the spatial extent of cold ice (Figure 2). The thermal change is dependent upon the specifics of a particular glacier's location, climate, or period considered; the time frame for thermal adjustment at various depths is presently underexplored. Such adjustments and uncertainty frustrate our understanding of the hydrology of nontemperate glaciers, as sections 4.1–4.4 of this review testify.

[12] In addition to the large-scale thermal regime, it is important to recognize that the surface ice in the ablation zone is a *transient thermal layer*. During winter, the glacier surface is cooled below the PMP as a winter cold wave penetrates down from the surface due to consistently cool air temperatures [*Ahlmann*, 1935; *Paterson*, 1994]; during spring and summer, the release of latent heat as meltwater refreezes at the ice surface and progressively higher air temperatures driving ablation ensure the seasonally cold surface layer initially approaches and then remains at the PMP [*Paterson*, 1994; *Sverdrup*, 1935]. The depth of this surface temperate ice zone is limited to only several meters, as governed by latent heat release, solar energy receipt, and the poor thermal conductivity of glacier ice (κ_{ice} < 2.1 W m^{-1} K^{-1} [*Bayley*, 2007; *Paterson*, 1994]). Work at Waldemarbreen (Svalbard) has shown that over the course of an annual cycle, ice at 10 m from the surface remained cold with a temperature close to −2.5°C and exhibiting less than 1°C fluctuation, while the uppermost 1 m fluctuated between 0°C (surface PMP) during summer and −8°C in the winter months [*Sobota*, 2009].

[13] It is clear from the above discussion, and Figure 1, that there are significant variations in the temperature structure of nontemperate glaciers, and their simple classification as "polythermal" is inadequate and may be misleading. The fact that glacier thermal regimes defy simple classification is also clear in the literature, with authors using descriptors that range from the nominally qualitative (e.g.,

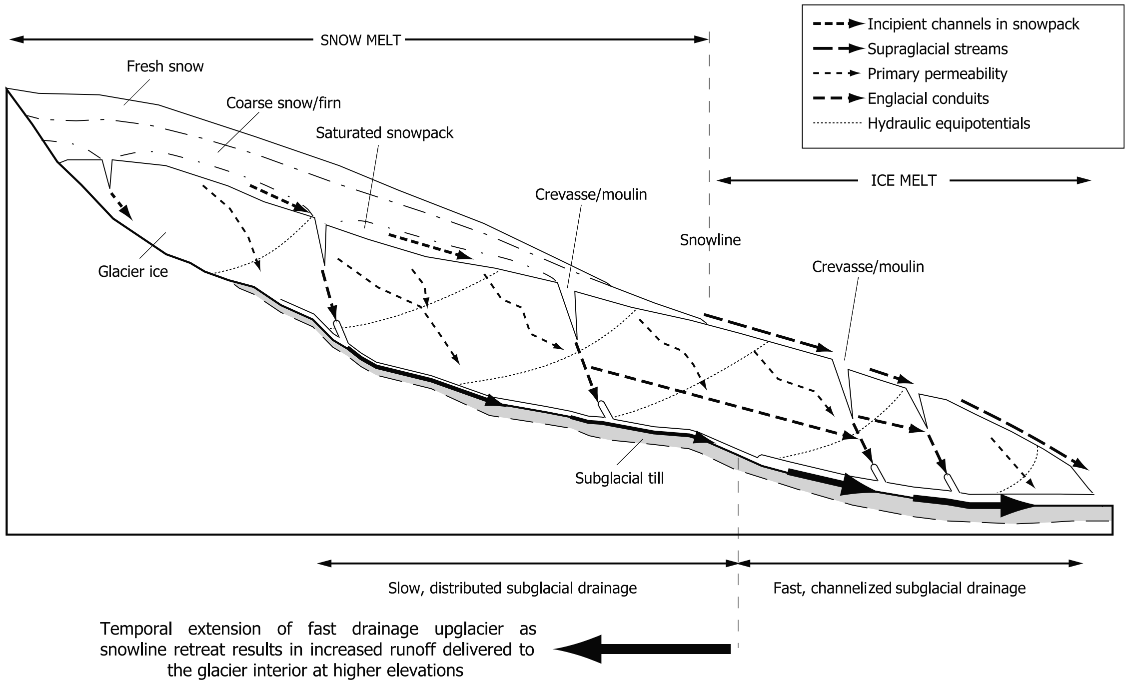

Figure 3. Temperate glacier hydrological model, after *Röthlisberger and Lang* [1987] and *Richards et al.* [1996].

"mostly cold-based" [*Etzelmüller and Hagen*, 2005]) to the quantitative (e.g., *Hodson et al.*'s [1997] use of ground-penetrating radar data to calculate the proportion of the glacier bed that is composed of temperate ice). Readers should note that some so-called polythermal and temperate glaciers may be indistinguishable (e.g., Figure 1i): Truly temperate glaciers, by definition, may have a limited occurrence, perhaps restricted to areas of >2 m of winter snowfall where both restricted penetration and seasonal elimination of the cold wave can occur [*Blatter and Haeberli*, 1984]. Here we suggest that "polythermal" should be used *only* for glaciers exhibiting a polythermal bed condition that, contrary to current convention, excludes instances where cold ice extent is limited only to narrow marginal areas and/or shallow near-surface layers (i.e., Figures 1e–1h). In view of this, the remainder of this review provides a brief overview of the hydrology of temperate ice before focusing on hydrological architecture found at nontemperate and polythermal glaciers typical of the polar regions.

3. OVERVIEW OF TEMPERATE ICE HYDROLOGY

[14] The broadly accepted, conceptual model for temperate glacier hydrology is summarized in Figure 3, for which, as was noted in section 1, extensive reviews are found elsewhere [*Benn and Evans*, 1998; *Fountain and Walder*, 1998; *Hooke*, 1989; *Hubbard and Nienow*, 1997; *Menzies*, 1995; *Paterson*, 1994; *Röthlisberger and Lang*, 1987]. However, in light of the varied extents and location of temperate ice within nontemperate glaciers, it is useful to briefly revisit a few aspects of temperate ice hydrology.

[15] The ablating thermally transient ice surface may have increased hydraulic permeability in response to enlargement of intercrystal voids by solar radiation and decreasing ice density within the uppermost 1.85 m [*LaChapelle*, 1959; *Llibourty*, 1964; *Paterson*, 1972]. Despite suggestions that meltwater might drain through the interstitial void spaces, flow rates of between 2.85×10^{-8} and 2.85×10^{-13} m s^{-1} [*Jordan and Stark*, 2001; *Raymond and Harrison*, 1975] and capillary action [*Lliboutry*, 1996] suggest this *primary permeability* is limited. *Theakstone and Knudsen* [1981] showed that the formation of supraglacial *rills* affirm that the near-surface ice permeability is low and should have a rapid hydrological response. Thus, it is the *secondary permeability* comprising crevasses and *moulins* that directs surface meltwater into the englacial environment [*Fountain and Walder*, 1998; *Röthlisberger and Lang*, 1987; *Stenborg*, 1968].

[16] The higher rate of ice deformation resulting from effective stresses in temperate ice typically limits surface crevasse depths to ~30 m [*van der Veen*, 1998b], but energy dissipation by flowing water may enlarge or propagate a conduit stemming from the crevasse or moulin. The current concepts that describe the subsequent englacial hydrology, particularly of temperate glaciers, are strongly conditioned by theoretical models of hydraulic potential presented by *Röthlisberger* [1972] and *Shreve* [1972]. Importantly, *Shreve*'s model in particular infers arborescent flowpaths but considers steady state conditions: isotropic ice and unvaried water flow. Such conditions do not reflect either glacier ice or the diurnal variations in melt rates driving water flow, and numerical models have cast doubt on the stability of a "Shrevian" englacial drainage system [e.g., *Cutler*, 1998; *Szilder and Lozowski*, 1997]. Indeed, neither the observations of temperate valley glacier moulins revealing vertical

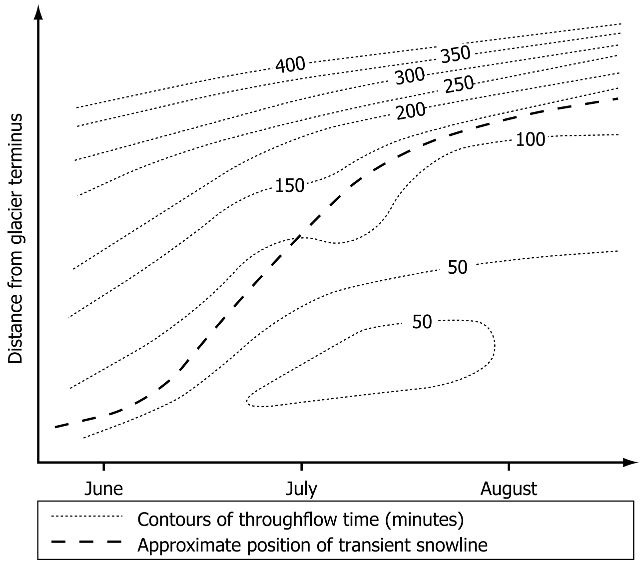

Figure 4. Idealized plot of the spatial and temporal variation in meltwater travel time to the proglacial environment as identified at Haut Glacier d'Arolla (modified from *Richards et al.* [1996]). Note the seasonal, up-glacier extension of fast transit times mirroring the retreat of the snowline.

descents depths of up to 60 m interrupted by small steps and subhorizontal galleries [e.g., *Piccini et al.*, 2000; *Reynaud*, 1987], nor the evidence of small (<0.1 m) englacial drainage structures with shallow dipping orientations [e.g., *Copland et al.*, 1997; *Harper and Humphrey*, 1995; *Raymond and Harrison*, 1975], conform to patterns driven by hydraulic potential solely governed by glacier geometry. Therefore there has been a growing realization that structural controls may provide a greater influence on the architecture of drainage features, and local hydraulic potential fields, than had previously been appreciated: *Gulley* [2009] suggested that structural and rheological characteristics were imperative for flowpath formation in temperate Matanuska Glacier, and *Fountain et al.* [2005a, 2005b] reported a network of small hydraulically connected fractures ~0.04 m in scale and commonly associated with blue ice within the extensive temperate ice of Storglaciaren, Sweden. Blue ice has an appreciable permeability and is thought to be formed either by feedbacks involved with concentrated ice strain [*Lliboutry*, 1996], by the annealing process as a crevasse closes [*Glasser et al.*, 2003; *Stenborg*, 1968], or by slow refreezing of meltwater [*Glasser et al.*, 2003; *Pohjola*, 1994], and highlights the hydraulic significance of ice rheology. A discussion of the current concerns in understanding englacial drainage is given by *Gulley et al.* [2009b].

[17] Where temperate ice exists close to and in contact with the bed, the hydraulic permeability of the lowermost several tens of meters increases [*Lliboutry*, 1996; *Rempel*, 2005]. Here it is possible that a Shrevian drainage system may develop, fed by water delivered by structurally determined flowpaths [*Fountain and Walder*, 1998], or a hydraulic connection to the subglacial environment may arise from fractures extending upward from the bed where basal water pressures exceed ice overburden pressure [*van der Veen*, 1998a]. *Rempel* [2005] highlights that the downward flow of meltwater can be impeded where subglacial water pressures are high. Nonetheless, once water reaches the discontinuity of the bed, flow is concentrated and routed through a subglacial drainage system fundamentally governed by the driving force of hydraulic potential. Subglacial drainage systems are now separated into two profoundly differing forms: fast and slow drainage [*Raymond et al.*, 1995]. The latter, in response to increasing water fluxes, will tend to evolve toward the former; however, it is likely that slow and fast drainage systems coexist both spatially and temporally [e.g., *Fowler*, 1987; *Hubbard et al.*, 1995]. The relatively rapid rate of temperate ice deformation makes it unlikely that englacial and/or subglacial conduits or drainage structures remain open once empty of water: For example, *Reynaud* [1987] observed 60% reductions in inactive, englacial conduit diameters at depths of ~100 m over a 20 day period. Current wisdom suggests drainage flowpaths close over winter months and reopen each summer and water pressures decrease as channel enlargement occurs, resulting in an up-glacier extension of an efficient subglacial system parallel to snowline retreat as secondary permeability becomes hydraulically active and progressively drives drainage inception [*Nienow et al.*, 1998; *Richards et al.*, 1996; *Sharp et al.*, 1993] (Figure 4).

[18] While there is a broad consensus of the applicability of the somewhat unified temperate glacier model (Figure 3), it has remained unverified, and clearly a number of uncertainties remain in need of further consideration. As *Fountain and Walder* [1998] had asserted over a decade ago, the supraglacial to subglacial linkage remains poorly understood, and the temporal and spatial variability in water pressures (and hydraulic potential) throughout temperate ice masses remains poorly constrained. Nonetheless, to provide an informative overview of likely meltwater flowpaths at the glacier scale, the Shrevian model of drainage has remained prominent in recent models of temperate glacier drainage [e.g., *Fischer et al.*, 2005; *Willis et al.*, 2009]. Given the complexities revealed in a glacier's thermal structure, it is necessary to lend detailed consideration of drainage processes and structures observed in nontemperate settings before comparisons or contrasts can be assessed.

4. POLYTHERMAL GLACIER HYDROLOGY

[19] Because of their extensive cold ice content, for similar geometry, many nontemperate or polythermal glaciers exhibit reduced rates of ice deformation and fracturing over considerable areas, resulting in a low crevasse density compared to their temperate counterparts [*Rabus and Echelmeyer*, 1997; *Etzelmüller et al.*, 1993]. High Arctic glaciers also tend to be located in settings with less topographic relief than their Alpine counterparts, which results in a lower aspect ratio and reduced surface slopes. As a result of these conditions, more emphasis can be placed upon supraglacial hydrological processes, which will tend to be more pervasive and evident. In instances where greater proportions of temperate ice are

found in polythermal glaciers, heightened effective stresses and ice deformation at depth can promote greater incidence of fracturing of cold ice layers, and the mechanisms behind englacial connections demand consideration.

4.1. Supraglacial Drainage

[20] At the start of the Arctic melt season, a proportion of meltwater freezing within and at the base of the snowpack and/or percolation zone forms *superimposed ice* [*Müller*, 1962; *Wakahama and Hasemi*, 1974], this can be a significant contribution to accumulation upon many High Arctic glaciers [e.g., *Koerner*, 1970; *König et al.*, 2002; *Repp*, 1988; *Wadham et al.*, 2006; *Wadham and Nuttall*, 2002; *Woodward et al.*, 1997]. Superimposed ice accentuates the reduced permeability of nontemperate glaciers so that accumulation of meltwater on the glacier surface is pronounced during the early period of the ablation season [e.g., *Hodgkins*, 2001] due to the low hydraulic transport rates of $<3 \times 10^{-4}$ m h^{-1} in snow and firn [*Schneider*, 1999]. The storage of water in slush and supraglacial lakes upon superimposed ice is a major feature upon High Arctic glaciers [e.g., *Boon and Sharp*, 2003; *Boon et al.*, 2010; *Hagen et al.*, 1991; *Hambrey*, 1984b; *Liestøl et al.*, 1980; *Liestøl*, 1988; *Müller*, 1962]. Examples from small nontemperate valley glaciers in Svalbard describe snowmelt lakes up to 3 m depth and with $\sim 1 \times 10^4$ m^3 water volumes that persist for 14–60 days before draining rapidly [*Hagen et al.*, 1991; *Liestøl et al.*, 1980; *Rachlewicz*, 2009]. These phenomena can represent a significant delay in or suppression of seasonal runoff patterns from nontemperate glacier catchments [e.g., *Braun et al.*, 2001; *Hodgkins*, 2001; *Rachlewicz*, 2009; *Schneider*, 1999; *Stenborg*, 1970; *Vatne et al.*, 1996]. Where lakes are absent, slush flows (mobilized saturated snow or firn) drain into topographic lows and can initiate or reopen antecedent supraglacial channel routes [*Boon et al.*, 2010; *Ferguson*, 1973; *Hagen et al.*, 1991; *Knighton*, 1985; *Marston*, 1983]. The processes and mechanisms of slush flows are discussed in greater detail by *Onesti* [1987].

