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Theoretical, contemporary observational and palaeo perspectives on ice sheet hydrology: processes and products

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Abstract

Meltwater drainage through ice sheets has recently been a key focus of glaciological research due to its influence on the dynamics of ice sheets in a warming climate. However, the processes, topologies and products of ice sheet hydrology are some of the least understood components of both past and modern ice sheets. This is to some extent a result of a disconnect between the fields of theoretical, contemporary observational and palaeo glaciology that each approach ice sheet hydrology from a different perspective and with different research objectives. With an increasing realisation of the potential of using the past to inform on the future of contemporary ice sheets, bridging the gaps in the understanding of ice sheet hydrology has become paramount. Here, we review the current state of knowledge about ice sheet hydrology from the perspectives of theoretical, observational and palaeo glaciology. We then explore and discuss some of the key questions in understanding and interpretation between these research fields, including: 1) disagreement between the palaeo record, glaciological theory and contemporary observations in the operational extent of channelised subglacial drainage and the topology of drainage systems; 2) uncertainty over the magnitude and frequency of drainage events associated with geomorphic activity; and 3) contrasts in scale between the three fields of research, both in a spatial and temporal context. The main concluding points are that modern observations, modelling experiments and inferences from the palaeo record indicate that drainage topologies may comprise a multiplicity of forms in an amalgam of drainage modes occurring in different contexts and at different scales. Drainage under high pressure appears to dominate at ice sheet scale and might in some cases be considered efficient; the sustainability of a particular drainage mode is governed primarily by the stability of discharge. To gain better understanding of meltwater drainage under thick ice, determining what drainage topologies are reached under high pressure conditions is of primary importance. Our review attests that the interconnectivity between research sub-disciplines in progressing the field is essential, both in interpreting the palaeo record and in developing physical understanding of glacial hydrological processes and systems.

Keywords: Glacial hydrology, ice sheet hydrology, meltwater, eskers, meltwater channels, glacial geomorphology.

1. Introduction

Ice sheet hydrology has long been recognised as a crucial component in the understanding of ice sheets, their behaviour and their evolution. The presence and distribution of water at the ice sheet bed is widely considered to control ice motion by facilitating basal sliding and substrate deformation (e.g. Alley et al., 1986; Engelhardt and Kamb, 1997), triggering enhanced or episodic fast flow via water pressure build-up (e.g. Kamb, 1987; Zwally et al., 2002), and impeding ice flow due to basal freeze-on, water piracy and bed stiffening (e.g. Alley et al., 1994; Tulaczyk et al., 2000; Bougamont et al., 2003). Whilst the time domain for ice sheet-scale behaviour has traditionally been considered to be from 100s – 1000s years (Alley and Whillans, 1984; Huybrechts and Oerlemans, 1990; Clark, 1994; Kleman et al., 2006; Kleman and Applegate, 2014; cf. Bamber et al., 2007) meltwater processes are highly spatially and temporally variable. This difference in conceptual and observational scales of different glaciological properties presents a problem for our understanding of the physical processes of meltwater drainage, of ice dynamic responses, of the land-forming footprint of drainage, and for implementation of these processes within large scale models.

After decades of studies of meltwater processes at a glacier-scale (e.g. Mathews, 1964; Stenborg, 1970; Iken and Bindschadler, 1986; Willis et al., 1990; Bingham et al., 2003; Mair et al., 2003), the last decade has seen a shift towards an ice sheet scale focus, largely out of concern for the rate of change and the future of the modern-day ice sheets. Pronounced seasonality of meltwater drainage linked to high summer supraglacial meltwater production has been observed in the ablation zone of the Greenland Ice Sheet (Zwally et al., 2002; Bartholomew et al., 2011a). On top of seasonal changes in subglacial efficiency, high-magnitude supraglacial lake drainage events connect the ice surface to the bed through ice up to 1000 m thick (Das et al., 2008; Doyle et al., 2013). In Antarctica, repeated drainage of water between subglacial lakes has been inferred, as well as water transfer towards the ice margin and interior freeze-on areas (Fricker et al., 2007; Wolovick et al., 2013). Such growing understanding of ice sheet scale drainage behaviour brings the scale of contemporary ice sheet research in line with that more typically inhabited by palaeo glaciology. Elements of palaeo meltwater drainage networks have long been recognised on the beds of Pleistocene ice sheets (e.g. Mannerfelt, 1945; Prest et al., 1968). Meltwater landforms have been widely used for reconstructions of past ice geometries and dynamics (e.g. Borgström, 1989; Kleman et al., 1992; Boulton et al., 2001; Margold et al., 2013a,b), but only limited focus has been directed towards the processes and function of the palaeo hydrological system.

In parallel to observational research, developments within the ice sheet modelling community increasingly allow for credible predictions of ice sheet responses and feedbacks under future climate scenarios (e.g. Gregory and Huybrechts, 2006; Greve et al., 2011; Applegate et al., 2012; Barrand et al., 2013; Edwards et al., 2014), and enable Quaternary (and older) ice sheet behaviour and its coupling to longer term climate regimes to be explored and better understood (e.g. Huybrechts and Oerlemans, 1988; Payne et al., 1989; Marshall and Clarke, 1997a, b; Tarasov and Peltier, 2004; Pollard and DeConto, 2009; Gregoire et al., 2012). Ice dynamics are arguably reasonably well captured in the latest generation of ice sheet models, but a realistic representation of ice sheet hydrology is a vital, missing component which must be incorporated for more robust simulations of likely ice sheet responses to increased meltwater production. For this we need a reliable, quantitative description of the processes and form of ice sheet hydrological drainage systems. We therefore consider it timely to review ice sheet hydrology as it stands and highlight the crucial issues for future consideration.

Ice sheet hydrology has generally been approached from different perspectives within three research sub-disciplines: *theoretical glaciology* which describes glacial processes in physical terms, contemporary *observational glaciology* within which real-time data concerning modern ice sheets are gathered, and *palaeo glaciology* which reconstructs the dynamics of past ice sheets. Although each of these fields is informed by the progress of the others, these research sub-disciplines traditionally work within different domains of space and time, with research conducted by different communities and developing along different trajectories. A number of reviews of glacial hydrology provide comprehensive summaries from different research perspectives (Clark and Walder, 1994; Hubbard and Nienow, 1997; Fountain and Walder, 1998; Benn and Evans, 2010; Bingham et al., 2010; Walder, 2010; Irvine-Fynn et al., 2011; Cook and Swift, 2012; Chu, 2014; Flowers, 2015). However, there have historically been few attempts to approach glacial hydrology from a fully cross-disciplinary perspective, leading to a disconnect of theoretical, contemporary observational and palaeo glaciology in their current understanding of ice sheet hydrology. With research agendas converging on the ice sheet scale, there is now the prospect for more integrated research designs that will tackle questions that have thus far been addressed semi-independently. Our effort here is to identify these disconnects between the findings and understandings of the sub-disciplines and to assess to what extent is it possible to merge the advances in each field into a more holistic view.

We first review the contributions of the three research sub-disciplines, focussing on the form, distribution and spatial scale of the hydrological system, and the timescales, rates and fluxes of meltwater delivery. We do not hope to be exhaustive in these reviews; other dedicated reviews better meet this demand. Our reviews rather serve as a basis for discussion of the areas of consensus in understanding and the disconnects, and significance thereof, between research fields, in order to address questions most urgently demanding attention. The discussion focusses primarily on subglacial drainage and inputs to subglacial drainage from the supraglacial domain. Here we identify a number of questions for which significant uncertainty remains, concerning the prevailing form, controls, variability, and spatial extent of subglacial drainage and the significance of high magnitude drainage events compared to long-term steady meltwater drainage. With this review, we aim to provide a critical basis for future collaborative efforts necessary to further our understanding of ice sheet hydrology and thus ice sheet behaviour as a whole.

2. Ice sheet hydrology: a theoretical perspective

In the subglacial domain, the spatial distribution and magnitude of effective pressure, p_e (defined as the ice pressure minus the water pressure, $p_e = p_i - p_w$), directly influences how efficiently the overlying ice can slide over its substrate. A key element of glacial hydrology, therefore, seeks to determine and explain how subglacial water pressures vary both spatially and temporally, in order to better understand the dynamics of ice flow. The rates of surface and basal melt, and the form of englacial transport, influence the mode and characteristics of subglacial drainage and the distribution of water pressure.