[21] After the removal of snow cover, the surface and ice dynamic conditions upon nontemperate glaciers in the Arctic can also facilitate the development of a *weathering crust* [*Müller and Keeler*, 1969] to depths of the order of several meters in the ablation areas [*Fountain and Walder*, 1998; *Larson*, 1978]. Observations of degrading surface ice layer at nontemperate glaciers with a density decreasing to ~ 0.5 g cm^{-3} at relatively shallow (<0.5 m) depths show that the seasonal weathering crust can be enhanced by the presence of so-called *cryoconite*, the biologically active dust common to glacier surfaces that can enhance localized melting [*Fountain and Walder*, 1998; *Hodson et al.*, 2008; *Müller and Keeler*, 1969; *Takeuchi*, 2002]. The development of the weathering crust in cold regions can also be promoted by shallow (<3 m) fracturing caused by large stresses from thermal expansion [*Sanderson*, 1978; *Takahashi and Wakahama*, 1975] and the heightened permeability of small-scale ice foliations [*Wakahama et al.*, 1973]. *Wakahama et al.* [1973], *Wakahama* [1978], *Larson* [1977], and *Fountain and Walder* [1998] report meltwater velocities of between 0.06×10^{-6} and 7×10^{-4} m s^{-1} within the weathering crust, which are several orders of magnitude greater than for temperate ice (see section 3). In contrast to the capillary impedance in temperate ice, the hydraulic limit of this weathering crust developed on cold ice will be dependent on the degree of development in the transient thermal layer, which in Svalbard may only be <2 m in depth [*Sobota*, 2009]. Hydraulically, the weathering crust can enable transient water storage [*Larson*, 1977, 1978; *Liestøl et al.*, 1980], suggesting the near-surface transient thermal layer may represent a *depth-limited* or *confined aquifer*.

[22] Evidence supporting the notion of near-surface (<5 m) water storage has been provided by geophysical data showing seasonally increased water content, material heterogeneity, and radar pulse attenuation [e.g., *Irvine-Fynn et al.*, 2006]. *Larson* [1977] reported temporary specific storage of 0.007 m d^{-1} at the stagnating Burroughs Glacier, Alaska, while a supraglacial catchment at polythermal Midtre Lovénbreen, Svalbard, revealed a similar specific storage of 0.005 m d^{-1} [*Irvine-Fynn*, 2008]. The weathering crust may have an important hydraulic role in delaying or damping meltwater runoff particularly over short time scales. *Irvine-Fynn* [2008], through a water budget approach, indicated supraglacial water release appeared to occur at Midtre Lovénbreen, with a release rate of ~0.002 to 0.008 m d^{-1}, likely related to reductions in the weathering crust's development and hydraulic conductivity, refreezing, and thermal contraction. Interestingly, the formation of the hydraulically active weathering crust may also aid in the development and persistence of shallow ice surface slopes [*Fountain and Walder*, 1998; *Liestøl et al.*, 1980]. Therefore, as recently suggested at temperate glaciers [*Kopczynski et al.*, 2008; *Munro*, 2011; *Shea et al.*, 2005], the supraglacial weathering crust may also play an important hydraulic role in fashioning the *hydrograph*, especially at diurnal time scales. Its seasonal impact, however, has yet to be studied.

[23] Despite water storage processes within the weathering crust, well-developed typically dendritic surface drainage systems (although often elongated with subparallel elements) evolve at nontemperate glaciers [*Kostrzewski and Zwoliñski*, 1995]. The processes of channel inception associated with weathering crust seepage, local ice surface topography, or microscale roughness characteristics have remained underexplored. The *drainage densities* (i.e., the ratio between total channel length and area) on polythermal valley glaciers typically range between <10 km km^{-2} near the snowline and >20 km km^{-2} at lower elevations [*Brykała*, 1998a, *Brykała and Araźny*, 2000]. Broadly, these values are similar to terrestrial fluvial systems [*Knighton*, 1998] but, in being wide ranging, likely reflect the spatial changes in contributing area, surface convexity, runoff volumes, and efficacy of channelization processes over the glacier surface. Over the course of the ablation season, drainage density has been shown to increase at Waldemarbreen, Svalbard, although not necessarily directly with elevation, and long-term changes may see drainage density increase as a glacier

retreats, thinning with decreasing surface gradients, and becoming increasingly cold [*Brykała*, 1998a].

[24] *Fountain and Walder* [1998] showed that, for a highly idealized supraglacial channel, the rate of incision by flowing water on ice ($\hat{d}$) can be crudely estimated as

$$\hat{d} = \frac{1}{2}\left(\frac{\pi}{2\eta}\right)^{3/8}\left(\frac{g\rho_w}{L_m\rho_i}\right)S^{19/16}Q^{5/8}, \quad (1)$$

where η is the Manning roughness (0.01 s m$^{-1/3}$ for ice), L_m is the latent heat of melting (3.35×10^5 J kg^{-1}), S is slope (m m^{-1}), and Q represents water flux or discharge (m^3 s^{-1}). Although this equation ignores ice deformation and neglects any vertical ice motion, calculations using glacier-typical values for S and η indicate down-cutting rates (m yr^{-1}) are proportional to $Q^{0.63}$: for $Q = 0.001$ m^3 s^{-1}, $\hat{d} = 2.9$ m yr^{-1}, and for $Q = 0.1$ m^3 s^{-1}, $\hat{d} = 52$ m yr^{-1} [*Fountain and Walder*, 1998]. A similar, theoretical analysis presented by *Isenko and Mavlyudov* [2002] illustrated the incision rate being dependent upon channel width (a proxy for Q) and slope, with incision rates of the order of 0.1 to 0.2 m d^{-1}. Accounting for the vertical component of ice dynamics, an equilibrium depth will be reached where, in the ablation area, ice emergence offsets incision rates [*Fountain and Walder*, 1998]. In contrast to their entirely temperate counterparts, which are characterized by higher rates of deformation and ablation, nontemperate glacier meltwater may perennially reoccupy stable, incised supraglacial channels because discharge is less important for streams' longevity [*Hagen et al.*, 1991; *Hambrey*, 1977; *Liestøl et al.*, 1980; *Marston*, 1983]. Where nontemperate glaciers show reduced dynamics, the relative absence or stability of crevassing also may aid stream stability [*Brykała*, 1998a]. As shown by equation (1), rates of down-cutting by open channels, even with small discharges, may readily exceed maximum surface ablation rates of <10 m yr^{-1} (or <0.14 m d^{-1} assuming a 70 day ablation season). For example, observations at Longyearbreen, Svalbard, documented supraglacial stream incision rates of ~0.07 m d^{-1} relative to surface ablation of 0.01 m d^{-1} [*Müller*, 2007]. Flowing water within an ice-bounded, open channel will trend toward a positive equilibrium water temperature, typically ~0.1°C [*Isenko et al.*, 2005a], demonstrating the active process of thermal incision: Meltwater temperatures of 0.005°C–0.01°C can account for incision of between 0.02 and 0.06 m d^{-1}, and commonly the correlation between stream incision and surrounding ice ablation is not statistically significant [e.g., *Marston*, 1983; *Pinchak*, 1972]. There is, however, no consensus on the relative importance of frictional heat energy contribution to incision rates, with suggestions ranging between <50% and 100% [cf. *Marston*, 1983; *Pinchak*, 1972; *Stock and Pinchak*, 1995]; this uncertainty remains unresolved and yet is highly significant in light of the development of englacial channels, as discussed below.

[25] Supraglacial channels tend to respond to increases in discharge with more exaggerated increases in flow velocity as opposed to changes in cross-sectional area [*Knighton*, 1981; *Marston*, 1983] and, in nontemperate glacier case studies, appear to exhibit similar power relations between hydraulic geometries and stream parameters [*Kostrzewski and Zwoliñski*, 1995]:

$$Q = wdv = c_1Q^b \cdot c_2Q^f \cdot c_3Q^m, \quad (2)$$

where

$$c_1c_2c_3 = b + f + m = 1, \quad (3)$$

for which stream width, depth, and water velocity are given by w, d, and v, respectively. Exponents b, f, and m are empirically derived, and constants (c) follow the power relation given in equation (3). Typical velocities are of the order 0.1 to 1.0 m s^{-1}, and *Brykała* [1999] reported that for supraglacial streams on Waldemarbreen over a diurnal melt cycle, water velocity increase is indeed the most significant ($m = 0.49$) in response to changes in Q followed by water depth ($f = 0.38$) and then width ($b = 0.13$); however, over the course of a week, the exponents were almost equivalent and adjustments in width became more significant ($b = 0.34$). With b typically <0.4, supraglacial channels show channel widening is somewhat inhibited [cf. *Leopold and Maddock*, 1953], thus enabling the genesis of deep, relatively narrow forms. However, *Kostrzewski and Zwoliñski* [1995] argued that supraglacial streams incise through the weathering crust and into the more dense ice below. Because the weathering crust has higher thermal and hydraulic conductivity, meltwater draining to the channel coupled with local air temperatures causes the widening of channel forms over time frames greater than the diurnal melt cycle. Channel width is therefore also a function of ambient summer air temperatures (or elevation), and downstream changes seem to indicate that for higher discharges and steeper gradients, increasing supraglacial stream velocity and width are more significant than variations in water depth [*Brykała*, 1998a, 1999; *Kostrzewski and Zwoliñski*, 1995].

[26] Small fluctuations in melt production (and thus Q) can result in observable modifications to channel walls. As Q increases, lateral thermal erosion concurrent with downcutting enlarges the channel width, while as Q falls, extension of width is suppressed but incision continues: The resultant form of channel cross section is of a sequence of cuspate bank segments [*Marston*, 1983; *Stock and Pinchak*, 1995]. As is seen at temperate glaciers, both meandering (with sinuosity of 1.3–2.2) [e.g., *Hansen*, 2001] and step-pool morphologies develop increased flow resistance by reducing mean channel section slope and streambed shear stress [*Vatne and Refsnes*, 2003]. Equation (1) also demonstrates that incision rates are highest where slopes are greatest, and as such, channels tend to evolve stepped long profiles; the lateral up-glacier migration of knickpoints associated with the step-pool sequences is rapid, perhaps up to ~0.4 m d^{-1} [*Gulley et al.*, 2009a]. In contrast to fluvial environments and other supraglacial stream studies [e.g., *Knighton*, 1981], more frequent meandering has been observed to develop on steeper gradient sections of the ice

surface [*Brykała*, 1998b]. Moreover, meander bends are able to migrate down-glacier and may leave abandoned sections where incision rates result in channel avulsion [*Marston*, 1983] and, where incision rates are much greater than ablation, can meander at depth, resulting in the development of overhangs and stream course patterns different from those expressed at the ice surface. At depths in excess of the weathering crust, the thermal down-cutting is greater than lateral widening, which suggests that the morphology of overhangs may relate to air temperature and circulation rather than solely to fluvial erosion [*Müller*, 2007]. Surprisingly, the hydraulics of supraglacial streams on nontemperate glaciers has been rarely investigated but is now receiving renewed attention following work on englacial flowpaths [e.g., *Gulley et al.*, 2009a; *Vatne and Refsnes*, 2003].

[27] In summary, it seems the supraglacial environment has a number of key hydrological roles and yet, for nontemperate glaciers, continues to be poorly understood or documented. The hydraulics of the weathering crust holds some analogy to groundwater within soil, emphasizing that near-surface ice should be considered a porous media. Supraglacial processes of water storage, at both seasonal and diurnal time scales, may modulate catchment hydrographs. As at temperate glaciers, the dynamic behavior and hydraulics of the weathering crust and supraglacial streams at the drainage network and glacier scale remain unclear. These represent more urgent research problems in the case of those nontemperate glaciers subject to significant melt rates because, in many instances, a proportionally greater quantity of runoff is routed across the glacier surface than is the case with their temperate counterparts. Moreover, as revealed in section 4.2, understanding the processes of supraglacial stream inception and flowpath hydraulics demands more comprehensive study.

4.2. Englacial Drainage

[28] The reduced ice deformation and flow velocity in many nontemperate valley glacier settings, typically <10 m yr^{-1} [*Bingham et al.*, 2006; *Hagen et al.*, 1993; *Pattyn et al.*, 2005; *Rippin et al.*, 2005a], commonly results in a characteristic dearth of secondary permeability [*Hagen et al.*, 2003; *Hodgkins*, 1997; *Liestøl*, 1988]. However, discrete, sometimes large, englacial drainage features do occur: For example, Austre Brøggerbreen, Svalbard, a now cold-based glacier with a crevasse-free surface, displays several large moulins [*Hagen et al.*, 1991; *Holtermann*, 2007; *Stuart et al.*, 2003; *Vatne*, 2001]. Clearly, drainage networks can exist within cold ice zones contradictory to previously held assumptions that the environment was too cold to support effective englacial hydraulic connections [e.g., *Paterson*, 1994]. A number of mechanisms have been proposed for the development of englacial drainage through cold ice, with recent, major advances in these conceptual models.

[29] At nontemperate glaciers, typical supraglacial streams can incise at rates of up to 0.3 m d^{-1} [*Gulley et al.*, 2009a; *Isenko and Mavlyudov*, 2002] and develop into englacial conduits [*Brykała*, 1998a; *Hagen et al.*, 1991, 1993; *Hodson*, 1994; *Liestøl et al.*, 1980; *Vatne*, 2001]. This englacial flowpath formation process, previously noted by *Röthlisberger and Lang* [1987], has recently been formalized in the literature by *Gulley et al.* [2009a] as "cut and closure" channels (Figure 5a). Importantly, this process ensures that the input of water to the englacial environment does not necessitate structural controls such as foliation, fractures, or crevasses. The persistence of snow bridges and metamorphosed and deforming ice over these incised flowpaths results in enclosed, high-roofed englacial channels (Figure 6). Meltwater refreezing on the incised channel walls near the ice surface may also aid in the isolation of a supraglacial stream within the englacial environment [*Vatne*, 2001], as can early season slush flows associated with supraglacial snowmelt floods [*Gulley et al.*, 2009a]. *Schroeder* [2007] has also suggested that where the incised channel valley becomes bridged by winter snow, internal air circulation and heat exchanges may also aid in the consolidation and regelation of the snow bridge. Because observed closure rates in these channels appear relatively low (between 0.07 m yr^{-1} [*Hansen*, 2001] and 0.14 m yr^{-1} [*Gulley et al.*, 2009a] near the glacier surface, and ~1 m yr^{-1} at <70 m depth [*Schroeder*, 1998a]), channel geometry may vary in space but cross-sectional area may remain large and drainage will usually be at atmospheric pressure [*Stuart et al.*, 2003; *Vatne*, 2001]. Such channels dominated by vertical incision can be termed *vadose channels*, which typically incise to and flow at a locally defined base level, particularly where incision is controlled by hydraulic barriers such as the glacier bed or basal *overdeepenings* [e.g., *Vatne*, 2001], a process also hypothesized in the temperate ice of nontemperate Storglaciaren [*Hooke and Pohjola*, 1994].

[30] The meander geometry of cut and closure englacial channels often mirrors a supraglacial form. Meander wavelengths of 10–50 m, sinuosity of 1.2–2.0, and lateral meander development rates of ~9 m yr^{-1} have been reported in Svalbard [*Hansen*, 2001; *Vatne*, 2001]. Where channel slopes are >0.3 m m^{-1}, step-pool sequences are more common [*Griselin et al.*, 1994; *Vatne*, 2001]. Contrasting to the more typical relationship for alluvial and supraglacial settings, *Hansen* [2001] reported that meandering of englacial conduits was inversely related to channel width, although this remains to be explored further. Cuspate forms on the ice walls indicate, like supraglacial streams, that the channel may cope with large changes in water flux by adjusting channel width over short, subseasonal temporal scales (Figure 6). Other morphological developments on ice channel walls can be formed by both circulating air and water, with size and shapes dependent on differing flow viscosities and velocities. The time frame of such morphological adjustments, however, remains uncertain.

[31] Given the formation process, it was initially believed that cut and closure channels would present a gradual, initial descent suggestive of supraglacial down-cutting below the ice surface [*Vatne*, 2001]. However, a large number of englacial channel explorations in Svalbard show very steep initial sections, with vertical shafts up to 80–100 m (approximately equal to the cold ice limit) interrupted by

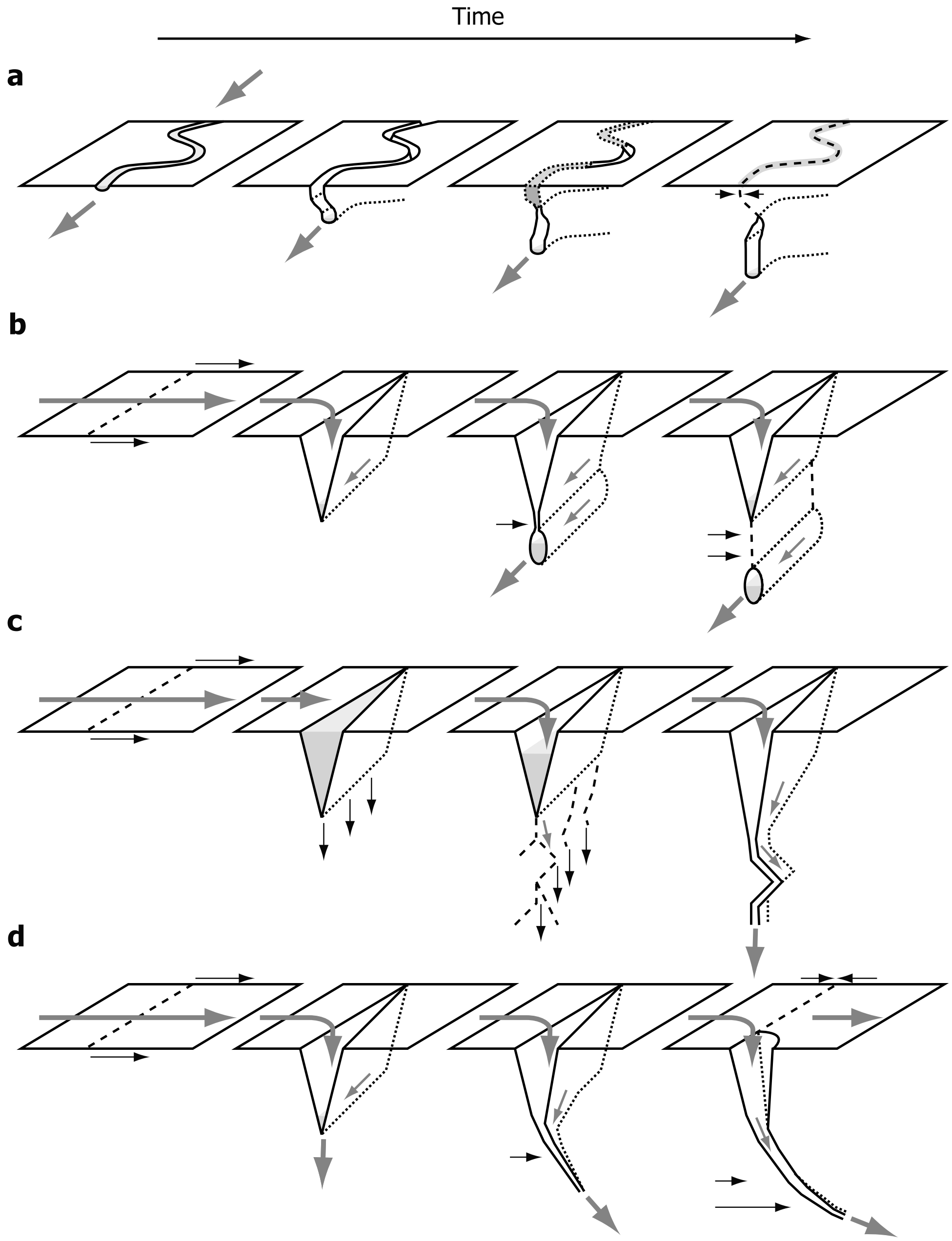

Figure 5. Schematic of the mechanisms of englacial drainage route formation in polythermal glaciers. Ice flow is left to right, water flow is represented by gray arrows, while conceptual forces acting on ice are shown by small black arrows.

horizontal or gently inclined sections, before yielding to a more subhorizontal, englacial route [e.g., *Eraso et al.*, 1992; *Griselin et al.*, 1994; *Řehák et al.*, 1990; *Schroeder*, 1994]. *Vatne and Refsnes* [2003] suggest that the englacial channels evolve to maximize flow resistance and channel stability, and initial meanders may develop into step-pool sequences, with pools deepening as a result of the excess heat gained in flow over steep stream sections and resulting in a positive feedback amplifying the step-pool geometry, and controlling the formation of subhorizontal channel sections which may then meander when a local base level is reached. The steps or knickpoints can migrate up-glacier with rates of 0.18–0.5 m d^{-1} [e.g., *Gulley et al.*, 2009a; *Hansen*, 2001] because melt rates are proportional to discharge and slope, and the equilibrium temperature of flowing water increases with channel size and slope [*Isenko et al.*, 2005a], providing a positive thermal feedback. Upon migration, large knickpoints will coalesce with smaller ones, resulting in relatively large steps [*Gulley et al.*, 2009a; *Hansen*, 2001]. Observations suggest steps of the order of 5–20 m in vertical dimensions are common [e.g., *Griselin et al.*, 1994; *Řehák et al.*, 1990; *Schroeder*, 1994; *Vatne*, 2001]. *Gulley et al.* [2009a] hypothesized that the development of step-pool sequences could result in the evolution of the cut and closure channel to a form that mimics a crevasse-formed moulin. Repeat investigations at Austre

Figure 6. Image of the englacial flowpath at Austre Brøggerbreen (image courtesy of G. Vatne). Note the cuspate forms on the channel walls, the high channel roof, and the stepped sequence in the channel floor.

Brøggerbreen show that, indeed, the rates of englacial knickpoint recession are of the order of 20 m yr^{-1} and over a period of 11 years, a gently sloping englacial channel was observed to develop into a 37 m deep moulin by knickpoint recession (Figure 7). Moreover, through this process, down-glacier motion of the englacial passage can be partially offset by up-glacier extension of the drainage structure, resulting in moulins with somewhat static positions and ensuring their longevity. This formation process enables the development and maintenance of what might be termed "classical moulins," which would appear anomalous on glaciers devoid of crevasses.

[32] The alternative formation of deep initial descents into the englacial environment is through crevassing or fracturing, as occurs at temperate glaciers. *Fountain and Walder* [1998] proposed that subhorizontal drainage structures could form simply by water flow in crevasse bottoms with incision rates given by equation (2); subsequent crevasse closure or ice creep will isolate or "pinch off" the drainage structure from direct connection to the supraglacial environment [*Fountain and Walder*, 1998; *Stenborg*, 1968] (Figure 5b). Support for this crevasse-base hypothesis has been presented: For example, at nontemperate Storglaciaren, *Holmlund* [1988] showed evidence of drainage structures oriented according to crevasse direction at depths of 60 m

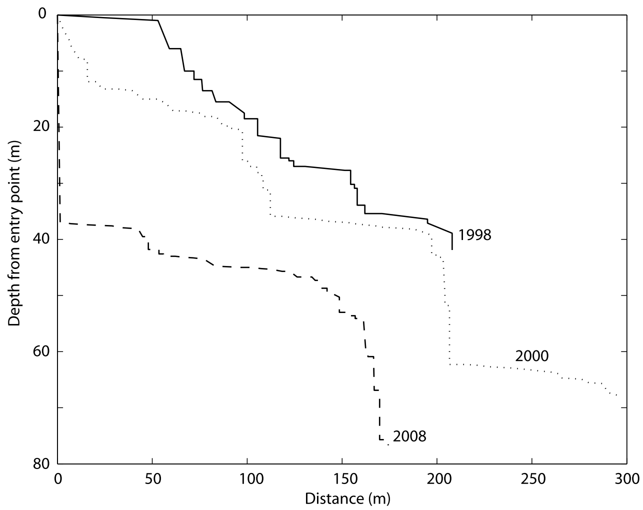

Figure 7. Schematic documenting the englacial flowpath and moulin development at Austre Brøggerbreen between 1998 and 2008 (G. Vatne, manuscript in preparation, 2011). Note the knickpoint recession, resulting in the deep moulin shaft apparent in 2008.

from the ice surface; these extended through cold ice into underlying temperate ice and were in excess of anticipated crevasse depths. *Mavlyudov* [1994] provided a conceptual model suggesting that crevasse-bottom incision results in the formation of sinuous englacial connections where transverse crevasses occur at the lateral ice margin or in regions of extending flow, while irregular crevassing can result in less structured hydraulic connections (Figure 8), an idea supported elsewhere [e.g., *Schroeder*, 1998a]. Where a crevasse or moulin becomes water-filled, the hydraulic head can promote fracture extension at the feature base [*Benn et al.*, 2007; *van der Veen*, 1998b; *Weertman*, 1973] (Figure 5c). Indeed, *Benn et al.* [2009a] assert that hydraulically driven fracture augmentation may play a crucial role in the delivery of water to a nontemperate glacier's interior. *Alley et al.* [2005b] consider the rate of hydrofracture propagation, accounting for refreezing at the fracture boundaries, giving a downward velocity of

$$u = \frac{QM}{-4\sigma'_T d_f \vartheta}, \qquad (4)$$

where Q is discharge (inflow) per unit length, u is deepening velocity, $M = 5 \times 10^9$ Pa, d_f = fracture depth, $\vartheta = 1.8 \times 10^{-3}$ Pa s^{-1}, and σ'_T is longitudinal crack-forming deviatoric stress. However, this was not explored further by *Alley et al.* [2005b]; rather, conclusions were drawn that high Q is required to maintain water pressure to continue the fracture process, especially in light of refreezing against cold ice boundaries, thus inciting the need for water storage (supraglacial lakes) and/or high tensile stresses. Observations at valley glaciers across the Arctic have shown supraglacial lakes to drain suddenly through fractures [e.g., *Boon and Sharp*, 2003; *Hambrey*, 1984b]. And *van der Veen* [2007] shows that even for large variations in tensile stress the depth of fracture penetration depth (d_f) can be reasonably approximated simply as a function of Q and time (t):

$$d_f \approx \left(\frac{p_w}{p_i}\right)^{2/3} Qt. \qquad (5)$$

[33] Although this latter model does not include the refreezing process, it is argued that the rate of refreezing compared to the rate of fracture penetration (assuming constant inflow which itself offsets refreezing) makes the effect negligible. In propitious circumstances, crevasses can propagate to depths >1 km in cold ice over short time periods (days) [*van der Veen*, 2007], suggesting that hydrofracture may connect the glacier surface to the ice bed [*Benn et al.*, 2009a]. Positive feedbacks between hydrology and ice dynamics may assist in accentuating or maintaining the hydrofracture connection [*Boon and Sharp*, 2003]. In the absence of a continuous water flux, refreezing and plugging or pinching off of water at the base will reduce the likelihood of water descending to greater depths [*Alley et al.*, 2005b; *Boon and Sharp*, 2003; *Weertman*, 1973]. However, this process is slow: For example, at an ice temperature of −20°C, closure by refreezing on crevasse sides occurs at a rate of 0.015 m d^{-1} [*van der Veen*, 2007], and thus a moulin in excess of typical crevasse depth can be readily formed. Moreover, it is possible that a sequence of hydrofracturing

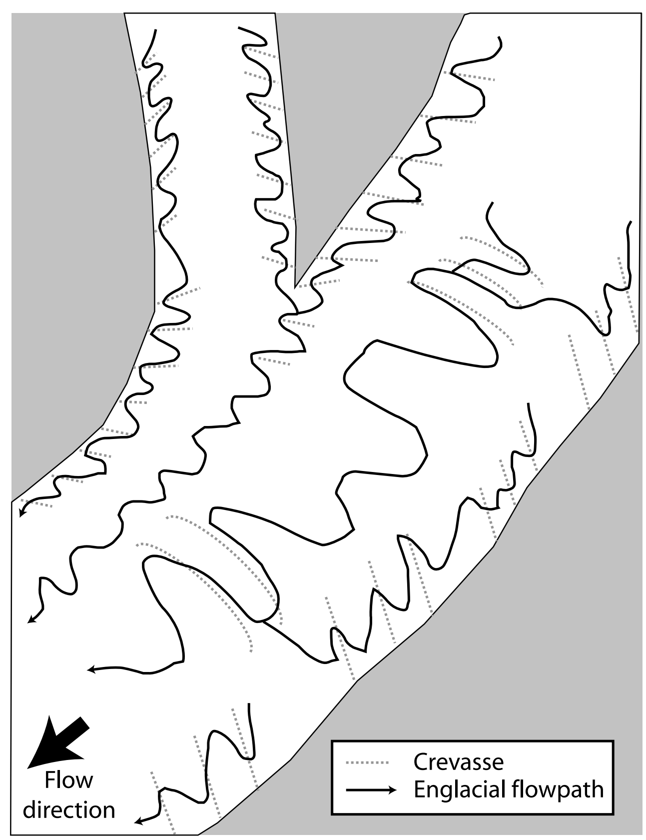

Figure 8. Conceptual model of englacial drainage route configuration in polythermal glaciers [after *Mavlyudov*, 1994].