A fundamental concept of glaciohydrological theory, concerning both englacial and subglacial domains, is that of the hydraulic potential of water in a glacier system, Φ, first introduced by Shreve (1972), whereby

$$\Phi = \Phi_0 + p_w + \rho_w g z \tag{1}$$

where Φ_0 is a (constant) background potential, p_w is the water pressure, ρ_w is the density of water, g is acceleration due to gravity and z is the elevation. For a connected hydraulic system water will flow down the negative gradient of Φ. In the subglacial case, the gradient of the hydraulic potential, ∇Φ, will be due to the ice sheet's geometry (glacier bed, z_b, and surface, z_s) and due to p_e as

$$\Psi \coloneqq -\nabla\Phi = -\nabla(p_i - p_e + \rho_w g z_b) \approx -\rho_i g \nabla z_s - (\rho_w - \rho_i) g \nabla z_b + \nabla p_e \tag{2}$$

where ρ_i and ρ_w are the densities of ice and water (after Hewitt, 2011; Shreve 1972). This relation shows that the slope of the ice surface has a stronger influence than the bed topography. Therefore, subglacial water will not always follow bed topography but can be driven uphill (for instance, when ∇p_e is not a dominating factor). However, Eq. (2) does not in any way specify the processes involved in the transport of water; these unknowns are tacitly implied to determine p_e. Theoretical glacial hydrology seeks to address these specific processes.

2.1. Englacial water routing

Unless water is stored supraglacially (in lakes or a snow-firn aquifer) or routed away as surface runoff, it is fed to the englacial system (Fig. 1). A small amount of water can be transported between the boundaries of individual ice grains (Nye and Frank, 1973; Lliboutry, 1971), but this form of transport is incapable of flow of any significance (Lliboutry, 1971).

Shreve (1972) considered temperate ice as a hydraulically connected system of englacial conduits, with flow dictated by an upslope gradient of Φ -11 times that of the ice surface, driving water towards the bed in the direction of ice surface slope perpendicular to the potentials of Φ. Expanding Eq. (1) to include the role of ice deformation (conduit closure) and melt caused by flowing water (conduit opening), Shreve (1972) examined the differential growth of interconnected conduits by comparing the conduit wall melt rate. He concludes that out of two conduits of different radius, the melt rate will be higher for larger channels which will therefore tend to feed off smaller conduits. A dendritic (arborescent) network is created (Shreve, 1972) with conduits oriented along $-\nabla\Phi$. Taking a similar approach, Röthlisberger (1972) relates the water pressure within tubular channels to the discharge, showing that the pressure drops with an increase in discharge. This would, at least conceptually, result in the same type of network suggested by Shreve (1972), since the larger low pressure channels would consume water from higher pressure zones.

A Shreve topology assumes a coherent ice mass acting as a hydraulically connected system, although this is under steady-state conditions. Due to this limitation, much recent research has turned to focus on hydro-fracturing through ice sheets, to better address observations that water primarily exploits weaknesses and structures within the ice mass (Fig. 1). Hydrofracture theory has also been invoked to explain short-term events such as the drainage of supraglacial lakes on Greenland (van der Veen, 2007; Krawczynski et al., 2009). Initially Weertman (1973) showed, by treating ice as an elastic material with negligible fracture toughness, that the difference in ρ_w and ρ_i could cause crevasse propagation through an ice sheet. Later studies have used a more sophisticated model in the form of linear elastic fracture mechanics (LEFM), which has the possibility to include a parameter for the fracture toughness of ice (Smith, 1976; van der Veen, 1998, 2007). The rate at which a water-filled crevasse propagates can be approximated by

$$d = \left(\frac{\rho_w}{\rho_i}\right)^{\frac{2}{3}} Q_r t \qquad (3)$$

where d is the depth of the crevasse, t is time, Q_r is crevasse filling rate (metres per hour) and ρ_w is the density of water (van der Veen, 2007). Thus the control on how a crevasse *propagates* is dependent mainly on Q_r, but crevasse *initiation* is dependent on ice mechanical properties and far field stresses. Single crevasses can be expected to exist in zones where tensile stresses exceed $30 - 80$ kPa, while a field of crevasses affects the distribution of elastic stresses and therefore requires larger tensile stresses (90−320 kPa). If an abundant supply of meltwater exists in such areas, crevasses can potentially penetrate an ice mass within hours to days (van der Veen, 2007; Alley et al., 2005), and water can rapidly be delivered through the englacial system.

Fountain and Walder (1998) focussed on the energy dissipation of flowing water at the bottom of a crevasse, rather than the initiation or propagation of the crevasse itself. Assuming that the energy is used to melt the surrounding ice, they find the relationship

$$\dot{d} = c_1 S^{\frac{19}{16}} Q^{\frac{5}{8}} \tag{4}$$

where $\dot{d}$ is the downcutting rate, c_1 is a constant dependent on the roughness of the channel (bottom of the crevasse), latent heat of melting and density of ice and water, S is the along flow slope of the channel and Q is the water flux. Even at modest values for Q ($5 - 10$ m^3 s^{-1}), a water channel in a longitudinal crevasse could have a down cutting rate of approximately 10 m a^{-1}. Discharge rates – and thus meltwater availability – are therefore key active controls on the englacial system and the transfer of meltwater to the bed: the water flux can change the form of the system which carries it, and does so rapidly via crevasse propagation and thermal downcutting.

2.2. Subglacial water routing

Subglacial drainage theories similarly seek to relate pressure and flux of water. Here, the bed properties are an additional determinant of system geometry; different theories variably treat the bed as a fixed, unchanging boundary, or one which interacts with and is shaped by the drainage system. Subglacial meltwater transfer can primarily be categorised as either *fast* or *slow* (Raymond et al., 1995). Röthlisberger (1972) described the interaction of melting ice and water flow through a semi-circular, pipe-like channel, and a system of such channels is widely thought to be the dominant process of fast water transport in alpine glaciers. Consequently, fast systems are often conceptualised as *channelised* or *discrete*. However, recent observations have raised the possibility of a fast water transport under ice sheets that may not be in the same configurations as that considered by Röthlisberger (see Sections 3 and 5). Therefore, we use the term *fast* in a way that includes, but is not restricted to *channelised/discrete* systems. In contrast, a range of hypotheses for a slow hydraulic system include the flow of water through: a connected set of water filled cavities (Walder, 1986; Kamb, 1987), a subglacial water film or sheet (Weertman, 1972; Creyts and Schoof, 2009), through a porous substrate aquifer (Flowers and Clarke, 2002), and downcutting into sediment (Walder and Fowler, 1994) or bedrock (Nye, 1973).

2.2.1. Fast systems

Given, in the idealised case, a circular (if englacial) or semi-circular (if at the interface between the ice and a hard bed) channel cross-section, and flow in the horizontal plane, the water pressure can be related to the discharge and the effective pressure (Fig. 1) as

$$\frac{dp_w}{dx} = c_2 Q^{\frac{-2}{11}} \left(\frac{p_e}{nA}\right)^{\frac{8n}{11}} \tag{5}$$

where Q is as above, n and A are parameters from the flow law (Glen, 1952), c_2 is a constant dependent on ice rheology, channel roughness/shape and the pressure melting effect, and x is the horizontal distance along channel (after Röthlisberger, 1972; Shreve, 1972; Walder, 2010). From Eq. (5) one can see that the change in water pressure decreases with discharge.

Conceptually, flow capture across a pressure differential would lead to a dendritic, channelised system.

A modification to the theory was made by Lliboutry (1983), who claimed that glacier sliding and bed separation (cavities) were a necessity for channels to exist, and that full channels (higher p_w than atmospheric) occur only at short periods of high discharge, in which the channel grows in size. Lliboutry suggested that the normal state would be water flowing at atmospheric pressure in a channel that is closing due to the plastic deformation of the overlying ice, a conclusion that was also reached by Hooke (1984). Furthermore Hooke et al. (1990) argued that the restriction to a semi-circular shape was unrealistic due to the fact that stress regimes would alter at the glacier bed interface where friction occurs. Moreover there would be preferential melt of the side of the channel walls under conditions of open flow. They showed that if the channel geometry was modified to be low and wide, the closure rates of a channel could be explained without having to prescribe the ice to be unusually soft (as in Röthlisberger, 1972). These modifications to the basic Röthlisberger theory predict a multi-arborescent, widespread system of low, wide channels cut into the basal ice.

2.2.2. Slow systems

An irregular bed topography and glacier sliding creates subglacial cavities, which may fill with water. Walder (1986) and Kamb (1987) considered conditions in which cavities are created on the lee side of bedrock outcrops, with an interconnection between such cavities in the form of small orifices. Initially these studies aimed to explain the onset of glacier surges, but have subsequently been included in the general notion of slow systems. The discharge can be related to the sliding speed (u_b) and hydraulic potential as

$$Q = u_b^m \left(\frac{d\Phi}{ds}\right)^{\frac{1}{2}} {p_e}^{-n} \tag{6}$$

where m is a constant and s is the along-flow path coordinate (after Fountain and Walder, 1998). For a given discharge, this system would result in a much lower p_e than in the fast system of Eq. (5). Moreover, according to Eq. (6), cavities with higher discharge would have higher water pressures than cavities with lower discharge. This does not therefore lead to a capturing mechanism as in the above described channelised systems. A dendritic system does not develop; rather, the topology of the system is governed by the irregular pre-existing distribution of bedrock obstacles. Comparing Eqs. (5) and (6) qualitatively, one can conclude that Röthlisberger channels (R-channels) are more dependent on the energy dissipation of flowing water while linked cavity systems are dependent on the sliding speed. Both Walder (1986) and Kamb (1987) noted that increasing the discharge (water pressure) above a critical level would lead to a situation where cavities unstably increase in size. It was suggested that this instability could lead to a situation favourable to R-channel initiation. However, the orientation of the cross section of the orifices/cavities in Kamb's 2-D model (across ice flow/surface gradient) would be perpendicular to the expected orientation of an R-channel (along ice flow/surface gradient).