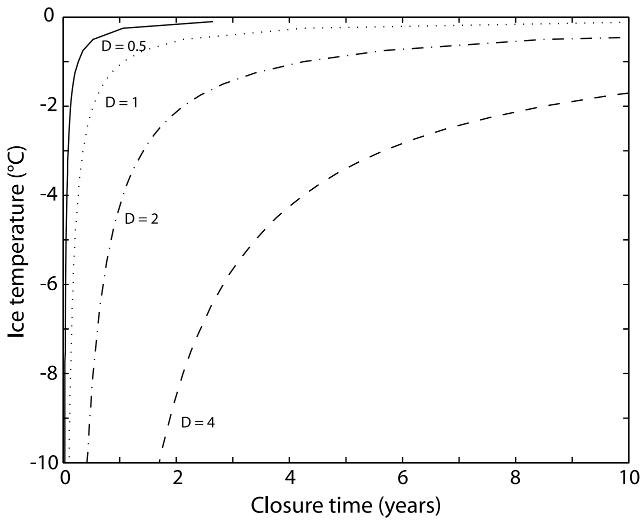

Figure 9. Graph illustrating the closure rates for water-filled englacial channels with differing diameters (D, in meters).

can occur, particularly if there is a cyclical nature of crevasse-filling and fracturing. The changes in effective stress field with increasing depth can result in the fracture of the flowpath becoming inclined at progressively shallower angles (Figure 5d) as has been observed at some polythermal glaciers [e.g., *Iken*, 1972, 1974]. Furthermore, hydrofracture can also spread laterally where there is a zone of minimal tensional resistance [see *Roberts et al.*, 2000]. *Mavlyudov* [2005a] reported a number of sites exhibiting horizontal hydrofractures and presented results from a numerical model that suggested that water flow for a duration of 9 h into such a fracture resulted in the inception of a channel form.

[34] Active fractures are not the only structure capable of carrying supraglacial meltwater to the polythermal englacial environment. Numerous cave explorations within cold ice have observed drainage architecture clearly related to ice structures, such as refrozen or annealed crevasses and fractures, which can provide locations for channel inception [*Benn et al.*, 2009a; *Řehák et al.*, 1990; *Schroeder*, 1998a; *Vatne*, 2001]. *Gulley and Benn* [2007] suggested that for a nontemperate Himalayan glacier, structural weaknesses such as fractures, debris bands, *thrust faults*, blue ice, or refrozen fractures may provide locations of hydraulically transmissive sites, facilitating flowpath inception and enlarging with increasing water flow. However, in cold ice, the potential refreezing of slowly percolating meltwater, in the absence of a continual flux of meltwater, would be likely to impede drainage development along very small fractures. Nonetheless, at subseasonal time scales, *Irvine-Fynn et al.* [2006] showed a number of activated and deserted hydrological features within a zone of cold ice. Disparate, isolated voids within the englacial environment are potentially remnants of abandoned or detached englacial hydrological systems [*Pohjola*, 1994], which highlight the likelihood of hydraulic preservation in polythermal glaciers. Cave explorations have reported small, active and inactive, elliptical or circular *phreatic channels* connecting with the much larger scale vadose or fracture-based channels [e.g., *Řehák et al.*, 1990; *Schroeder*, 1998a; *Vatne*, 2001]. The hydraulic efficiency and longevity of these fracture-based, vadose, or phreatic flowpaths is, however, dependent upon the mechanisms of channel closure.

[35] The closure rate (r_c) of an air-filled circular englacial channel due to ice creep can be given by

$$r_c = RAn^{-n}p_i^n, \tag{6}$$

where R is the conduit radius, p_i is the ice overburden pressure, and the ice deformation parameter A is temperature dependent; in most instances $n \approx 3$ [see *Paterson*, 1994; *Röthlisberger and Lang*, 1987]. With vadose channels, the closure rate is dictated by the ice thickness and temperature, thus likely to show an increase with depth as noted above, a critical consideration for both englacial and subglacial domains (see section 4.3). However, if a channel is water-filled, determination of the closure time (t_c) only needs incorporation of the effect of refreezing and, following *Lunardini* [1988],

$$t_c = \frac{LR^2}{4\kappa_{ice}T_i}, \tag{7}$$

for which L is latent heat of fusion and T_i is ice temperature (Figure 9). Thus, as both *Fountain* [1993] and *Mavlyudov and Solovyanova* [2003] suggest, ice temperature exerts a fundamental control on cavity dimensions. Yet numerical experiments suggest ice depth is more critical to englacial channel survival than ice temperature but show the potential longevity of englacial drainage networks in relatively thin cold ice within the ablation zone of a polythermal glacier [e.g., *Isenko et al.*, 2005b]. Other researchers have proposed that air circulation and latent heat exchange with liquid water in cold ice zones may further aid the persistence of ice-walled channels in the absence of continued meltwater provision [*Eraso*, 1992; *Schroeder*, 2007]. The net result is that englacial channels can persist in cold ice and may be reoccupied year after year. The longevity may extend up to 50 years earlier [e.g., *Schroeder*, 1998b], especially given that cold ice in the upper layers of most nontemperate glaciers prevents ice deformation from closing crevasse or moulin entrances to the englacial environment over annual time scales [*Vatne*, 2001]. Consequently, the englacial flowpaths represent a large reservoir that may store water in the glacier interior if outflow is impeded [*Schroeder*, 1998a], and seasonal change in the interstitial water content within the temperate ice of a polythermal glacier [*Irvine-Fynn et al.*, 2006] also indicates the englacial environment to behave as an aquifer.

[36] Evidence for longer-term changes in hydrological connections comes in the form of abandoned, fossil moulins, or "crystal quirks" (Figure 10), which are apparent on nontemperate glacier surfaces [*Holmlund*, 1988; *Mavlyudov and Solovyanova*, 2003; *Stenborg*, 1968]. These features form where an englacial flowpath becomes flooded and refreezes, caused either by glacier motion isolating moulins from their original meltwater supply or by internal changes

Figure 10. A crystal quirk: an indication of refrozen moulin forms. Note the radial crystal structure, emphasized by the distribution of cryoconite.

sealing off a hydraulic connection. Damming of englacial channels (and vadose channels approaching the bed) is more likely to occur at depth and is caused by (1) ice creep, (2) deposition of slush flows, (3) in situ freezing of meltwater volumes particularly if inflow water discharges are reduced, or (4) by substantial ice collapse [*Griselin*, 1992; *Griselin et al.*, 1994; *Gulley et al.*, 2009a]. Where the water flux is far in excess of channel enlargement and water evacuation, hydraulic damming can occur and water pressures may exceed ice overburden pressures. In these instances, the channel course can be readily rerouted to higher englacial elevations as processes of cut and closure, hydrofracturing, or channel inception begin again [*Gulley et al.*, 2009a]. Newly formed hydrofractures will be directed to zones of reduced ice overburden pressure, and so it is plausible that hydraulic damming can propagate fractures upward to connect with other preexisting or developing drainage features (e.g., moulins and vadose channels [see *Roberts et al.*, 2000]). In temperate ice, such vertical flowpath reorganization [e.g., *Hooke and Pohjola*, 1994; *Kirkbride and Spedding*, 1996; *Lliboutry*, 1983] has been termed a *hydraulic jump*. Evidence for such restructuring in nontemperate glaciers has been provided by direct englacial observations revealing sequences of relict or inactive drainage structures, as both voids and frozen features [e.g., *Benn et al.*, 2009a; *Gulley et al.*, 2009a; *Vatne*, 2001; *Vatne and Refsnes*, 2003].

[37] In summary, englacial drainage structures appear to be common in polythermal glaciers, yet they are less likely to form in uncrevassed areas [*Gulley et al.*, 2009b; *Mavlyudov*, 1994; *Stenborg*, 1968] and do not necessarily convey water directly to a subglacial system. The processes of cut and closure, hydrofracturing, and channel inception at structural discontinuities appear to be the most important [*Gulley et al.*, 2009b], resulting in a drainage structure composed of both vadose and fracture channels [*Mavlyudov*, 1994]. Glacial confluences provide apt locations for such flowpath development, where ice structures and stress fields are likely to be considerably disturbed [*Benn et al.*, 2009b; *Gulley et al.*, 2009b; *Mavlyudov*, 1994]. Consideration of englacial drainage, at least partially, as a *fracture network*, also appears to hold considerable promise for the future modeling of nontemperate glacier hydrology [*Flowers and Clarke*, 2002; *Fountain et al.*, 2005b; *Irvine-Fynn et al.*, 2005a]. Critically, while similarities to temperate ice hydrology can be identified, it remains somewhat unclear how, in polythermal glaciers, the drainage structures developed through cold ice zones subsequently interact and develop with englacial drainage of temperate ice areas.

4.3. Comments on Subglacial Drainage

[38] Initially, the presence of a cold-temperate ice transition zone (the CTZ) was conceptualized as a "thermal dam," imparting profound influence upon ice dynamics and hydrology by partially or entirely precluding both regelation and meltwater flow down-glacier from the CTZ [*Blatter and Hutter*, 1991; *Clarke et al.*, 1984; *Jania et al.*, 1996; *Jarvis and Clarke*, 1975]. The perceived influence of the "dam" might explain the early assumptions that the hydrology of polythermal glaciers was unable to develop to the extent thought to exist at temperate glaciers [e.g., *Menzies*, 1995; *Rabus and Echelmeyer*, 1997]. However, the emergence of turbid, highly mineralized water from beneath nontemperate glaciers has long been observed [e.g., *Hambrey*, 1984a; *Hodson and Ferguson*, 1999; *Müller and Iken*, 1973; *Skidmore and Sharp*, 1999; *Vatne et al.*, 1995, 1996; *Wadham et al.*, 1998], suggesting that the CTZ certainly does not preclude the drainage of subglacial water into the forefield. Indeed, research has shown thin films of water can exist at the ice-rock interface beneath cold ice areas due to the hypersalinity of the mineralized waters resulting from solute rejection during pressure-driven melt and regelation [e.g., *Cuffey et al.*, 1999; *Echelmeyer and Zhongxiang*, 1987]. However, still relatively little is known about the form of active subglacial drainage beneath nontemperate ice masses.

[39] The messages highlighted in sections 4.1 and 4.2 stressed that the manner in which supraglacial meltwater accesses the bed of nontemperate glaciers results in a marked heterogeneity in input distribution, and drainage structures through cold ice may incise to the ice-bed interface. However, water reaching the glacier bed may also be sourced from basal and frictional heating in the temperate ice areas, as discussed in section 2 [e.g., *Aschwanden and Blatter*, 2005]. Further, the shortened summer season and supraglacial water storage at high latitudes will also have an influence on the development of any subglacial drainage system. Nevertheless, at several polythermal glaciers in Svalbard, research has demonstrated that subglacial drainage structures appear to correspond to drainage paths determined by subglacial hydraulic potential assuming glacier geometry and water pressures approaching that of ice-overburden [*Hagen et al.*, 2000; *Pälli et al.*, 2003; *Rippin et al.*, 2003]. These examples are glaciers exhibiting predominantly temperate ice-bed interfaces (e.g., Figures 1h and 1i). Contrastingly, for colder-bedded John Evans Glacier (Figure 1f), *Copland and Sharp* [2001] suggested bed

topography and lowered water pressures dictated the apparent hydraulic configuration across the glacier bed. *Pälli et al.* [2003] concluded that in some polythermal glaciers, low water pressures may occur up-glacier of the CTZ, even during the ablation season; the inference was that meltwater delivery to the subglacial system may be diminished and that the cold terminus may cause channel constriction (hydraulic damming), causing heightened water pressures close to the snout compared to up-glacier locations. A similar water pressure configuration was suggested for nontemperate Stagnation Glacier [*Irvine-Fynn et al.*, 2006] in response to the observation of the drainage of interstitial water from the temperate ice zone along a path coincident with topographic lows at the glacier bed. Where the influence of bed topography on subglacial hydrology is great, either discharge is sufficiently high that the rates of drainage channel enlargement are in excess of rates of closure [*Hooke*, 1984] or meltwater discharge is lower than the capacity of a stabilized subglacial system [*Sharp et al.*, 1993]. Given the reduced rates of ablation, and despite the temporary retention of water in the snowpack in polar regions, the former is less likely, while the processes of englacial flowpath formation discussed above would indicate englacial channels could descend to, and persist at, the glacier bed. Such incision could leave relatively large drainage system capacity with stable channels even in the thinner, colder regions close to the snout. Certainly, *Gulley et al.* [2009a] argue that cut and closure vadose channels may cut down to the glacier bed to form subglacial pathways in both temperate and cold ice zones. This appears to be the process by which subglacial channels at Tellbreen have been formed and remain directly linked to supraglacial drainage despite the fact the glacier now displays a dominantly cold thermal configuration akin to that shown in Figure 1e [*Bælum and Benn*, 2011]. Although in some instances such channels may have considerable thicknesses of overlying ice, the cold temperatures and reduced effective stresses can promote the persistence of developed drainage architecture. Where the overlying ice is temperate, basal water may connect with other regions of the glacier bed; where temperate ice areas are spatially discontinuous, connections may be established; however, where significant portions of the glacier are frozen to the bed, such flowpaths are unlikely to be hydraulically connected to other regions of the subglacial environment [*Gulley et al.*, 2009a].

[40] As *Fountain and Walder* [1998] emphasized, most glaciers are likely to be underlain by a spatially variable or discontinuous layers of rock debris or rock-ice mixes, but the hydraulic conductivity is likely to be low ($<1.9 \times 10^{-5}$ m s^{-1}) due to the processes of regelation and/or cold wave penetration [*King et al.*, 2008; *Kulessa and Murray*, 2003]. The subglacial drainage system at polythermal Trapridge Glacier is thought to be dominated by distributed water flow in and over a permeable sediment layer, rather than in subglacial channels, with flow velocities of ≤ 0.1 m s^{-1} [*Stone and Clarke*, 1996]. Nonetheless, some oddities have been observed in that subglacial system; for example, *Murray and Clarke* [1995] recorded transverse effects with oscillations in water pressures in subglacial channels out of phase with water pressures at distances away from the channels, and *Kavanaugh and Clarke* [2000] reported transient high water pressure pulses thought to be generated in response to rapid changes in water volumes or abrupt glacier motion affecting the existing drainage system even at significant distances from the point of observation. These observations clearly demonstrate that even in a slow drainage system, hydraulic connections and variations at a point can be influential over large subglacial areas. Importantly, under certain conditions the subglacial hydrology may link to groundwater beneath the temperate ice zone: Unfrozen areas or *talik* may readily survive beneath many High Arctic glaciers, insulated by the overlying ice itself [*Hagen et al.*, 1993; *Liestøl*, 1977], resulting in subglacial water interacting with or developing or utilizing groundwater flowpaths [e.g., *Griselin et al.*, 1994; *Haldorsen et al.*, 1996; *Haldorsen and Heim*, 1999; *Lauritzen*, 1991], especially over karst-prone terrain [*Ford*, 1993].

[41] Despite some interpretations being drawn on subglacial drainage systems beneath polythermal glaciers, this remains a key area of uncertainty—as with all glaciers. The forms of subglacial flowpaths are assumed to reflect fast and slow drainage types theorized for temperate ice, and so often focus has been directed to the inferences that can be drawn from observations of seasonal outflow. However, spatially, the combination of discrete meltwater delivery and thermal variation would suggest that the spatial and temporal transitions of a subglacial system as idealized for wholly temperate glacier beds may not apply definitively, and local water storage may readily occur.

4.4. Seasonal Outflow

4.4.1. Summer

[42] Evidence of water storage and increasing englacial or subglacial water pressures following the onset of the melt season has been observed though vertical ice surface displacement at John Evans Glacier, Canada [*Bingham et al.*, 2006]. This water storage can continue until water pressures exceed some notional threshold value and, consequently, a drainage system develops, enabling water to breach or rupture any constrictions imposed by the CTZ or icing at the glacier margin [*Bingham et al.*, 2006; *Jania et al.*, 1996; *Schroeder*, 1998a; *Skidmore and Sharp*, 1999; *Wadham et al.*, 2001]. As a consequence, seasonal meltwater outflow is lagged relative to the onset of the melt season, at least until meltwater release, usually in the form of artesian discharges, or *upwellings* [e.g., *Baranowski*, 1973; *Hambrey*, 1984a; *Hodson et al.*, 2005; *Hodson and Ferguson*, 1999; *Iken*, 1974; *Irvine-Fynn et al.*, 2005c; *Jania et al.*, 1996; *Pälli et al.*, 2003; *Skidmore and Sharp*, 1999; *Wadham et al.*, 2001]. The nature and timing of the lag appear to be determined by the mechanics of the outflow constriction, or they may relate to the volume of meltwater produced or stored within the glacier prior to and during the initial ablation period [*Hodson et al.*, 2005; *Skidmore and Sharp*, 1999; *Wadham et al.*, 2001]. Therefore the manner in which upwelling or outflow occurs is variable both in

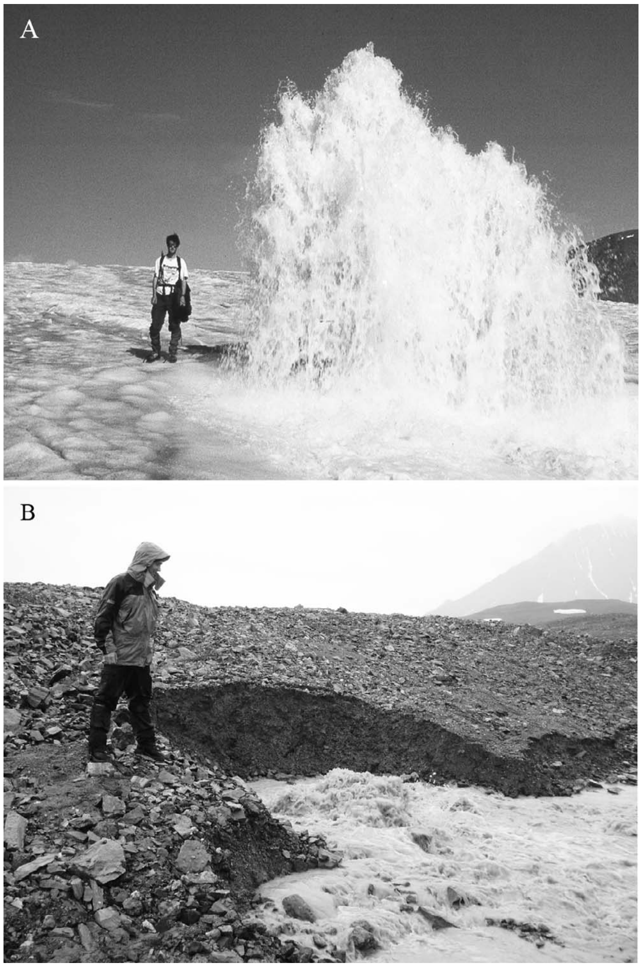

Figure 11. Upwellings observed at polythermal glaciers (a) in the supraglacial environment at John Evans Glacier (image courtesy of M. J. Sharp) and (b) the proglacial environment at Midtre Lovénbreen (image courtesy of T. D. L. Irvine-Fynn).

time and space and between differing polythermal glaciers (Figures 11 and 12).

[43] In some examples, outflows are evident as upwellings upon the ice surface in both accumulation and ablation areas [e.g., *Baranowski*, 1973; *Glowicki*, 1982; *Irvine-Fynn et al.*, 2005c; *Rucklidge*, 1956; *Skidmore and Sharp*, 1999]. These supraglacial fountains are often only relatively short-lived (several days) but clearly indicate enhanced water pressures exceeding ice overburden pressure within the drainage system: High water pressures at depth can enable weaknesses in the glacier ice to be exploited [*Skidmore and Sharp*, 1999] or promote basal fractures [*van der Veen*, 1998a]. Ice *thrust faults* [e.g., *Hambrey and Müller*, 1978], longitudinal folding, foliation, and other structural discontinuities in the ice represent the sorts of features exploited by meltwater emerging from the glacier in this manner [e.g., *Hambrey and Glasser*, 2003; *Mavlyudov*, 2005a; *Murray et al.*, 2000]. The processes of hydrofracture (section 4.2) driven by heightened water pressures at or close to the glacier bed may also explain water ascent though the ice mass to the surface [e.g., *Roberts et al.*, 2000]. Similarly, with suggestions that large englacial flowpaths are readily preserved in cold ice, it is possible that these present avenues to carry water to the surface at times of high water pressures [e.g., *Dewart*, 1966].

[44] In other examples, seasonal outflow has occurred through channel openings or *portals*, or the initial upwelling has developed swiftly into a portal at the margin [e.g., *Boon and Sharp*, 2003; *Irvine-Fynn et al.*, 2005c; *Rippin et al.*, 2003], for which two alternative mechanisms have been suggested: englacial routing and *hydraulic jacking*. The former relates to the opening or reopening of vadose englacial drainage pathways through cold ice (at least initially) blocked by frozen zones, potentially linked to hydraulic jumps, as detailed above. The latter involves meltwater volumes in excess of the drainage system capacity, which prompts

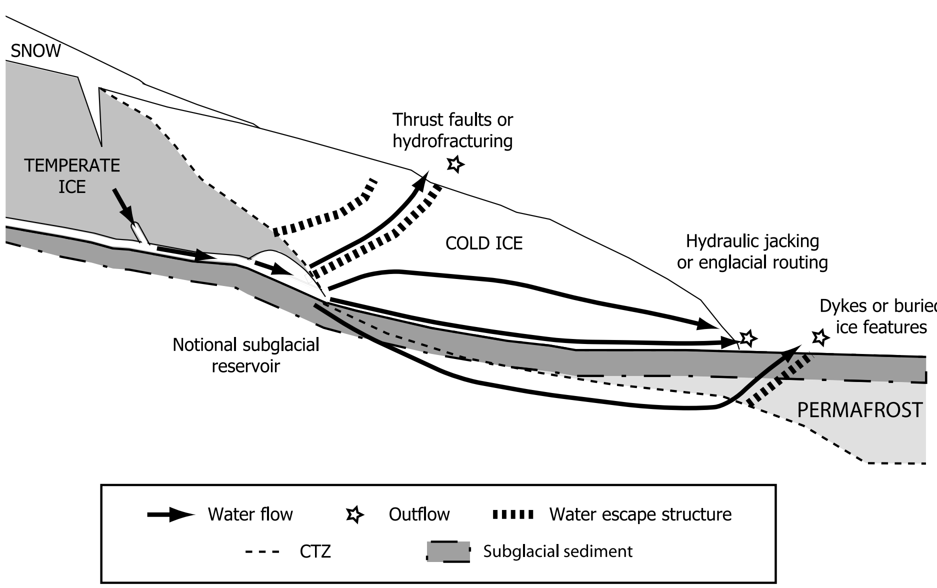

Figure 12. Potential outflow routes enabling water to pass from the temperate ice interior to breach the cold ice margin of polythermal glaciers.

instability, decoupling, and failure at the at nontemperate glacier's ice-bed interface [*Kavanaugh and Clarke*, 2001; *Murray and Clarke*, 1995]. Changes in the englacial system or the penetration of new hydrofractured crevasses to the glacier bed (described in section 4.2) may result in transient meltwater pulses which can subsequently drive subglacial water pressure increases [*Flowers and Clarke*, 2000]. The effectiveness of hydraulic jacking depends on how fast water spreads out at the bed and rates of water input to (and drainage from) the bed [*Röthlisberger and Lang*, 1987]:

$$P_{hj} = P_0 - \frac{\tau_b}{\tan(\alpha + \varphi)}, \tag{8}$$

where P_{hj} is water pressure necessary for jacking, P_0 is related to mean overburden pressure with respect to the average bed slope (α), τ_b is basal shear stress (rigid body, friction free model), and φ is the slope of the water reservoir or bed cavity. *Rippin et al.* [2005a, 2005b] suggested that hydraulic jacking may explain reduced basal drag and enhanced ice velocities observed at polythermal Midtre Lovénbreen, Svalbard; however, this was unsubstantiated becuase the dates of observation did not include the time prior to and during the commencement of subglacial upwelling when water pressures would likely be highest. Unless hydraulic jacking is spatially limited relative to the size of the glacier, the process could lead to a dynamically unstable ice mass. *Kavanaugh and Clarke* [2001] demonstrated that hydraulic decoupling at the ice-bed interface at Trapridge Glacier extended only over an area ~2×10^4 m^2, and observations interpreted to show hydraulic jacking do not suggest larger-scale basal instability at polythermal glaciers [e.g., *Bingham et al.*, 2006; *Rippin et al.*, 2005a, 2005b]. Rather, it seems that following brief instability, a high-capacity drainage route becomes defined through the cold margin or interfacial sediments and subsequently water pressures drop, leading to a more stable situation [*Kavanaugh and Clarke*, 2001; *Rippin et al.*, 2005b]. *Bingham et al.* [2006, 2008], in comparing hydrological and ice dynamical data, indicated a possible downglacier propagation of the pressurized water reservoir suggestive of a temporally progressive hydraulic jacking mechanism allowing gradual and spatially discrete drainage system development beneath a cold ice margin. *Murray and Clarke*'s [1995] observations at Trapridge Glacier indicate hydraulic jacking may be complex with pressure-driven uplift not necessarily occurring uniformly over space. The applicability and detail of this process therefore remain unclear with only limited data available, and the role of features such small-scale basal hydrofractures may be underestimated [e.g., *Mavlyudov*, 2005a].

[45] In some cases, subglacial waters appear to escape through permafrost in the glacier forefield rather than at the ice margin. Permafrost has a spatially variable hydraulic conductivity and can therefore influence both the location and timing of water emergence near the glacier [e.g., *Wainstein et al.*, 2008]. The immediate glacier forefield can exhibit perennially frozen, partially frozen, or unfrozen (talik) zones, depending on the sediment or rock types, the ice/water and the solute content, changes in the ice limit, and the groundwater flow history [*French*, 1996; *Haldorsen et al.*, 1996; *Haldorsen and Heim*, 1999]. Recent glacier retreat ensures that the depth of permafrost is still relatively shallow at the glacier margin and aggradation is likely to be both lagged after ice retreat and spatially discontinuous [e.g., *French*, 1996; *Kniesel*, 2003]. Further, structural weaknesses in frozen sediments, especially if previously deformed by glacier motion, can be exploited by subglacial waters during the seasonal outburst events [e.g., *van der Meer et al.*, 1999]. *Boulton and Caban* [1995] suggested that hydrofracturing and fluidization of forefield sediments or rocks can also allow water outflow, particularly at larger-scale polythermal ice sheet margins. Fluidization of sediments under high water pressure presents a mechanism by which the route becomes closed off once water pressures decrease later in the melt season, in that it leaves distinctive geological deposits termed *dykes* [*Boulton and Caban*, 1995]. However, it remains unclear how this process may operate on an annual basis at smaller valley glaciers, with reduced meltwater volumes (and therefore hydraulic pressures). In explaining upwellings in front of smaller retreating glaciers, *Hambrey* [1984a] argued that rather than being dependent on sedimentary weaknesses or processes, meltwater drainage may simply exploit hydraulic links between buried, stagnant ice and englacial and/or subglacial drainage flowpaths in the active glacier body (see also section 5). Therefore the occurrence of proglacial upwellings does not necessitate flowpaths through sediments but rather may link to englacial or subglacial cut and closure channels or similar relict structures hidden beneath melt-out *diamict* and tills.

[46] The configuration and progression of drainage in nontemperate glaciers will undoubtedly have influence on the seasonal hydrograph form, as well as on sediment and solute yields from individual glacier catchments. *Wadham et al.* [1998] reported that dissolved solutes emerging from Finsterwalderbreen, Svalbard, indicated a seasonal decline in water residence time and increasing diurnal cycles, implying the development of a more responsive and efficient subglacial drainage structure over the course of the ablation season. However, *Vatne et al.* [1996] found no hydrochemical evidence of drainage evolution at polythermal Hannabreen, Svalbard, but rather the potential coexistence of a fast (perhaps englacial) and slow (subglacial) meltwater drainage structures. Seasonal exhaustion of suspended sediment to subglacial runoff has also been recorded at a number of polythermal glacier basins and thought to indicate a transition toward a fast, efficient subglacial drainage system [e.g., *Hodson and Ferguson*, 1999; *Irvine-Fynn et al.*, 2005c]. In contrast, *Vatne et al.* [1992, 1995] and *Hodson and Ferguson* [1999] have reported continued increases in suspended subglacial sediment from some polythermal glaciers, interpreting the change to be driven by increasing volumes of water being delivered to the subglacial environment, at least partially controlled by individual thermal structure, rather than a simple process of subglacial drainage system evolution.

[47] Dye tracing has also been used to allude to properties of hydrological pathways in nontemperate glaciers and confirming supraglacial connection to seasonal outflows [e.g., *Bingham et al.*, 2005; *Irvine-Fynn et al.*, 2005a; *Vatne et al.*, 1995; *Vatne*, 2001]. Results from traces at John Evans Glacier over the course of 2 years showed a distinct change in descriptive indices during the drainage system bridging the formation and development of the seasonal upwelling [*Bingham et al.*, 2005]. Moreover, this change in drainage efficiency also appeared to show an up-glacier progression as the season advanced, as seen at temperate glaciers (Figure 4). However, both *Vatne et al.* [1995] and *Bingham et al.* [2005] attest to the annual variability in throughflow velocities attributed to seasonal contrasts in meltwater availability and runoff history. Outflow from the internal drainage system may be significantly modulated by cold margin restriction, which is unique for individual years and conditions. Interestingly, once subglacial upwelling had been initiated, *Bingham et al.* [2005] report a strong dependency between throughflow velocity and Q. Such a dependency suggests that once an efficient drainage system was initiated by upwelling, there was evidence for system stability. However, to date, there is not a wealth of data available to assess seasonal changes in internal drainage systems. While the form of dye traces may reveal variations in flowpath configuration [e.g., *Vatne et al.*, 1995], assessments of hydrograph form may also provide information on drainage structure [e.g., *Hooke et al.*, 1988]; however, with the complexity of supraglacial drainage revealed in section 4.1, it is not known how effective these analyses may prove to be.

[48] As an intriguing aside, during 2005 at Midtre Lovénbreen, the emergence of seasonal outflow was not limited to a single site, with two upwellings apparent close to the glacier margin but separated by a distance of ~500 m [*Irvine-Fynn et al.*, 2005a; *Irvine-Fynn*, 2008]. Multiple upwellings are not uncommon at polythermal glaciers, particularly at larger ice masses [e.g., *Schroeder*, 1998a]. However, at Midtre Lovénbreen, dye tracing showed flowpath divergence: A total of 11 singular dye injections from six different moulins across the glacier surface were all recovered at both upwelling sites over a 16 day period [*Irvine-Fynn*, 2008]. The dye travel time to the two upwellings exhibited a difference of 1.5 h, reflecting the differing hydraulic conditions of flowpaths, perhaps connected to a multichannel or fracture network in the active glacier interior, and contrasting to the anticipated arborescent drainage pattern predicted by hydraulic potential [*Rippin et al.*, 2003]. Such artificial tracer investigations, as with sediment or solute tracers, reassert that the assumption of a temperate glacier hydrology model for the outflow of meltwater may not be valid for nontemperate ice masses.