To explain glacier sliding, Weertman (1962, 1972) had early considered the concept of a continous subglacial water film (mm in thickness) at the ice-bed interface. In a series of papers Lliboutry (see 1979) argued that the resulting stresses needed in Weertmans's formulation to reproduce normal ice velocities would be too large to be supported by either ice or bedrock. Lliboutry suggested that bed separation on the lee side of bed protrusions must be a significant process. Later the stability of Weertman's water film was questioned by Walder (1982). Creyts and Schoof (2009) have recently expanded the concepts originally presented in Weertman (1972) and Weertman and Birchfield (1983). They showed that, on an uneven bed with protrusions to partially support the overlying ice, several stable states of water sheet thickness could exist. This idea is qualitatively similar to the slow system described by Walder (1986) and Kamb (1987), in that the water sheet would be expected to increase in height with decreasing effective pressure, but different in that the system can adapt to changes in hydraulic potential and effective pressure from one steady-state to another. Therefore, with changes of water sheet thickness, different levels of efficiency can result from the same basic system.

2.2.3. ‘Active beds’ – slow systems?

The above descriptions of the hydrological system treat the bed as a stable, passive participant in meltwater delivery, and these concepts are typically applicable to ‘hard beds’. Nye (1973), however, argued that R-channels were likely to be advected and squeezed shut against bedrock irregularities due to the movement of overlying ice. He concluded that, in addition to a very thin Weertman type water film (important for regelation processes), a more persistent type of channel cut into the underlying bedrock would exist. This type of channel is commonly referred to as a Nye-channel. Such channels are known to exist, but have received little theoretical attention; Fountain and Walder (1998) treat Nye channels as a component of a slow drainage system.

Sediment substrates are known to underlie large parts of contemporary and palaeo-ice sheets, yet how water flows through or interacts with a subglacial sediment layer remains to a large degree uncertain. It is unlikely that the underlying substrate could be permeable enough to evacuate all water through porewater, aquifer-type processes (Alley, 1989), even in an ice sheet setting where all melt is generated at the bed and water volumes are small. Shoemaker (1986) found that in order for the till to be an efficient aquifer, it must be integrated with a channel system in which channel spacing must be very close (on the order of $10^2 - 10^3$ m). Walder and Fowler (1994) described till as a creep flow material, and showed that, as long as the creep of till is faster than that of ice, subglacial channels cut into the sediment (canals) could exist. Due to sediment erosion and transport, the shape of canals would have a low aspect ratio. For a low ice surface slope typical for ice sheets or ice streams ($\sin \alpha = 0.001$, where α is the ice surface slope), sediment canals together with flow through the till would be the stable hydrological configuration. Contrary to R-channels, a system of canals would have p_e close to zero and would not be arborescent. For fine-grained sediment, canals would have an approximate depth of 10 cm at a discharge of $1 m^3 s^{-1}$ (Walder and Fowler, 1994). Ng (2000) expands on the canal concept by including mechanisms for sediment transport and downstream variability of the canal. The type of inverse relationship between effective pressure and discharge as presented by Walder and Fowler (1994) is concluded to hold, but that the discharge and sediment flux are both needed to determine the downstream distribution of the effective pressure.

Kyrke-Smith and Fowler (2014) continue the idea of Creyts and Schoof (2009) that flow can occur as a water film beneath a glacier supported by bed protrusions, and additionally incorporate the processes of sediment erosion and transport into their model. They find that a uniform water film of about 5 mm would be unstable, and that shallow ‘streams’ would form. Kykre-Smith and Fowler (2014) suggest that these streams would be more swamp-like in nature and would be in the order of a centimetre deep and hundreds of metres wide. While this process of water flow would be considered slow, the study offers the possibility of a swamp transitioning to a channel, although the physics behind this transition is unknown.

2.2.4. Interactions between subglacial hydraulic systems

It is unlikely that any one hydraulic system would be exclusive to an ice sheet, either spatially or temporally. Which of multiple coexistent components comes to dominate may depend on water flux, the ‘permeability’ of different domains, and the system geometry. Walder and Fowler (1994) argued that a till aquifer can coexist with sediment canals and R-channels, the dominant component depending on the ice surface slope (low slopes favouring canals). Hewitt (2011) examined how water may be drawn from a porous medium into channelised flow, and the spatial scales which may be appropriate in such an integrated system. He provides an analytical criterion for the position of channel initiation (i.e. channel length: distance up-ice from the margin) and a spacing relation coupled to this extent criterion. For an idealised ice sheet (characteristic length 1000 km; thickness 1 km) channel length would be 10-100 km and spacing 7-15 km. Qualitatively, longer channels should develop where there are steep slopes, a high discharge, and a low permeability in the distributed system. These same conditions reduce the spacing between channels of a given length. Hewitt (2011) concludes that channel length and spacing are governed by the (in)efficiency of the distributed system.

Schoof (2010) modelled the switch between a (inefficient) cavity system and a (efficient) channelised system according to the equation

$$\frac{dR}{dt} = c_4 Q \Psi + u_b h - c_5 p_e^n R \quad (7)$$

where R is the cross-sectional area of the cavity or channel, *h* is a characteristic height of bed protrusions, ψ is the hydraulic potential gradient along the conduit and c_4, c_5 are constants. For a steady-state, Eq. (7) implies that there exists a critical discharge, Q_c (corresponding to the lowest p_e). Channels will form if $Q > Q_c$, otherwise the system will consist of cavities. Modelling a transient case with variable water input shows that a relatively low p_e cavity system exists at time of low water input while channels dominate when the input is high. The switch between the systems is characterised by a peak in high p_w (low p_e). The model predicts that, during channelisation, the mean channel/cavity size will decrease but the variance of R will increase. Both Schoof (2010) and Hewitt (2011) see discharge as a critical parameter, to both temporal switches and spatial scale controls.

2.3. Theoretical ice sheet hydrology: summary

Relating the topology and processes acting in a subglacial system to the water pressure therein is a focal point of theoretical subglacial hydrology. Conceptually, one can subdivide subglacial water transport into two modes: slow or fast. Theoretically, the water pressure of a linked cavity system will increase with an increase in discharge, leading to a natural limit of the system at ice overburden pressure where the topology or the character of the system must change. Conversely, the water pressure in R-channels decreases with an increase discharge, resulting in an ability to transport water without the water pressure reaching overburden pressure. Hence, the common idea of a slow system being characterised by distributed cavities and a fast system by arborescent R-channels stems from the theoretical aspects considered in the 1970s and 80s. Theoretical developments have been stimulated by observations of more complex behaviours and topologies in modern and palaeo ice sheets, and a desire to apply numerical implementations to realistic settings. Ultimately, however, theory is an idealised representation of simplified systems. The limits of current theory do not exclude alternative system topologies that could arise from or result in alternative pressure regimes for so-called fast or slow systems.

3. Ice sheet hydrology: contemporary observations

Theoretical and observational glacial hydrology have progressed in reasonably close association, seeking to develop theory in order to explain observations of glacial hydrological systems. However, the impenetrable nature of glacier ice to GPS (global positioning systems) signals and the difficulties of obtaining direct observational data, particularly at ice sheet scale, constrain our knowledge of englacial and subglacial drainage systems. The nature and dynamics of these systems is thus inferred from surface and proglacial monitoring and from ice dynamic proxies for subglacial and englacial efficiency. Notwithstanding such limitations, observations and glacial hydrological theory lead us to a conceptual framework for the hydrological system resembling Figure 1.

3.1. Supraglacial and englacial reservoirs and pathways

The ablation zone of ice masses is often characterised by networks of streams running across the ice surface (Ferguson, 1973). These streams provide a mechanism for efficient transport of meltwater to the margin, or into supraglacial lakes and moulins, from which water can be delivered into the englacial and subglacial systems. In the snowpack and firn, refreezing and percolation of meltwater can produce superimposed ice (Fig. 1D), lowering primary permeability and encouraging meltwater storage within supraglacial ponds and slush zones in topographic lows (Müller, 1962; Boon and Sharp, 2003; Wright et al., 2007; Irvine-Fynn et al., 2011). Vertical percolation rates are highly heterogeneous (Gerdel, 1954; Schneider and Jansson, 2004; Campbell et al., 2006) and meltwater retention in supraglacial and firn zones buffers ice sheet and sea level response to surface melting (Fig. 1D, Harper et al., 2012; Humphrey et al., 2012; Forster et al., 2014).