4.4.2. Winter

[49] Assuming the hydrological outlet is frozen closed during winter months, and either reopened or abandoned in favor of an alternative location during the following ablation season, the form of drainage structures, as discussed in sections 4.1–4.3, suggest the potential for water storage within the glacier's drainage system. *Schroeder* [1998a] discusses how, as a glacier becomes near stationary in winter, the blocking of drainage exits cause upstream retention of water in macrovoids (e.g., moulins), microvoids (e.g., foliations), and interstitial spaces. During winter, a flooded vadose channel may store water volumes of ~10^3 m^3 or more [*Schroeder*, 1998a]. Observations of a mapped englacial flowpath at Austre Brøggerbreen in 1998 and 2000 showed that a water volume of ~8×10^3 m^3 was retained in a single englacial channel, while at Hansbreen, *Benn et al.* [2009a] estimated annual water storage volumes of 1.3×10^6 m^3 in englacial conduits. However, the total water storage in channel forms will be dependent on the density and dimensions of channels, glacier size, and the rate at which summer season outflow is curtailed. Investigations of some englacial channels during the winter season have simply reported liquid water within channel pools, sometimes with thin ice lids rather than fully inundated conduits [e.g., *Griselin*, 1992; *Vatne*, 2001].

[50] *Irvine-Fynn et al.* [2006], having observed temporal variations in the interstitial water content of the temperate ice zone at Stagnation Glacier, presented a simple conceptual model of the winter increases in water pressures and volumes in a polythermal ablation zone location (Figure 13). The numerical model presented by *Aschwanden and Blatter* [2005] shows that interstitial production of meltwater to ice volume by strain heating alone can reach up to 10 g kg^{-1}, particularly at depth. The apparent englacial water table is not restricted to the temperate ice zone, however, with observations of water levels in moulins remaining at >60% of the ice depth compared to the local depth of temperate ice, which is significantly less than 50% of the ice thickness [cf. *Benn et al.*, 2009a; *Pälli et al.*, 2003; *Schroeder*, 1998a, 1998b, 2007]. *Irvine-Fynn et al.* [2006] termed this combined internal storage potential as the *specific retention capacity* of the glacier. Throughout winter, interstitial water draining under gravity from within the temperate zone, as well as water originating from within the open channels or supplied by liquid precipitation events, can increase down-glacier water pressure in the closed or plugged drainage system. As noted above, when pressures periodically exceed a notional threshold value, short-term outflow can occur, which can have important implications on flowpath closure rates. Observations of winter water levels in englacial channels have revealed water level fluctuations of up to 10 cm h^{-1} and numerous ice "rings" (3 cm thick) in a deep moulin shaft suggestive of progressive but irregular variations in water table levels of up to 50% of the glacier thickness [*Schroeder*, 1998b]. These variations reflect sporadic changes in the volumes of water retained within the glacier. Where, in response to ice dynamics or purely a function of thermal regime, water flow is continuous and/or sufficient to offset the processes responsible for drainage system closure, the winter outflow may be more continuous.

[51] Cyclic or continuous water outflow during the winter months at land-terminating glaciers may prompt the genesis of proglacial *icing* (also referred to in the literature as *naled* or *aufeis*). Icings develop from seepage, effusion, or periodic outflow of glacial meltwater that subsequently freezes,

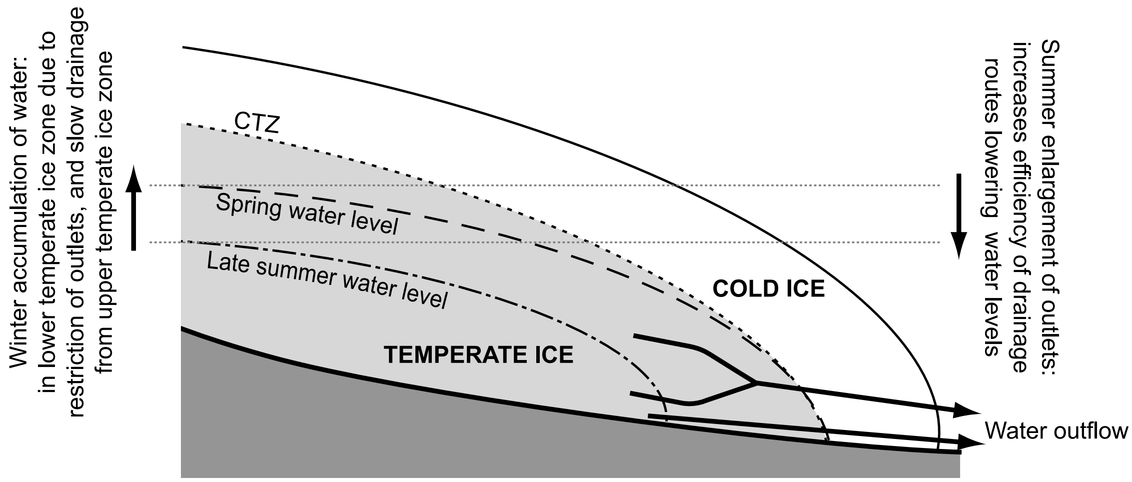

Figure 13. The conceptual model of specific storage capacity (redrawn from *Irvine-Fynn et al.* [2006]). Reprinted with permission from John Wiley.

often in topographic depressions, during the winter months [*Åkerman*, 1982; *Carey*, 1973; *Gokhman*, 1987; *Liestøl*, 1977]. Laboratory experiments and field observations of icings have revealed that laminated structures are common and result from cyclical accretion [*Moorman and Michel*, 2000a; *Schohl and Ettema*, 1986, 1990; *Wadham et al.*, 2000]. However, the structures of icings appear to be more intricate than can be explained by a simple model, and it is suggested icing grows in a series of stages with complex and ever changing hydrological flow patterns [*Moorman and Michel*, 2000a; *Wadham et al.*, 2000].

[52] *Bukowska-Jania and Szafraniec* [2005] reported that during the summer of 1990, 217 proglacial icings remained observable in Svalbard and exhibited a distinct inverse relationship between glacier size and icing to glacier area ratio: 0.5% to 0.1% for glaciers of <10 km^2 to >50 km^2, respectively. The icings formed within the limits of the recent (Neoglacial) maximum glacier extents, suggesting a link to active glacier hydrology [*Bukowska-Jania and Szafraniec*, 2005]. However, icing area may be misleading, as topography has a strong control on their extent and thickness, and the seasonal or interseasonal longevity of icings appears related to proximity to glacier ice where they are thickest [*Bukowska-Jania and Szafraniec*, 2005]. The size and depth of icings can be used to estimate the winter season outflow rate: Measurements of icings at Werenskioldbreen, Dronbreen, and Austre Lovénbreen (all in Svalbard) revealed outflow of 0.06, 0.03, and 0.005 m^3 s^{-1}, respectively [*Baranowski*, 1982; *Griselin et al.*, 1994; *Hansen*, 2001], emphasizing the large volumes of water retained after the summer season in certain nontemperate glaciers. A complete review of icings in Svalbard is given by *Gokhman* [1987].

[53] Because icing formation demands effusion of water during winter months, it is thought to be indicative of glaciers exhibiting a polythermal regime experiencing continued or periodic drainage from the temperate zone [*Åkerman*, 1982; *Baranowski*, 1977; *Hambrey*, 1984a; *Liestøl*, 1977, 1988; *MacKay and Løken*, 1974]. However, observations at Scott Turnerbreen suggested that cold or dominantly cold nontemperate glaciers (Figures 1b and 1f) may exhibit ice-marginal, subsurface water storage during the winter months, the drainage of which may form into icings [*Hodgkins*, 2001; *Hodgkins et al.*, 2004]. Consequently, it is unclear whether icings can be viewed as symptomatic of a polythermal regime.

4.4.3. Annual Water Storage and Release

[54] Seasonal peak meltwater discharge from nontemperate glaciers usually occurs during the initial snowmelt period, although the runoff magnitude does not necessarily relate to any meteorological forcing due to the storage processes at the ice surface. In Svalbard, the majority of snowmelt often occurs within a 14 day period from the start of the melt season [e.g., *Bruland et al.*, 2001], although this is clearly variable with catchment elevation and location. Following the demise of the snowpack, as the snowline retreats up-glacier to the equilibrium line altitude (ELA), ice melt commences with discharge hydrographs conditioned by prevailing weather patterns. At High Arctic and continental locations, where nontemperate regimes are common, incident radiation fluxes may account for >70% of seasonal ablation [see *Hodson et al.*, 2005; *Willis et al.*, 2002, Table 3]. It is important to note that as *Hodson et al.* [2005] showed for one glacier, <40% of the summer meltwater volume may be accounted for by ablation of glacier ice; the remainder is explained by snowmelt and summer precipitation. However, comparable figures are not widely available for other polythermal ice masses.

[55] As highlighted throughout section 4, the relative density of structures carrying meltwater into the glacier interior may be dictated by a combination of thermal regime and glacier dynamics. Both chemical and terrain analyses have suggested that on polythermal valley glaciers only a lesser proportion of the supraglacial waters may travel through the glacier interior; values of between 10% and 36% of the total meltwater flux have been quoted [e.g., *Griselin et al.*, 1994; *Hodson et al.*, 2005] and 25% to ~50% of the surface area [e.g., *Flowers and Clarke*, 2000; *Irvine-Fynn*, 2008]. Estimated using tracing techniques and assuming

straight-line or glacier-parallel travel distances, throughflow velocities at a number of nontemperate glaciers have been consistently shown to be of the order of 0.1–0.6 m s^{-1}, although it is noted this may be an underestimate, as sinuosity of channels and complexity of englacial or subglacial drainage is ignored [*Bingham et al.*, 2005; *Irvine-Fynn et al.*, 2005a; *Vatne et al.*, 1995]. The velocities are comparable to similar data at temperate glaciers, and the wide range of dispersion indices reported for traces conducted at nontemperate glaciers is likely to reflect the differences in flowpaths both in time and between sites (see section 4.4.1). A total of 88 dye tracing experiments conducted over 18 years at Austre Brøggerbreen suggest variations in input discharge dominate over system change for a cut and closure type channel [*Holtermann*, 2007; *Vatne*, 2001; G. Vatne, manuscript in preparation, 2011]. Dye experiments at moulins on Midtre Lovénbreen have shown a strong association between tracer velocity and system volume (the product of water flux and velocity), which is suggestive of open channel flow, as may be expected for vadose channels, and show similarity between englacial/subglacial flowpaths and supraglacial streams [*Irvine-Fynn*, 2008]. Drainage structures formed by processes outlined in section 4.2 may be readily able to cope with dynamic discharge without necessitating flowpath change. Importantly, the relatively limited dye trace data ensures that as discussed in above, it is still unclear whether changes in throughflow velocities and dispersion relate to input discharges or internal system configuration and/or change thereof; this remains to be explored further.

[56] Despite the apparently rapid throughflow in nontemperate glaciers, using a water budget approach by modeling melt and comparing to proglacial discharge, several studies have highlighted seasonal water storage and release within glacierized catchments in Svalbard. At Erikbreen, during 1990, total summer discharge was estimated to be 10% greater than volumes of ablation, while in the following year, runoff was only 93% of the summer melt [*Vatne et al.*, 1992]. This apparent specific water storage and release was attributed to superimposed ice formation or water storage within the internal drainage system, although *Vatne et al.* [1992] acknowledge uncertainties surrounding their calculated water volumes. Similarly, *Hodson et al.* [2005] used a 5 year time series at Midtre Lovénbreen to show that annually, specific storage ranged between −0.39 and +0.68 m, compared to specific runoff values of from +1.1 to +1.5 m. These results implied that the differences in storage can be explained by contrasts in progression of sequential ablation seasons. The energy (or melt volumes) required to establish a surface-to-forefield drainage system may be dependent on the water volumes stored or released from the previous winter and/or internal changes necessary to enable the hydrological connections [*Hodson et al.*, 2005]. Exploring this notion further using daily water budgets at the same glacier, during the melt season of 2005 a peak specific storage of ~0.11 m was found with evidence of both supraglacial and subglacial storage processes; this water volume was subsequently released during the same ablation season [*Irvine-Fynn*, 2008]. The presence of antecedent structures as locations for water storage may provide an important mechanism by which to explain interannual variations in volumes of meltwater storage, in addition to the reservoirs represented by the development icings and supraglacial refreezing processes. Furthermore, as noted in section 4.4.1, these variations in stored water volumes in such a drainage system may have an impact upon the progression, timing, and possibly the location of seasonal outflow [*Bingham*, 2003; *Bingham et al.*, 2006; *Hodson et al.*, 2005].

5. GLACIER-PERMAFROST INTERCONNECTIVITY

[57] The nontemperate environment also promotes peculiarities for drainage processes in the form of permafrost and buried ice. The cold wave penetration is not limited to the thermal regime of glaciers but also includes the accretion and maintenance of permafrost in the ice proximal terrain, and limits the development of the seasonally thawed active layer to ~1 m depth [e.g., *Humlum et al.*, 2003; *Moorman and Michel*, 2000b]. As a result of these periglacial conditions, runoff is highly interdependent on precipitation and melt rates [*French*, 1996; *Hodson et al.*, 1998]. With regions of nontemperate glaciers broadly exhibiting recent retreat, many of the residual moraine ridges now exposed and degrading close to the ice margins and in proglacial forefields have been found to be ice cored, and instances of buried icings have also been noted [e.g., *Brandt et al.*, 2007; *Etzelmüller et al.*, 1996; *Evans and England*, 1992; *Gibas et al.*, 2005; *Hambrey*, 1984a; *Hoelzle*, 1993; *Kozarski*, 1982; *Lønne and Lyså*, 2005; *Lukas et al.*, 2005; *Lyså and Lønne*, 2001; *Moorman and Michel*, 2000b; *Sletten et al.*, 2001; *Ziaja*, 2005]. Glacier ice may also be buried by sediment deposition from significant outflow events, for example, the draining of an elevated, marginal ice-dammed lake [e.g., *Moorman and Michel*, 2003]. The resulting ridges from these processes and buried ice forms reflect glacier downwasting from previous extents, either since recent Neoglacial maxima or from more contemporary surge cycles, and the melt-out of entrained debris. More detailed conceptual models of ice-cored moraine formation and preservation are presented elsewhere [e.g., *Etzelmüller et al.*, 1996; *Lønne and Lyså*, 2005; *Paul and Eyles*, 1990; *Sletten et al.*, 2001].