3.1.1. Supraglacial lakes and their drainage

Storage of water in supraglacial lakes and ponding is commonplace in the ablation zone of the Greenland Ice Sheet, and can be present on Arctic and alpine glaciers where influenced by debris cover, slush zones and closed surface fractures. Lakes provide a point source for the delivery of large volumes of meltwater to the bed (Fig. 1) through existing fractures or weaknesses in the ice, supplying heat and preserving open moulins (Alley et al., 2005). Their catastrophic drainage provides one mechanism for rapid transfer of large quantities of meltwater directly to the subglacial system (van der Veen, 1998; Alley et al., 2005), and hydrofracture processes are thought to be important in this transfer (Fig. 1A). Krawczynski et al. (2009) found that lakes of 250-800 m diameter contain enough water to drive hydrofracture to the bed through 1 km of subfreezing ice; such lakes encompass 98% of surface-stored meltwater in central-west Greenland, which could reach the bed through full fracture propagation. Furthermore, basal water pressures greater than overburden could be produced when moulins are water-filled. Greenland lake drainages have been shown to increase seismicity, horizontal ice surface velocities and cause vertical uplift (Das et al., 2008, Doyle et al., 2013), with peak event discharges on the order of Niagara Falls' ~8700 m^3 s^{-1} having been recorded (Das et al., 2008).

Box and Ski (2007) estimate that a moulin with a cross-sectional area of 10 m^2, penetrating through 800 m thick ice, would hold only 0.1% of an 'average' lake's drained volume at any one time, based on lake dimension statistics for a sample of lake outburst events in west Greenland. Elevated basal water pressure could therefore be maintained, at least locally, for hours to days (e.g. Box and Ski, 2007; Doyle et al., 2013). Selmes et al. (2011) report that 13% of Greenland's lakes undergo rapid drainage each year (averaged over a 5-year period). Of these fast drainage events, 61% occurred in the southwest sector of the ice sheet, in contrast to only 1% in the southeast. This highlights pronounced regional differences in supraglacial meltwater drainage processes, in addition to large regional differences in lake numbers and total lake area (Selmes et al., 2011). The net effect of short-term drainage events on longer-term ice surface motion and subglacial network efficiency remains unclear. Sundal et al. (2011) suggest that seasonal, surface-melt induced speed-ups may be relatively insignificant at an annual timescale, although their data were restricted to marginal regions. Furthermore, recent modelling indicates that single-event lake drainages, although large, transfer a much smaller proportion of annual melt to the bed than the more steady flow of meltwater streams through moulins (Clason et al., 2015). Under future climatic warming scenarios the spatial coverage of supraglacial lakes may extend further into the interior of the Greenland Ice Sheet (Leeson et al., 2015), and may have potential to deliver meltwater to the ice sheet bed in areas of widespread inefficient subglacial drainage. However, the extent to which lakes forming at higher elevation could drain through new surface-to-bed connections is uncertain. Drainage may be limited by the relative dearth of existing crevasses through which water can reach the bed, producing tensile stresses in otherwise compressive areas (Stevens et al., 2015).

3.1.2. Moulins and englacial conduits

Moulins, from centimetres to 10s of metres in diameter, form through the intersection of supraglacial streams with surface crevasses, and transport meltwater into the englacial system (Fig. 1; Fig. 2A). At both valley glacier and ice sheet scale, moulins connect drainage directly to the bed. Dye tracing studies and water pressure measurements (e.g. Nienow et al., 1996; Hock et al., 1999; Fudge et al., 2008), direct moulin observations (e.g. Holmlund, 1988; Reynaud and Moreau, 1995; Schroeder 1998; Steffen et al., 2009) and numerical modelling (e.g. van der Veen, 1998; Clason et al., 2012) indicate the connectivity of moulins and crevasses to the englacial and subglacial systems. Furthermore, moulins propagating the full ice-thickness, possibly persistent for multi-year timescales, have been interpreted from repeat ice penetrating radar profiles (e.g. Catania et al., 2008; Catania and Neumann, 2010). Development and maintenance of these systems is considered highly dependent on the water flux. In areas of closely spaced crevasses the stress required to maintain hydrofracture is considerably larger than for individual crevasses (van der Veen, 1998), but where hydrostatic pressure from meltwater inflow is sufficient, drainage of saturated crevasse fields could deliver substantial volumes of meltwater to the bed, as has been demonstrated for Jakobshavn Isbrae in west Greenland (Lampkin et al., 2013). However, cold surface layers (Holmlund and Eriksson, 1989; Blatter and Hutter, 1991; Pettersson et al., 2003) and retention of meltwater in the snow and firn packs (Stenborg, 1970; Schneider, 1999; Irvine-Fynn et al., 2011) not only dampen supraglacial water delivery but hinder the ability of crevasses or moulins to be driven through the ice by hydrofracture. Unless continuous meltwater delivery is maintained, low ice temperatures may promote refreezing and plugging of fractures, and cessation of further fracture propagation (e.g. Boon and Sharp, 2003).

There has been a recent shift towards a greater focus on englacial hydrology, yet the exact form of surface-to-bed connectivity remains largely unobserved and unknown. This is especially true for high magnitude drainage events; a stark contrast to traditional thinking on long-term steady basal melting. On a macro-scale the englacial system consists of moulins, crevasses, fractures and enclosed horizontal conduits (Fig. 2). Rather than the dendritic topology of veins and conduits first hypothesised by Shreve (1972), direct imaging, exploration and speleology suggest a complex network of fractures and conduits (e.g. Vatne, 2001; Fountain et al., 2005a, b; Benn et al., 2009; Gulley et al., 2009a). Benn et al. (2009) directly explored englacial drainage systems in Alaska, Nepal and Svalbard, finding that hydrofracturing could lead to englacial system development across a large range of glaciological regimes (Fig. 2). Fracture formation and geometry was found to be controlled by supraglacial meltwater ponding and exploitation of resealed fractures and ‘crevasse traces’ (cf. Holmlund, 1988), while repeated formation of conduits and fractures in the same places was postulated to yield stable locations for subglacial recharge (Benn et al., 2009; Catania and Neumann, 2010). Fountain et al. (2005a, b) argue that fracture systems dominate englacial water flow, and Gulley et al. (2009a) found that even in crevasse-free areas of polythermal glaciers, englacial channels could evolve through downward incision of meltwater channels, as previously described by Vatne (2001), coining the term ‘cut-and-closure’ channels (Fig. 2C). Gulley et al. (2009a) observed down-cutting rates of up to 0.33 m day^{-1}, allowing full incision to the bed, although channel growth occurred more rapidly via knickpoint migration upstream. Compiled data on observed englacial conduits indicate lengths ranging from 10s to 1000s metres at depths ~2-100 m below the ice surface (Gulley et al., 2009b). Little evidence has been found to support convergence of water-filled capillaries into larger conduits or for flow pathways strongly related to hydraulic gradient (cf. Shreve, 1972). Rather, slow- and fast-flow englacial components may co-exist. Large, spatially distributed networks of small fractures (tens of millimetres in width) found throughout the ice column likely operate simultaneously alongside fast, efficient flow through englacial channels and moulins (cf. Jansson, 1996; Fountain et al., 2005a, b).

3.2. Subglacial configuration

Much of our understanding of the configuration of subglacial drainage systems is inferred from monitoring of other, more accessible parts of the hydrological system or from ice dynamic proxies. Monitoring of proglacial discharge for its physical and chemical characteristics and for artificial tracer investigations (e.g. Willis et al., 1990; Nienow et al., 1996, 1998; Bingham et al., 2005; Cowton et al., 2013), study of boreholes drilled from the surface (e.g. Fountain, 1994; Hubbard et al., 1995; Mair et al., 2003; Fudge et al., 2008) and use of surface velocity patterns as a proxy for meltwater drainage events and subglacial drainage efficiency (e.g. Iken and Bindschadler, 1986; Kamb, 1987; Jansson, 1996; Mair et al., 2002; Bartholomew et al., 2012) are commonly used to deduce the likely form of the subglacial drainage network. Water pressure fields, meltwater delivery rates and outflow properties are related to meltwater routing through the subglacial domain, often then conceptualised as efficient or inefficient, or a channelised or distributed system (Fig. 1).

3.2.1. Insight from small ice masses

Investigation of proglacial discharge and surface-to-bed boreholes has for many years been implemented on alpine and arctic glaciers, and many major advances in subglacial hydrology come from this environment (see reviews by Hubbard and Nienow, 1997, and Fountain and Walder, 1998). Tracer studies widely show seasonal evolution of the subglacial drainage system (e.g. Seaberg et al., 1988; Willis et al., 1990; Hock and Hooke, 1993; Nienow et al., 1996, 1998), with dye return curves measured in proglacial outlets typically evolving from a slow, diffuse and multi-peaked return to a single-peaked fast return as the melt season progresses. Corresponding transition from a distributed to a channelised drainage configuration is widely assumed, with subsequent enlargement of channels, headward extension and 'rationalisation' of the network (Fig. 1) associated with seasonal up-glacier extension of surface meltwater production and retreat of the snowline (e.g. Nienow et al., 1998; Bingham et al., 2005). Borehole measurements of water pressure support a general model of seasonally enhanced drainage efficiency, with summer low pressure zones and synchronous fluctuations across boreholes (e.g. diurnal) taken to indicate a connected, channelised system (e.g. Fountain 1994; Hubbard et al., 1995; Harper et al., 2002; Fudge et al., 2008).

Superimposed on this seasonal model is considerable spatial and temporal heterogeneity. Punctuated events such as supraglacial lake drainage events (Chu et al., 2009) and outburst floods or jökulhlaups (Russell, 1989; Russell et al., 1990; Mernild et al., 2008; Werder et al, 2009; Russell et al., 2011) are detected in proglacial discharge. Large borehole pressure variations, high gradients and sites disconnected from the main hydraulic system have been reported across relatively short distances (e.g. Murray and Clarke, 1995; Fudge et al., 2008). Hubbard et al. (1995) report diurnal interaction between a high pressure ('distributed') system and a low pressure channel, with increasing melt and rising pressure forcing water from the channel into the surrounding distributed system. While simultaneous variance in water pressure has been documented in closely clustered boreholes, observations by Fudge et al (2008) suggest that in late summer across-glacier connectivity is limited, and infer that long down-glacier water flow is necessary to drive large diurnal pressure fluctuations recorded within a relatively small area of the bed: the system is channelised, but not classically arborescent. Interpretations of drainage topology, connectivity and efficiency often make assumptions that borehole studies have difficulty substantiating; local borehole pressures may not reflect the pressure conditions of the wider system (Fudge et al., 2008; Meierbachtol et al., 2013), and limited spatial coverage of boreholes restricts the possibility of connecting with channelised drainage (Andrews et al., 2014).

An observed correspondence between borehole water level fluctuations and surface ice flow velocities led Iken and Bindschadler (1986) to conclude that sufficient basal water pressures could locally hydraulically jack a glacier from its bed. Kamb (1987) drew similar conclusions from monitoring at Variegated Glacier during its 1982-83 surge event, further suggesting that a switch in subglacial regime from a low pressure, channelised system to a high-pressure linked-cavity (distributed) system may act as a trigger mechanism for glacier surges and fast ice flow. Despite uncertainties in linking basal water pressure to drainage topology, that changes therein can directly precipitate ice dynamic responses has been demonstrated in numerous settings (e.g. Jansson, 1995; Mair et al., 2003; Bingham et al., 2003; Sugiyama and Gudmundsson, 2004). On individual Arctic and alpine glaciers, small scale spatial patterns in surface ice velocity change in accordance with local basal water pressure changes are now well-established, both in terms of high pressure 'spring events' (e.g. Jansson and Hooke, 1989; Mair et al., 2003) and subsequent dampening of velocity responses as the melt season progresses and the drainage network becomes increasingly efficient.

3.2.2 Scaling up: ice sheets

The last decade has seen a shift of focus to the ice sheet scale when Zwally et al. (2002) argued that seasonal velocity patterns of outlet glaciers of the Greenland Ice Sheet are linked to seasonal drainage changes in a similar way to alpine contexts. The mode of surface meltwater delivery to the bed via moulins and crevasses modulates basal sliding (Colgan et al., 2011), and may further influence ice dynamics via cryo-hydrologic warming (Phillips et al., 2010; 2013). Seasonal meltwater-forced velocity behaviour of some of Greenland's outlet glaciers is now well-documented (e.g. van de Wal et al., 2008; Shepherd et al., 2009; Bartholomew et al., 2010, 2011a, b; Sundal et al., 2011) and similar behaviour is exhibited by both terrestrial and marine-terminating outlets (Sole et al., 2011). However, uncertainty remains on the extent to which inland expansion of supraglacial meltwater penetration to the bed can offset a reduction in mean annual ice velocities caused by self-regulation of the subglacial drainage system and decreased winter velocities (Sundal et al., 2011; Sole et al., 2013; Doyle et al., 2014; van de Wal et al., 2015).

Bartholomew et al. (2011a, b) observed velocity response to seasonal melting below 1000 m a.s.l., and interpreted the expansion of an efficient subglacial drainage system to more than 50 km from the ice margin of Leverett Glacier in southwest Greenland. Dye tracing and instrumentation of meltwater conduits on the Greenland Ice Sheet attempts to better constrain the inferred subglacial network development, and reveals remarkable similarity with the seasonal evolution of alpine and arctic glacier hydrology from an initially inefficient system to one which rapidly delivers meltwater through an efficient network (Chandler et al., 2013; Cowton et al., 2013). Chandler et al. (2013) find that moulins >40 km from the ice margin, where the ice is ~1 km thick, connect with an efficient subglacial configuration, though at 57 km dye returns remain slow even at the end of the season, indicating that efficient drainage is currently limited to below 57 km in this region of the Greenland Ice Sheet.

However, these discrete proxy measurements provide relatively little insight to the actual configuration of subglacial drainage system. This is largely inferred, based on an assumption that 'efficient' drainage can be correlated with a channelised topology and delayed or diffuse delivery reflects a high pressure distributed system. This assumption is increasingly challenged. Harper et al. (2010) argue that basal crevasses act as subglacial meltwater stores and yield an irregular distribution of basal water pressure. Furthermore, measurement of borehole pressures in southwest Greenland, twinned with numerical modelling of channel maintenance, do not support the presence of stable systems of relatively low pressure more than 17 km inland of the margin (Meierbachtol et al., 2013), in contrast to Chandler et al. (2013); perhaps pointing to a different type of subglacial network. High ice-velocity events were observed by Cowton et al. (2013) up to 14 km from the Greenland Ice Sheet margin, even after the inferred channelisation of the subglacial system, perhaps due to meltwater inputs exceeding the drainage system capacity. Based on direct channel exploration in conjunction with dye tracing analyses, Gulley et al. (2012a, b) suggest that elevated pressure-induced sliding can occur in both channelised and distributed systems, and emphasise that dye breakthrough curves therefore are problematic for reliably determining changes in subglacial system configuration. Indeed, changes in efficiency of the non-channelised drainage system may account for observed velocity changes during the late melt season (Andrews et al., 2014). Finally, our knowledge of subglacial channel size is very much limited to those seen emerging at glacier fronts, which can be up to a few 10s of metres in width and may already have been subject to partial collapse, and to the very few direct observations beneath glaciers (Gulley et al., 2012a; Fig. 2D,E). Away from the most marginal outlets of the Greenland Ice Sheet, we have very little understanding of ice sheet subglacial hydrology where ice is thicker and overburden pressure higher. This has important implications for inferring the form, length scales and connectivity of subglacial systems.

In Antarctica the vast scales, limited surface meltwater availability and the logistical challenges of mounting field campaigns have prompted the application of alternative research strategies, which have in turn revealed rather different elements of the subglacial hydrological system. Many studies of Antarctic subglacial hydrology assume dominance of distributed drainage (Ashmore and Bingham, 2014). Geophysical techniques complemented by borehole studies revealed saturated sediments and high basal water pressures at the bed of the Whillans ice stream, stimulating a widely adopted conceptual model of fast ice stream flow facilitated by saturated soft beds (Alley et al., 1986; Blankenship et al., 1986; Engelhardt et al., 1990; Engelhardt and Kamb, 1997). A widely explored corollary of this model is that loss of meltwater from the subglacial sediment body, via freeze-on or via re-routing ('water piracy'), drives slow-down, sticking or stoppage of fast ice flow (Alley et al., 1994; Tulaczyk et al 2000; Christoffersen and Tulaczyk, 2003; Bougamont et al., 2003; Vaughan et al., 2008, Beem et al., 2014). In the Siple Coast sector, a highly heterogeneous pattern of basal melting and freezing has been detected and modelled (e.g. Vogel et al., 2003, 2005; Joughin et al., 2004; Jacobel et al., 2009), with a peak combined discharge across the whole Siple Coast grounding line (~300 $m^3 s^{-1}$, Carter and Fricker, 2012) comparable with the much smaller but supraglacially-fed Leverett Glacier in west Greenland (Bartholomew et al., 2011b). Beneath the Rutford Ice Stream, ice-penetrating radar and seismic methods identify a patchy mosaic (kilometre-scale) of saturated deforming sediments and ponded water bodies (e.g. King et al., 2004; Smith et al., 2007; Murray et al., 2008). The arrangement of the inferred ponded 'free water' is consistent with small cavities or with broad, shallow canals, in the order of 10 m wide and 0.1 m deep, incised into the soft bed (King et al., 2004; Murray et al., 2008).