[58] *Etzelmüller* [2000] presented results from proglacial regions in Svalbard, finding a significant reduction in surface elevation over the course of a 20-year period inferred to be thermoerosion ice-cored features. Such erosion may contribute significant quantities of debris for hydrological transport in Arctic glacierized catchments [*Etzelmüller*, 2000; *Etzelmüller et al.*, 2000; *Lyså and Lønne*, 2001]. Terrain analyses and resistivity sounding have indicated that buried ice in nontemperate glacier forefields may be quite widespread, not only confined to moraine ridges [*Hoelzle*, 1993; *Irvine-Fynn et al.*, 2010; *Irvine-Fynn et al.*, 2011; *Lukas et al.*, 2005; *Schomacker and Kjær*, 2008]. The viscous dissipation of heat from flowing meltwater [e.g., *Hodson and Ferguson*, 1999; *Irvine-Fynn et al.*, 2005b] and melt-induced thaw slumps [e.g., *Bennett et al.*, 2000; *Lukas et al.*, 2005] may result in the rapid thermoerosion of

ice-cored moraine complexes. As was discussed in section 4.4, *Hambrey* [1984a] suggested that despite active glacier margins appearing well defined, the proglacial zone may exhibit extensive buried *dead ice* that can remain hydraulically linked to englacial or subglacial drainage systems. *Moorman and Michel* [2000a] document an englacial channel in the active Stagnation Glacier, Canadian Arctic, which was fed by a flowpath originating in the lateral ice-cored moraine ridge. *Moorman* [2005] details the change in hydrological linkages at Stagnation Glacier over a decade as the glacier thinned and the ice-cored moraine degraded—a process accelerated by hydrological processes—and suggests that other flowpaths preserved in stagnant (cold) ice can become reactivated. *Ziaja* [2005] also reported reshaping of the proglacial environment by the collapse of former glacial channels at sites in central Svalbard. The development of drainage networks in active nontemperate glaciers appears to play a potentially significant a role in the longer-term catchment hydrology, even following glacier retreat.

6. TOWARD UNDERSTANDING ICE SHEET HYDROLOGY

[59] Contemporary ice sheets are known to display nontemperate (if not polythermal) margins and outlet glaciers [e.g., *Baral et al.*, 2001; *Breuer et al.*, 2006; *Funk et al.*, 1994; *Greve*, 1997, 2005; *Huybrechts*, 1992], a thermal structure largely inherited from paleoclimatic history [*Rogozhina et al.*, 2011]. Though our review focuses on nontemperate valley glaciers, a number of hydrological corollaries exist within ice sheet environments. This section aims to provide only a brief illustration of these overlapping topics and their recent advances, rather than an exhaustive summary, since ice sheet hydrology remains a fledgling subject area. Having said this, we note that the number of process studies specifically focusing on the polar ice sheets has increased significantly in recent years as a direct result of the perceived role that hydrology plays as a positive feedback on ice sheet dynamics under a warming climate [*Lemke et al.*, 2007; *Meehl et al.*, 2007]. Critically, there is no specific reason per se that ice sheets should behave significantly differently from nontemperate valley glaciers; the processes of meltwater generation, refreezing, routing, storage, and dynamic coupling are the same albeit significantly upscaled in magnitude and space.

[60] In Antarctica, surface melt largely occurs during relatively warm summer periods upon ice shelves and the coastal margins of the ice sheets, where it is strongly influenced by surface topography. The majority of snowmelt refreezes within the snowpack and so makes little contribution to runoff as evident in ice layer structures [*Das and Alley*, 2005; *Tranter et al.*, 2010]. Where ice is exposed, such as in blue ice areas, subsurface melting to depths of ~1.5 m is possible through preferential absorption of incident radiation, coarsened crystallography, and greenhouse-type conditions [e.g., *Bøggild et al.*, 1995; *Liston and Winther*, 2005]. Similarly, dust and cryoconite accentuate melt and hydrological development [*Bagshaw et al.*, 2010; *Fountain et al.*, 2008; *MacDonell and Fitzsimons*, 2008], directly contributing toward ~13% of the seasonal melt volumes in the case of the valley glaciers in the McMurdo Dry Valleys [*Fountain et al.*, 2004]. These near-surface melt processes both form and accentuate "cryolakes" that have highly varied size, extending from 1 m to 2×10^3 m, although typically limited to shallow (<1 m) depths, and in turn furthering local melt genesis [*Bøggild et al.*, 1995; *Tranter et al.*, 2010; *Winther et al.*, 1996]. Because water readily refreezes in response to cold air and ice temperatures, the hydrology (including the lakes) becomes restricted to the near-surface weathering crust, with liquid water drainage functioning beneath a veneer of cold surface ice up to ~0.5 m in thickness [*Bagshaw et al.*, 2010; *Bøggild et al.*, 1995; *Winther et al.*, 1996]. This veneer can decay for brief periods during the summer months. The hydrological connections may form subaerial rills, and where drainage is concentrated into streams, these exhibit meandering, step-pool sequences and may extend for more than 1 km [*Fountain et al.*, 2004; *Tranter et al.*, 2010; *Winther et al.*, 1996].

[61] In Greenland, summer air temperatures tend to increase to values above freezing and thereby enhance the surface melt rates near the margins [e.g., *Hall et al.*, 2008]. However, meltwater retention [*Greuell*, 2000] and dust and cryoconite [*Bøggild*, 1997; *Bøggild et al.*, 2010] remain important to ablation, potentially increasing seasonal runoff volumes by ~50% [*Bøggild et al.*, 2006]. Supraglacial lakes forming during summer at the progressively exposed ice surface have long been observed [e.g., *Echelmeyer et al.*, 1991]. More recent observations of these lakes have included estimates of dimensions of typically $<10^3$ m diameter and <12 m deep [*Box and Ski*, 2007], although larger lakes have been recorded [e.g., *Echelmeyer et al.*, 1991; *Sundal et al.*, 2009], and many appear to exhibit spatial continuity, recurring annually in the same position [*Box and Ski*, 2007]. As the ablation season progresses, the lake depressions fill at rates of up to ~5 mm h^{-1} [*McMillan et al.*, 2007], accentuating ice ablation by up to 170% through changing the ice surface energy balance [*Lüthje et al.*, 2006]. Nonetheless, there is considerable intra-annual and interannual variation in ponded water volumes [*Sneed and Hamilton*, 2007]. A significant proportion of the supraglacial lakes examined by *McMillan et al.* [2007] and those by *Box and Ski* [2007] exhibited both increases and reductions in area as the summer season progressed. As has been observed for supraglacial lakes on nontemperate valley glaciers [e.g., *Boon and Sharp*, 2003], lakes on the Greenland Ice Sheet have been observed to drain through hydrofractures driven by water volume [*Das et al.*, 2008]. Model runs have been used to illustrate that lakes containing $>8 \times 10^4$ m^3 may be capable of driving hydrofractures through ~1 km of cold ice [*Krawczynski et al.*, 2009]. Using geophysical methods, *Catania et al.* [2008] have inferred data to suggest the penetration of fracture-like features to the bed of the Greenland Ice Sheet, and modeling has shown these fracture paths may be long-lived, persisting for several years [*Catania and Neumann*, 2010]. However, as *Krawczynski*

Figure 14. Large moulin at the surface of the Greenland Ice Sheet at the close of the ablation season (image courtesy of A. L. Hubbard). Note the step-pool sequences in the supraglacial channel.

et al. [2009] caution, smaller lakes may drain by hydrofracture but demand greater water volumes for the meltwaters to penetrate to the bed. To date, only *Das et al.* [2008] have provided unequivocal evidence for meltwaters directly reaching the ice bed through hydrofracture; therefore it remains unclear whether this form of drainage is the most significant in delivering meltwater to the ice sheet sole, particularly in light of the occurrence of large, persistent moulins (Figure 14) and the recognition of nonponded meltwater as a major contributor to total runoff volumes [e.g., *Sneed and Hamilton*, 2007].

[62] Despite the uncertainty surrounding the mechanisms by which surface-derived meltwaters reach the bed of the Greenland Ice Sheet, the linkage has been inferred by seasonal sediment plumes [e.g., *Chu et al.*, 2009; *McGrath et al.*, 2010] and ice dynamics [e.g., *Bartholomew et al.*, 2010; *Palmer et al.*, 2011; *Shepherd et al.*, 2009; *Zwally et al.*, 2002]. *Shepherd et al.* [2009], using high-temporal-resolution geodetic measurements of ice surface displacement during the ablation season, note a strong diurnal signal in both horizontal and vertical surface displacement at a number of locations stretching up to 80 km within the western margin of the Greenland Ice Sheet; because the vertical uplift across all sites was in-phase, though slightly retarded behind the daily surface melt cycle, they argued the diurnal horizontal acceleration and uplift (up to 0.15 m) were directly driven by surface-to-bed meltwater coupling. For ice within 35 km of the western ice sheet margin, ice motion of one outlet glacier was linked to progressive meltwater fluxes delivered to and draining from the ice bed [*Bartholomew et al.*, 2010]. Indeed, the remote sensing–based investigation of *Sundal et al.* [2011] specifically argues that the behavior of the land terminating glaciers of the southwestern margin of the Greenland Ice Sheet "mirrors that of mountain glaciers": In years of high melt, although peak summer surface velocities are of a larger magnitude than corresponding moderate or low-melt years, this is more than offset in that the duration of the summer speedup period in these high-melt years is significantly shorter, resulting in an overall mean reduction in flow velocity during them. Drainage system development and evolution stabilizes efficiency of meltwater outflow, reducing basal effective pressures and a concomitant slowdown in overall flow velocity [*Sundal et al.*, 2011]. Observations of decreasing annual ice velocities over 17 years have been interpreted as the hydrological system compensating for variability in meltwater input and negating accelerated ice flow [*van de Wal et al.*, 2008]. Indeed, *Russell* [1993] hypothesized a stable, retained, high-capacity drainage system to explain anomalous supraglacial lake drainage events during winter months at the Greenland ice margins. As *Howat et al.* [2010] demonstrated, supraglacial drainage events linked to the ice sheet bed may even appear to reduce outlet glacier motion, even for marine-terminating glaciers, by increasing hydrological system efficiency. This at least appears to be the case at the marginal 35 km of the Greenland Ice Sheet on which many studies focus, where ice thicknesses are still relatively thin and hydrostatic stresses and flowpath closure rates are going to be retarded compared to up-glacier zones of the ice sheet.

[63] In the ice sheet interior of Greenland, the spatially discontinuous presence of subglacial water has been conjectured from geophysical surveys [*Oswald and Gogineni*,

2008]. As such, and assuming the absence of surface-to-bed linkage, subglacial hydrology may be analogous to Antarctica where ~12×10^3 km^3 of water is stored in >70 subglacial lakes that exist in areas of low basal topographic relief, broadly matching the distribution of pressure melting [*Dowdeswell and Siegert*, 1999; *Dowdeswell and Siegert*, 2003]. Areas of subglacial water storage may result not only from ice thickness but also from geothermal heat and tectonics [*Bell et al.*, 2006; *Fox-Maule et al.*, 2005]. Recent work has demonstrated that these subglacial lakes are linked [*Fricker et al.*, 2007; *Fricker and Scambos*, 2009; *Fricker et al.*, 2010; *Gray et al.*, 2005; *Smith et al.*, 2009; *Wingham et al.*, 2006] and may exhibit continuous or periodic connections, as neatly summarized by *Clarke* [2006]. It is these basal connections that, through hydrologic-dynamic coupling, have been shown to influence ice flow velocities at Antarctic margins [e.g., *Bindschadler and Choi*, 2007; *Stearns et al.*, 2008]. Despite the notion that discrete, spatially dispersed water delivery is likely to have the greatest influence on ice sheet dynamics [*Bell*, 2008], we have yet to discern the exact mechanisms and conditions of hydraulic architecture at the bed of ice sheets, despite the similarity (albeit at different spatial scales and in the absence of surface derived water) to the slow drainage systems detailed at valley glaciers [e.g., *Kamb*, 1987]. However, one should not degrade the relative importance of environmental conditions such as ice shelves, calving conditions, along-flow stress transfers, and adverse bed slopes to ice sheet mass dynamics.

7. SUMMARY AND IMPLICATIONS

[64] The purpose of this article, as stated at the outset, was to provide a summary of the recent advances in understanding of glacier hydrology with specific focus on ice masses characterized by nontemperate thermal regimes. In detailing the processes by which a glacier's thermal regime is conditioned, it is clear there is a wide range and variability in the distribution of cold and temperate ice. *Hodson et al.* [1997] suggested using a combination of numeric indices to describe the thermal regime of a glacier: the proportion of the bed along the flow or centerline at PMP and the proportion of temperate ice over the ice thickness at the ELA. Such indices enable comparisons between nontemperate glaciers to be drawn in a more meaningful way than is provided by qualitative descriptions. However, such an approach limits classification to glaciers with geophysical data, which itself has some uncertainty in the confident delineation of thermal structures. Rather, we suggest building upon the strategy used by *Blatter and Hutter* [1991] and *Etzelmüller and Hagen* [2005]: to use the categories illustrated in Figure 1 as a universal classification scheme of thermal structure, thereby avoiding nominally subjective descriptions. Here we assert the importance of using the term polythermal only to refer to glaciers exhibiting large proportions of their bed with the contrasting ice types and water content. The hydrology and dynamics of such polythermal glaciers is undoubtedly linked to their thermal regime, including configurations dictated by past climates or dynamics.

7.1. Nontemperate Glacier Hydrology

[65] There is a growing awareness of the limitations of the macroscale, conceptual model of temperate ice hydrology, with the emerging importance of microscale, structural controls on flowpath development and configuration. The importance of these structural and thermal controls on hydrology is better illustrated in nontemperate glaciers. The limited depth of the seasonally temperate ice surface layer and reduced ablation rates ensure characteristically marked supraglacial drainage behavior and structural form. The underlying cold ice, and commonly associated low surface gradients, results in formation of superimposed ice at the base of the snowpack and water storage within a glacier's snowpack or in melt lakes at the commencement of the ablation season. Subsequently, as the ice is progressively exposed, the formation of a deep weathering crust appears to have a significant influence on supraglacial runoff, with debris and cryoconite influencing the inception of surface rills, and thermal erosion at the banks of larger supraglacial channels widening the channel form, and aiding in their persistence year to year. Clear analogies with hillslope hydrology can be drawn, as noted by *Wakahama et al.* [1973], in considering the weathering crust as a porous media with a depth-limited storage capacity and actively dynamic water table. The underlying cold ice with lowered deformation rates allows the larger supraglacial channels to thermally incise to great depths with meandering patterns and step-pool morphology that may or may not be reflected in the ice surface expression of the flowpath. However, the details of hydraulics and the incision process and rates currently remain inadequate.

[66] Deep incision of supraglacial channels through cold ice coupled with ice deformation and/or snow bridge consolidation lead to cut and closure vadose englacial channels. Englacial hydrological connections can also be formed by processes of the burial and advection of channels formed in accumulation area firn and snowpack or through the hydraulic exploitation of existing fractures and other structural features with greater permeability. An alternative process of englacial flowpath inception is hydrofracture, which may extend crevasses to greater depths from the surface, be prompted by supraglacial water storage, or exploit inherent structural weaknesses within or at the bed of the glacier. The thermal regime, associated ice dynamics and fracture density, and the supraglacial hydrology are important to these processes becuase they demand continued meltwater delivery. In ice masses exhibiting an absence of crevasses, discrete cut and closure channels are most likely, with isolated incidence of hydrofracture in propitious circumstances; for those with moderate crevassing, combinations of vadose and fracture flowpaths prevail; where crevasse density is high, englacial flow paths may be absent due to the reduced source areas feeding potential drainage structures. Nonetheless, in nearly all instances, it seems that englacial flowpaths become seasonally constricted by freezeup of their outlet during winter

months, but the residual englacial structures remain open or water-filled, providing internal meltwater reservoirs, especially where ice creep closure and refreezing rates are low. An area of uncertainty remains over the manner in which the drainage structures through cold ice develop once in contact with temperate ice zones. However, the englacial drainage structures may readily evolve into subglacial systems or deliver meltwater inputs to generate or sustain subglacial drainage, and it appears englacial channels may provide flowpaths to enable water to breach the cold ice margin that typifies nontemperate glaciers. Corollaries between englacial hydrology and groundwater flow are apparent when considering fracture networks and discrete vadose channels delivering meltwater to depth and in viewing a nontemperate glacier body as an aquifer.

[67] It is clear that where proportions of the glacier bed are cold or frozen, the subglacial drainage system is reduced in terms of spatial extent compared to temperate bedded glaciers. Nonetheless, certainly in polythermal glaciers, subglacial drainage structures do exist and observations of nontemperate glacier dynamics showed seasonal and subseasonal variations in ice velocities not dissimilar to those recorded at temperate glaciers [*Andreason*, 1985; *Blatter and Kappenberger*, 1988; *Iken*, 1972; *Müller and Iken*, 1973; *Rabus and Echelmeyer*, 1997]; such velocities necessitate the presence of subglacial water and suggest that it is incorrect to assume a reduced surface-subsurface hydraulic coupling in polythermal glaciers. Moreover, it is likely that at the macroscale, subglacial drainage at polythermal glaciers, particularly those with extensive temperate ice areas, may exhibit similarities with the long-standing Shrevian model, perhaps forgiving modeling based upon glacier-scale hydraulic potential [e.g., *Copland and Sharp*, 2001; *Flowers and Clarke*, 1999; *Hagen et al.*, 2000; *Pälli et al.*, 2003; *Pattyn et al.*, 2009; *Rippin et al.*, 2003]. However, water delivery to the glacier bed in nontemperate examples may be more spatially discrete than at temperate counterparts, and the controls imposed by spatial variability in mesoscale and microscale thermal regime and water content should not be underestimated.

[68] At nontemperate glaciers the subglacial system (as with the englacial system) can be isolated during cold winter months. Water storage in such a system aids its longevity. However, it appears that winter season outflow of glacial meltwater can occur with a wide variety of potential mechanisms, from preserved flowpaths in buried glacier ice to open talik or groundwater flow to periodic breach of drainage structure constrictions by elevated water pressures. Critically, the exact pathways and driving forces behind perennial reinitiation of seasonal outflow remain poorly investigated, despite annual sediment and solute yields from nontemperate glacierized catchments broadly appearing to reflect the internal drainage system structure.

[69] The variability in thermal structure is likely to result in hydrological structures that are unique for individual glaciers, and a unified conceptual model is unlikely to be forthcoming. Nonetheless, continued attention to the structure, functionality. and dynamics of glacier drainage systems is necessary if the scientific goal sought is to predict dynamical responses of nontemperate ice masses and their impact upon sediment and solute fluxes to sensitive downstream ecosystems.

7.2. Ice Sheet Hydrology?

[70] A wide variety of authors discuss the likelihood of polythermal conditions in the large ice sheets, not only in the contemporary ice sheets but also during past ice ages as evidenced in paleoglaciology, geomorphology, and modeling [e.g., *Heine and McTigue*, 1996; *Kleman and Hättestrand*, 1999; *Kleman and Glasser*, 2007; *Marshall and Clark*, 2002; *Rolandone et al.*, 2003]. Nontemperate valley glaciers are therefore more reflective of ice sheet boundary conditions than their entirely temperate cousins and may be more readily studied and constrained than the more complex ice sheet systems. As shown throughout section 6, distinct analogies between ice sheets and nontemperate valley glaciers are evident for supraglacial, englacial, and subglacial hydrologies. A number of recent studies specifically argue that an ice sheet should not behave any differently from a valley glacier and that the dynamically active, fast flowing zones of ice sheets, characterized by ice streams and outlet glaciers, have a similar response time (years rather than millennia) and, it has been argued, are direct analogs of their smaller counterparts. In the past decade, our understanding of hydrodynamical processes within the surface, englacial, and basal domains of ice sheets has gained massively due to innovations on two fronts: (1) remote sensing that has revealed macroscale insights into ice mass acceleration and loss and how these relate to hydrological processes operating across the ablation zone or at marine termini of polar ice sheets and (2) field techniques that have enabled observations of ice sheet flow, ablation, and hydrological processes that previously could only be studied at valley glacier scale to be upscaled and investigated on the ice sheet scale. Field-based process studies do not come cheap, as the logistics and field technological costs spiral rapidly over the greater spatial scales demanded for ice sheets. However, such studies are invaluable in indicating that the general processes observed from valley glaciers (be they temperate or polythermal) are generally sound and can probably be applied to larger domains. Critically, however, knowledge gained at outlet glaciers should be treated with caution and not extrapolated to be an assumed understanding of ice sheet hydrology: Such hydrology remains one of the key areas of uncertainty for modeling ice sheet behavior at all time scales [*Marshall*, 2005]. While the bulk of literature examining ice sheet hydrology focuses on ice dynamical coupling, it is worth recalling that *Clayton* [1964], *Clapperton* [1968], and *Mavlyudov* [2005b] have all suggested that the drainage structures formed in polythermal ice may, themselves, have been influential not only in landscape formation but also in the down-wasting of the ice sheet margins.

7.3. Future Research Directions

[71] In reviewing the hydrology of polythermal glaciers, five key research directions become apparent:

[72] 1. The transition of a glacier's thermal regime. The manner and rate at which the distribution of cold and temperate ice vary are imperative contexts within which to understand hydrological and ice dynamical processes. Not only are long-term changes poorly constrained but also interstitial water mobility appears more rapid than theoretically expected and demands consideration.

[73] 2. Understanding supraglacial hydrology. Models typically ascribe rapid runoff, and little work has been devoted to the processes of supraglacial drainage evolution. To date, the relative proportions of meltwater following purely supraglacial pathways, and long-term and seasonal temporal changes within contributing areas, have not been explored in depth. The analogies between surface hydrology observed in Antarctic environments and polythermal glaciers highlight that this area deserves renewed research attention: in particular, the hydrological role of the weathering crust, the thermodynamic contribution near-surface water and cryoconite deposits may have to meltwater genesis, and the hydraulics involved in supraglacial stream incision.

[74] 3. Ascertaining englacial connections. Further speleological approaches to exploring englacial hydrology [*Gulley*, 2009; *Gulley et al.*, 2009a; *Řehák et al.*, 1990; *Vatne*, 2001] are needed, although readers should note that speleology as a form of scientific glacial research to comprehend processes of englacial drainage is not a new idea, as had been formally proposed by *Halliday and Anderson* [1970] and discussed extensively by *Mavlyudov* [2006a]. Indeed, *Mavlyudov*'s [2006b] *Internal Drainage Systems of Glaciers* presents ideas and concepts, some of which remain to be tested or validated, but the book is in Russian and so not yet fully appreciated in the more dominant Western literature. Consideration of the englacial environment in terms of karst-like structures or that of a fracture network may prove fruitful in modeling meltwater throughflow. Nonetheless, more attention needs to be directed toward hydrological processes at the CTZ.

[75] 4. The subglacial drainage system. The mechanisms, pathways, and dynamics of drainage beneath polythermal ice masses remain unclear, and methods such as temporal sequences of ground-penetrating radar surveys to explore variation in hold potential. Microbial activity releases heat, albeit very small, and it remains unclear whether there are feedbacks between the subglacial ecosystem and thermal and/or hydrological functioning beneath nontemperate glaciers. Further, the exact nature, form, and dynamics of seasonal upwellings demand further exploration.

[76] 5. Furnishing realization of ice sheet hydrology. Focus upon the manner in which supraglacial waters are transferred within an ice sheet demands continued attention, particularly in terms of whether hydrology may reciprocally influence thermal regime, and so too for processes and timing of subglacial water redistribution. However, recognition of the current focus on marginal areas needs be placed in context. While it is clear that coupled field-based and remote sensing approaches to exploring ice sheet phenomena are yielding promising advances in knowledge, the potential in the advancement and use of existing ideas and observations drawn from nontemperate valley glaciers should not be overlooked in terms of driving the conceptual development and elucidation of a clearly complex problem.

7.4. Implications of Environmental Change

[77] Glaciers across the high latitudes where polythermal regimes are dominant have thinned or retreated in recent decades [e.g., *Nuth et al.*, 2007; *Sharp et al.*, 2011]. Trends and projections for the Arctic and similar environments suggest broadly rising temperatures and increasing precipitation but declining snow extents with lengthened summer seasons [*Christensen et al.*, 2007; *Hinzman et al.*, 2005; *Kattsov et al.*, 2005], all accentuating runoff. Glacier retreat is typically coupled with thinning, and the hydrology will reflect glacier thermal structural controls. In some situations, as cold ice becomes increasingly dominant for smaller glaciers, development and persistence of vadose channels or enlargement of existing structures are likely. The thermal change will also reduce the extent of any subglacial system. Ironically, the large cross-sectional areas of vadose drainage structures within cold ice have a carrying capacity capable accommodating higher summer meltwater discharges. However, the processes of incision may be accelerated and protracted and potentially greater volumes of meltwater may be delivered to the bed. The consequence of the delivery of meltwater to the bed of small thinning glaciers may be a change in the number of crevasses. For example, in the instance of the cooling glacier (Figure 2, left), although increasing cold ice extents would theoretically reduce the incidence of crevassing, a reduction in ice plasticity may result in more fracturing at confluences, instabilities where meltwater reaches the bed or where cirques join main glacier tongue as stresses increase. This increase in crevassing could impose a new fracture-based glacier hydrology with unknown consequences upon ice dynamics. Where glaciers become increasingly temperate (Figure 2, right), the likelihood is for increasing water delivery to the bed and potentially accelerating the transfer of ice to elevations of ablation. The dynamic response of nontemperate glacier is therefore intimately dependent on the hydrological structures and functioning within the ice mass. Crucially, large glaciers will take a longer time frame to respond to climatic change and the associated thermal and hydrological reactions will be delayed.

[78] The transfer of sediment and solutes from nontemperate glaciers subject to decline is likely to increase, at least initially. The release of subglacial sediments may be reduced as cold ice beds at small glaciers increase, but this may be offset by the unconsolidated sediment exposed at the ice margins, which can be entrained by periglacial and fluvial processes [e.g., *Lukas et al.*, 2005; *Porter et al.*, 2010], while instances where temperate ice increases so too will subglacial sediment release. Paleoenvironmental records do suggest greater fluxes of sediment during times of decreasing glacier extent [e.g., *Elverhøi et al.*, 1995] and models suggest that sediment yields will increase under a warming climate [*Syvitski*, 2002; *Warburton*, 2007]. The

increased precipitation and exposure of reactive subglacial sediments will, over short time scales, increase the fluxes of crustally derived solutes in a warming climate. However, the rates and fluxes invoked by these changes are not known.

7.5. Concluding Comments

[79] Although there have been a growing number of investigations exploring topics from larger-scale hydrological structures in nontemperate glaciers to the specifics of solute transfer, the majority of these have occurred on glaciers with long histories of glacial research. Undoubtedly, observations and data sets collated at a single site over a number of years, if not over decades, is vital to comprehend a glacier's hydrology and its internal dynamics in time and with respect to the driving force of climate. It would seem, however, that to better understand the spatial variation in hydrological behavior, and to test current conceptual models, there is a concurrent need to both continue with incremental research achievements at individual glaciers and broaden the number of field sites. Researchers need to endeavor to initiate new long-term monitoring programs on new or under-explored glacier sites. In all of the areas for future scientific endeavor, it is necessary to continue with field-based research but utilizing a combination of techniques including dye tracing, hydrological monitoring, hydrochemical analyses, melt modeling, speleology, ground-penetrating radar, and electrical resistivity tomography. As *Hodgkins* [1997] asserted, researchers still need to extend the volume of hydrological data gained at nontemperate glaciers, to emphasize the specificity of nontemperate hydrology to avoid inappropriate generalizations, and to expect ice dynamical modelers to both include and press for information that provides more accurate boundary conditions. The devotion to subglacial hydrology in recent glaciological research, due to the immediate connection to ice dynamics, has resulted in a poor recognition of supraglacial, near-surface, and englacial flowpaths, which may yet prove to be significant in the hydraulic and biogeochemical functioning of nontemperate glaciers, as well as in their dynamics. Through the upsurge in interest in polythermal glacier hydrology, coupled with the recent developments in data acquisition and theoretical developments, and similarities alluded to here, *Sharp et al.*'s [1998] observation that nontemperate glacier flowpaths hold distinct analogies to hillslope hydrology may yet be an intriguing line of future enquiry and glaciological progress.

GLOSSARY

Accumulation area ratio: the ratio of accumulation area at the end of the melt season to the total glacier area.

Cold-temperate ice transition zone (CTZ): the region of gradual transition from cold glacier ice (with no liquid water content) to warm or temperate ice at the pressure melting point (containing intracrystalline liquid water).

Cryoconite: granular, organic-rich, biologically active debris found upon glacier surfaces.

Dead ice: typically glacier ice now isolated from the active glacier body, characteristically covered with sediments and moraine diamict.

Diamict: unsorted glacial sediments, also termed till.

Dynamic thinning: glacier thinning and mass loss in excess of the imbalance between surface mass balance (climatically determined from annual melt and accumulation volumes) and ice mass replenishment by flow.

Glacier hypsometry: the distribution of glacier area over elevation.

Hydraulic jacking: uplift of an ice mass by exaggerated subglacial water pressures.

Hydrograph: the time series showing changes in the runoff (discharge) over a range of temporal scales.

Icing: sheet-like bodies of ice formed during winter by the refreezing of emergent discharge, typically filling topographic depressions close to glacier margins; also termed *naled* or *aufeis*.

Interstitial: for ice, the space between individual crystals, which may be water-filled.

Phreatic channels: channels that are typically circular in cross section and water-filled, usually where channel water is pressurized (i.e., at ice-overburden pressure).

Pressure melting point: the melting point of ice under pressure. There is an inverse relation between pressure and ice's melting point.

Primary permeability: a measure of the microscale ability of ice to transmit liquid water (for example, through interstitial void space).

Rills: narrow and shallow incisions carrying runoff at the surface.

Secondary permeability: a measure of the mesoscale to macroscale ability of ice to transmit liquid water (for example, through crevasse structures).

Superimposed ice: refrozen snow or ice melt at the glacier ice surface.

Surface mass balance: the difference between a glacier's accumulation and ablation, excluding dynamic changes and calving.

Talik: a layer or zone of perennially unfrozen ground within permafrost.

Thrust faults: where deformation of ice result in pressures causing ice layers being pushed up over other zones of ice; the resultant interface is the fault.

Upwelling: seasonal emergence of waters at locations within a glacier catchment, often isolated from existing fluvial and/or drainage features.

Vadose channels: channels that are deep relative to their width and canyon-like in shape because vertical incision dominates. These usually occur where water pressures are atmospheric.

Weathering crust: a shallow layer of porous ice at the glacier surface produced by differential melt rates.

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