There are recent indications that, in the marginal zone, subglacial channels may in fact be developed in Antarctica despite virtually no contributions of surface melt to the subglacial environment. Distributed canals and subglacial channels are interpreted under Thwaites Glacier based on radar scattering signatures (Schroeder et al, 2013), and a transition zone between distributed and channelised drainage is interpreted around 50 km from the grounding line. Le Brocq et al. (2013) similarly interpret grounding zone channelised meltwater, argued to be responsible for sub-ice shelf channels via grounding line meltwater plumes at the Filchner-Ronne Ice Shelf. Radio echo-sounding and satellite laser altimetry reveal hundreds of ponded, subglacial lakes under both the West and East Antarctic Ice Sheets, with typical length-scales of 5-10 km (Siegert et al., 2005; Carter et al., 2007; Smith et al., 2009). Repeat imaging has shown that these lakes periodically drain and fill, episodically delivering large volumes of water through the subglacial system of the ice sheet interior (Gray et al., 2005; Wingham et al., 2006; Fricker et al., 2007). Delivery is inferred to be via channels with network lengths of 10s-100s km (Wingham et al., 2006; Wright et al., 2012; Wolovick et al., 2013), though with no confirmed topology. Drainage occurs from lake basin to basin with timescales for individual events of months to years and associated maximum discharge estimates of 40-50 $m^3 s^{-1}$ (Wingham et al., 2006; Fricker et al., 2007). Wolovick et al. (2013) describe subglacial lake networks terminating in regions of basal accretion ice (Bell et al, 2011a) which indicate stability of an internally confined hydrological system over 10s kyr.

While Bell et al. (2011b) have identified regions of accretion ice beneath Petermann Glacier, in Greenland, similar extensive interior hydrological systems to those in Antarctica have yet to be identified beneath the Greenland Ice Sheet. Recent detection of *subglacial* lakes in Greenland via radio echo-sounding and remotely sensed ice surface elevation changes have been linked to the *supraglacial* meltwater system as sources of lake recharge via hydrofracture and englacial routing, and as triggers for lake drainage (Palmer et al., 2013; Howat et al., 2015; Willis et al., 2015). Willis et al. (2015) observed the recharge of a subglacial lake in conjunction with surface meltwater inflow into crevasses bordering the lake basin. Howat et al. (2015) theorise an increase in subglacial drainage efficiency in response to increased surface melting as a trigger for lake drainage. The abundance of surface meltwater and its supply to the bed of the Greenland Ice Sheet may then explain the relative dearth of subglacial lakes in comparison to the Antarctic Ice Sheets, such that a larger proportion of Greenland is underlain by efficient subglacial drainage, restricting the formation and maintenance of subglacial lakes (Palmer et al., 2013). The steeper ice surface profile of the Greenland Ice Sheet in comparison to the Antarctic Ice Sheet may also hamper subglacial lake formation through its influence on hydraulic potential (Livingstone et al., 2013).

3.3. Contemporary ice sheet hydrology: summary

Seasonal changes in drainage efficiency on the Greenland Ice Sheet appear to be similar to the seasonal development of an efficient channelised system seen in alpine glaciers. The extent of channelisation inferred from observations in west Greenland is, however, limited to around 50 km inland from the ice sheet margin. Hence, exporting knowledge of alpine glacier hydrology to larger systems may introduce a bias into interpretation of observations, since a clear correlation between drainage system form, pressure distribution and drainage efficiency has not yet been established. Structural controls on the form of the englacial system and on the initiation of surface-to-bed connections, including the drainage of supraglacial lakes, raise the possibility of stable locations for subglacial recharge on an inter-seasonal timescale. A growing proportion of the Greenland Ice Sheet is characterised by supraglacial lakes, the drainage of which influences ice dynamics and subglacial hydrology on at least a local scale. The physical form of the subglacial drainage system in response to such delivery, and its capacity to leave a landform imprint, remains unknown. Sequential drainage of subglacial lakes in Antarctica, however, hints at an internal, non-climatic control on the mode and timing of subglacial drainage, and the relative roles of the internal glacial system versus surface meltwater production in governing drainage processes on contemporary ice sheets remain poorly understood.

4. Ice sheet hydrology: palaeo ice sheets and the geological products

Studying the geological products of past ice sheet hydrological systems offers a broader-scale and longer-term view compared to contemporary observations. Glaciated landscapes also provide direct 'access' to the palaeo-subglacial domain. Meltwater landforms have long been documented at range of scales (Jamieson, 1863; Högbom, 1885, 1892; Mannerfelt, 1945; Ives, 1958; Soyez, 1974; Borgström, 1989; Dyke, 1993; Kleman et al., 1992; Russell et al., 2001; Mäkinen, 2003, Burke et al., 2008, Livingstone et al., 2010; Margold et al., 2011). They have widely been used to infer past ice sheet geometries and dynamics, based on relatively simple assumptions regarding their formation. Meltwater landforms and sediments have, however, been less often used to infer properties of the (palaeo) *hydrological* system (though see, for example, Shreve, 1985; Brennand and Shaw, 1994; Piotrowski 1997; Lowe and Anderson 2002; Shaw, 2002; Boulton et al., 2007a, b; Piotrowski et al., 2009; Boulton et al., 2009; Nitsche et al., 2013; Phillips and Lee, 2013; Storrar et al., 2014a; Burke et al., 2015; Lee et al., 2015; Livingstone et al., 2015).

Meltwater in supra-, sub- and proglacial environments has the capacity to create landforms; these subsequently have different preservation potential depending on the domain and the glacial phase (advance or retreat) in which they formed (see Mannerfelt, 1945; Sugden and John, 1976; Evans, 2003; Bennett and Glasser, 2009; Benn and Evans, 2010, for overview). Subaerially, meltwater channels are carved at the surface and margins of glaciers as well as in the glacier fore-field, and shorelines are eroded by wave action and accompanying processes in glacial lakes. Sediments are deposited where water loses the capacity to carry them farther: in kame terraces, on outwash plains and in deltas. In en- and subglacial environments, the pressure field is expected to influence the routing of meltwater and associated erosion and deposition, in turn determining the spatial distribution of meltwater landforms (Shreve, 1972, 1985; see Section 2.2). Subglacial streams erode meltwater channels, deposit eskers (these might also form in en- or supraglacial channels; Fitzsimons, 1991; Russell et al., 2001, Burke et al., 2008, 2009), or create more complex and composite features like meltwater corridors (Rampton, 2000). Subglacial meltwater channels and tunnel valleys are often seen to be evidence of Nye-channels, whereas eskers are taken to represent the sedimentary infill of former R-channels (Clark and Walder, 1994).

4.1. Ice-marginal meltwater landforms

Lateral and submarginal meltwater channels document drainage of supraglacial meltwater along the glacier margin, where ice has been pinned against a topographic slope. These channels tend to be relatively small (metre-scale in cross-section) and occur in well-developed series on high ground (Kleman et al., 1992; Dyke, 1993; Greenwood et al., 2007; Margold et al., 2011, 2013b). With increasing size of the channels (tens of metres in cross-section) their sinuosity increases and connections to the bed in the form of subglacial chutes or short esker segments appear; these are usually considered to be submarginal, at the lateral margin but beneath the ice surface (Mannerfelt, 1949; Greenwood et al., 2007; Syverson and Mickelson, 2009; Lovell et al., 2011; Margold et al., 2011, 2013b). It has traditionally been inferred that lateral channels form where the ice margin was cold based, preventing downward percolation (Kleman et al., 1992; Dyke, 1993; Kleman and Borgström, 1996). Where lateral meltwater channels are a dominant landform (parts of Scandinavia, the Canadian Arctic and Cordillera), cold-based or polythermal ice can be expected to have prevailed during deglaciation (Borgström, 1989; Dyke, 1990; Kleman et al., 1992; Dyke, 1993; Sollid and Sørbel, 1994). However, lateral meltwater channels have been reported to form at the margins of decaying warm-based Alaskan glaciers (Syverson and Mickelson, 2009) and our palaeo-glacial interpretations must thus be made with caution.

Historically, much attention has been paid to glacial lakes reconstructed from fossil shorelines and perched deltas (Jamieson, 1863; Gavelin and Högbom, 1910; Frödin, 1925; Charlesworth, 1955; Ives, 1960; Jansson, 2003). However, the focus of these studies has been on reconstructing past ice margin geometry and few attempts have been made to link the volume and evolution of glacial lakes to the rest of the ice sheet hydrological system. This is true also of other ice-marginal landforms such as outwash plains or kames (Clague and Evans, 1993; Evans et al., 1999; Livingstone et al., 2010; Kehew et al., 2012a).

4.2. Subglacial meltwater landforms

Glaciohydrological theory and observations at the margins of contemporary ice masses tell us that subglacial meltwater must be channelised, at least in parts. Palaeo-subglacial meltwater channels (also termed tunnel valleys or tunnel channels) are widely detected in the landform record (e.g. Sissons, 1958; Walder and Hallet, 1979; Sugden et al., 1991; Huuse and Lykke-Andersen, 2000; Rains et al., 2002; Lowe and Anderson, 2003; Greenwood, 2008; Hughes, 2009; Kehew et al., 2012b; Fig. 3), though they can be difficult to distinguish from their proglacial counterparts and extensions (Benn and Evans, 2010, p. 296). Often characterised by an anastomosing topology and undulating long-profiles, individual channel segments may be 100s metres – kilometres in length, whilst connecting to form more extensive networks (Greenwood et al., 2007; Kehew et al., 2012b). The scale envelope of subglacial channels is vast, the smallest of decimetre-metre scale, and their morphology, topology and their exposure in sediment sections indicate ephemeral, braiding systems in relatively small, disconnected patches (e.g. Walder and Hallet, 1979; Sharp et al., 1989; Piotrowski, 1999; Greenwood, 2008; Hughes, 2009). In contrast, subglacial channels on Antarctic palaeo-ice stream beds reach hundreds of metres depth (Nitsche et al., 2013), similar to the tunnel valleys 2-3 km wide, 100s m deep and up to 100 km long associated with near marginal drainage of the Fennoscandian and Laurentide Ice Sheets (Cutler et al., 2002; Sandersen et al., 2009; Stewart and Lonergan, 2011; Stewart et al., 2013). Whilst the large Antarctic channels cut into crystalline bedrock, the tunnel valleys at the margins of the palaeo-ice sheets of the northern hemisphere are mostly incised in sediments. The latter case might be associated with ‘canals’ theorised by Walder and Fowler (1994) and Ng (2000) while subglacial meltwater channels cut into bedrock might be more closely associated with Nye channels (see Section 2.2.3.).

Given the hidden nature of the contemporary subglacial environment and the range of scales concerned, it is difficult to find appropriate modern analogues or to witness active formation of the types of meltwater channels observed in the palaeo records. As a consequence, the mode of formation of these landforms remains uncertain (Ó Cofaigh, 1996; Kehew et al., 2012b), whether a product of steady-state down-cutting, a single high magnitude or multiple lower magnitude erosional events. Given such uncertainties, meltwater channels in the landform record have often been used as a simple indicator of palaeo-flow directions, or proxy for ice surface slope direction, rather than for specific hydrological processes.

At an intermediate scale, a number of sculpted forms (referred to as p-forms, plastically sculpted forms, Dahl, 1965; or s-forms, sculpted forms, Kor et al., 1991) appear to be the products of erosion by water flow, in cases pressurised (e.g., Gray, 1981; Shaw, 1988; Glasser and Nicholson, 1998). However, there is considerable uncertainty in these processes (Glasser and Bennett, 2004; Bennett and Glasser, 2009, p. 146-147; Benn and Evans, 2010, p. 272-277), and whether a primarily meltwater or ice flow origin is responsible.

Esker ridges are typically 10s-100s metres wide and 1-10s m high (Aylsworth and Shilts, 1989; Hebrand and Åmark, 1989; Syverson et al., 1994; Huddart et al., 1999). They occur in segments or ‘beads’ a few kilometres long, sometimes with interlocking ridges and often terminating in deltaic sediments (Warren and Ashley, 1994; Hebrand and Åmark, 1989; Mäkinen, 2003). Esker segments line up into systems that stretch – apparently continuously – for hundreds of km over the beds of the former northern hemispheric Pleistocene ice sheets (Prest et al., 1968; Aylsworth and Shilts, 1989; Clark et al., 2000; Boulton et al., 2001; Storrar et al., 2013; Storrar et al., 2014b; Fig. 4). Whether this reflects an extensive, continuous drainage network under large portions of the ice sheet, or incremental (time-transgressive) formation shortly behind a retreating margin, is a matter of some uncertainty.

Sediment architecture, esker geometry and distribution, and relationships to channel theory have been widely used to inform this debate. In the central parts of the Laurentide Ice Sheet, continuous ridges 10s-100s km in length have been observed, with fans or deltas only at their distal ends (Brennand, 1994; Brennand and Shaw, 1996; Brennand, 2000). Sediment clast characteristics documenting long transport distances and low palaeo-flow variability

(Brennand, 1994; Brennand and Shaw, 1996; Brennand, 2000) argue for esker operation at great lengths. Tributary connections of many esker networks (albeit low stream order; Storrar et al., 2014) further gives the impression of a single, continuous system. However, Clark and Walder (1994) argued that since channelised drainage should develop where the ice surface slope is relatively steep, eskers should form only close to the ice margin, and extensive networks must therefore be a time-transgressive product of a retreating margin. Hooke (2005) suggests that secondary ridges that branch and braid from the main stem are formed when the overburden pressure (ice thickness) is not sufficiently high to keep the water conduit on the crest of the primary sediment ridge; these systems are implied, therefore, to form relatively close to the margin. Beaded eskers in southern Sweden and south-western Finland support an incremental formation model (Hebrand and Åmark, 1989; Mäkinen, 2003), where the networks also closely fit with the regional end moraines (Punkari, 1997). The changing sediment fraction and the repeated successions of sedimentary units have been interpreted to document seasonal changes in sedimentation at the esker mouth (Mäkinen, 2003).

A link to supraglacial meltwater supply is implicit in a model interpreting seasonal sedimentation, a conclusion also reached by Storrar et al. (2014a) on the basis of the increasing frequency of eskers associated with warming and the deglaciation of the Laurentide Ice Sheet. Recent comparison of modelled subglacial meltwater pathways under the Laurentide Ice Sheet to the mapped esker networks found a poor match between the two, interpreted as an indication of a time-transgressive origin of the long esker systems and a supraglacial source of the meltwater drained (Livingstone et al., 2015). Sedimentation related to high discharges from either supraglacial or subglacial lake drainage events is interpreted by Burke et al. (2012a, b) for a pair of eskers in southern British Columbia. In a case study from southern Alberta, Burke et al. (2015) suggest that the shape, network complexity and the internal architecture of eskers ridges reflect both the flow power and sediment supply in the conduit, arguing that these inferred relationships might also be applicable elsewhere.

Esker properties have perhaps provided the closest links between the palaeo record and developing theory of ice sheet hydrological systems (e.g. Shreve, 1985; Clark and Walder, 1994; Boulton et al., 2007a,b). Despite incomplete understanding of their formation processes, eskers have traditionally been the landform of choice for broad-scale inferences on past ice sheet retreat patterns (De Geer, 1929; Dyke and Prest, 1987; Kleman et al., 1997; Boulton et al., 2001; Putkinen and Lunkka, 2008; Greenwood and Clark 2009a, b; Clark et al., 2012; Margold et al., 2013a, b).

Subglacial lakes are expected to leave a geological trace, but research into their identification in the palaeo-record is in its infancy. Livingstone et al. (2012) provide criteria, largely sedimentological, for palaeo-subglacial lake identification, and recent modelling efforts attempt to predict their location (Evatt et al., 2006; Livingstone et al., 2013). Such predictions test earlier hypothesised subglacial lake locations (e.g. Shoemaker, 1999; Munro-Stasiuk, 2003; Christoffersen et al., 2008; Lesemann and Brennand, 2009) and offer an improved picture of an integrated palaeo-glaciohydrological system.

4.3. Spatial distribution

The distribution of meltwater landforms on palaeo-ice sheet beds appears to be almost ubiquitous (Figs. 3, 4). The spatial distribution of meltwater landforms as a whole, and landform types as a subset, have been argued to be controlled by a number of factors; we consider (1) the coupled effects of subglacial thermal organisation and ice dynamics, and (2) the underlying geology.

4.3.1. Controls exerted by subglacial thermal organisation and ice dynamics

The thermal conditions at the bed fundamentally dictate patterns of and capacity for ice sheet drainage, and hence the distribution of the resultant glacial meltwater landforms. It has often been suggested that different landform types display zonation across the glacial landscape, and that the subglacial thermal organisation of ice sheets – and resultant ice dynamic regimes – controls such distribution patterns (Boulton, 1972; Sugden and John, 1976; Hall and Sugden, 1987; Dyke, 1993; Sollid and Sørbel, 1994; Kleman and Hättestrand, 1999; Kleman and Glasser, 2007). Assemblages of glacial landforms indicative of a well-lubricated or a stiff and frozen bed, and their spatial distribution, were used by Kleman and Glasser (2007) to conceptualise the subglacial thermal organisation of ice sheets as a series of warm-based corridors and intervening frozen zones.

That the meltwater record appears to be nearly ubiquitous at ice sheet scale points to the superimposition of deglacial mass melting signatures upon the earlier thermal organisation of the ice sheet. At a regional scale, the thermal history of the ice sheet may have a more significant role. Palaeo-ice streams on the Canadian Shield, identified on the basis of mega-scale glacial lineations and shear-margin moraines, operated during deglaciation in what had previously been a cold-based interior of the Laurentide Ice Sheet. Here, eskers have been found to be less frequent than in the surrounding areas (Kleman and Glasser, 2007; Storrar et al., 2013; Margold et al., 2015). It has been suggested that distributed meltwater drainage within the ice streaming zones may explain the lower frequency of eskers in these areas (Storrar et al., 2014b; Livingstone et al., 2015). Ice dynamics might affect not only the frequency of esker occurrence, but also their spatial pattern. While coherent and uniformly oriented esker networks, such as those in Keewatin, Labrador or parts of Scandinavia, have generally been interpreted as indicating frontally retreating ice sheet margins (though see Sections 4.2, 5.2 and 5.3), more chaotic esker patterns have variously been explained as a result of (1) a complex ice streaming history in a given area (Victoria Island; Storrar et al., 2014b), (2) interaction of a low-profile ice sheet with a variable topography (Canadian Arctic Archipelago; Storrar et al., 2014b), (3) subglacial drainage governed by local potentiometric surfaces in stagnant ice or subglacial drainage in stagnant ice fed into broader subglacial channels (southern Alberta, Brennand, 2000), or (4) by a transition from a coherent active ice surface to a more complex and stagnant ice body (Margold et al., 2013b).

Frozen areas of ice sheet beds have traditionally been identified by the presence of relict surfaces and landscapes, and an *absence* of landforms associated with subglacial melt delivery and soft-sediment deformation. Such sectors occupy considerable areas of ice sheets (Kleman and Glasser, 2007) and can be taken to indicate that we should not expect a (sub)glacial hydrological system to be widely developed in an ice sheet. Lateral meltwater channels have been considered indicative of cold-based or polythermal ice (Borgström, 1989; Dyke, 1990; Kleman et al., 1992; Dyke, 1993; Sollid and Sørbel, 1994). However, they have been reported to form at the margins of decaying warm-based Alaskan glaciers (Syverson and Mickelson, 2009), and surface-to-bed downcutting of a supraglacial dendritic network through the 'cut-and-closure' process (Vatne, 2001; Gulley et al., 2009) has been observed under cold-based glaciers in Svalbard (Naegeli et al., 2014). How the thermal condition of an ice sheet generates, directs and diverts the hydrological system is a fundamental question, but is non-trivial to extract from the landform record.

4.3.2. Geological controls

At a regional to ice sheet scale, eskers have been correlated with areas of hard bed lithologies, whilst incised channels occur where ice sheets covered less resistant bedrock or thick sedimentary sequences (Clark and Walder, 1994; Boulton et al., 2009). Soft, deformable beds are argued to favour distributed drainage through a network of shallow 'canals'. As a generality, these associations hold true for glaciated terrains of the northern hemisphere and, indeed, it has been argued that channel development at a local scale intimately follows local substrate geological variability (e.g. Sandersen and Jørgensen, 2012) and may be strongly controlled by the interaction with the local groundwater system (Piotrowski, 1997; Piotrowski et al., 2009). However, numerous local landform arrangements display complexity beyond the general landform–substrate associations. Many channel systems incise hard bedrock (Clapperton, 1968; Sugden et al., 1991; Nitsche et al., 2013; Jansen et al., 2014) whilst well-developed esker networks occur locally over thick sediments and sedimentary bedrock (Flint, 1930; Greenwood, 2008; Storrar et al., 2014b; Burke et al 2015). Furthermore, esker chains may transform into deeply cut channels, and vice-versa (e.g. Johansson, 1995, 2003; Hättestrand and Clark, 2006; Greenwood, 2008; Hughes, 2009; Margold et al., 2013a; Atkinson et al., 2014), whilst eskers are commonly observed on the floors of large tunnel valleys (see Kehew et al., 2012b). The landform record points to conditions conducive to both esker and channel formation within the same inter-connected system; the length-scale of synchronously operational networks is, however, a crucial unknown.

A different landsystem model has been advanced for the Antarctic continental shelves (e.g. Wellner et al., 2001, 2006; Graham et al., 2009), where the beds of former ice streams display a common landform succession from subglacial meltwater channels cut deeply into crystalline bedrock (Lowe and Anderson, 2002; Nitsche et al., 2013) to increasingly elongate glacial lineations across and beyond the seaward transition to sedimentary bedrock. Here, any traces of subglacial drainage disappear and it is hypothesised that meltwater is delivered through small canals or a deforming till aquifer (Ó Cofaigh et al., 2002; Graham et al., 2009; Noormets et al., 2009; cf. Murray et al., 2008). A single reported meltwater channel occurs in the outer shelf sedimentary environment of the Ross Sea (Alonso et al., 1992; Wellner et al., 2006) and eskers are notably absent. Meltwater landforms are not especially common in other marine glaciated environments, with the exception of some rare examples of eskers (e.g. Todd et al., 2007; Todd and Shaw, 2012; Feldens et al., 2013) and the tunnel valleys of the North and Barents Seas (Huuse and Lykke-Andersen, 2000; Stewart and Lonergan, 2011; Bjarnadóttir et al., 2012; Stewart et al., 2013). It should, however, be noted that that much research in marine environments has focused on palaeo-ice stream troughs, where fast flowing ice may impede channelised drainage. In contrast, the inter-ice stream banks are typically heavily scoured by icebergs, which might have destroyed any evidence of drainage.

4.4. Temporal properties

At an ice sheet scale, meltwater landforms in the palaeo-record are typically interpreted as a product of stable flow, and thereby used to reconstruct palaeoglaciological regimes corresponding to years, decades or longer (e.g. Kleman et al., 1992; Dyke 1993; Kleman et al., 1997; Clark et al., 2000; Boulton et al., 2001; Clark et al., 2012; Margold et al., 2013a, b). However there remains much uncertainty over the timescales of landform creation (Kehew et al., 2012b). Meltwater channel erosion may be a product of incremental but steady erosion (Boulton et al., 2009), single high-magnitude events (Brennand and Shaw 1994; Cutler et al., 2002; Hooke and Jennings, 2006), or repeated erosional episodes over time periods from ~100 years to multiple glaciations (Jørgensen and Sandersen, 2006; Sandersen et al., 2009; Nitsche et al., 2013). Sedimentological and morphological arguments are used to support each alternative and it is possible that all are viable models in different contexts. Contrary to the typical long-term view of palaeo glaciology, englacial eskers have been observed to form in single jökulhlaup events in Iceland (Russell et al., 2001; Burke et al., 2008). On the basis of its single, point source and its specific internal architecture, Burke et al. (2012a) interpret an esker in southern British Columbia as a possible evidence of supraglacial lake drainage, whilst the ‘Labyrinth’ network of massive channels and other high-magnitude flow traces in the Dry Valleys, Antarctica, have been interpreted as a result of subglacial floods with estimated discharges of $1.6–2.2 \times 10^6$ $m^3 s^{-1}$ and velocities of 11-15 m s^{-1} (Denton and Sugden, 2005; Lewis et al., 2006). Short duration and high-magnitude drainage events are likely to leave a geological and geomorphological product. Reliably distinguishing between such landforms and those formed under steady flow conditions remains a challenge for the palaeoglacial community.

4.5. Palaeo-ice sheet hydrology: summary

Thanks to the extensive occurrence of the meltwater landforms on the past ice sheet beds, the meltwater record has been widely used for palaeoglaciological reconstructions of Pleistocene ice sheets, but mainly as a simple indicator of palaeo-flow directions, ice surface slope directions, and ice margin positions. The mode of formation, especially for landforms of subglacial origin (subglacial meltwater channels, eskers), remains uncertain and different interpretative models have been developed for different regions subjected to past glaciations (cf. Brennand, 2000 and Mäkinen, 2003).

Meltwater landforms display a vast range of scales, contrasting topologies and arrangements (extensive esker networks, esker-channel-esker successions, chaotic distribution of channels). Landform creation is subject to processes beyond the glaciohydrological: processes of sediment mobilisation and deposition. Understanding of the operation of these processes and related controls on the landform distribution, such as the underlying geology (i.e. 'boundary conditions' in theoretical and numerical modelling approach) is incomplete. Similarly, the temporal and spatial scales of formation remain unclear: whether single or multiple drainage events, or gradual steady-state drainage; and how spatially connected an operational drainage network may be. Nevertheless, the palaeo record offers a spatial context which is difficult to extract from the contemporary systems: the whole ice sheet bed is exposed with the resulting landforms often easily accessible, which is in stark contrast to the indirect character of study of contemporary ice sheet hydrology.

Figure 5 summarises an ice sheet's hydrological system from a palaeoglaciological perspective, in terms of the geological record that we have today to decipher. This record is a composite resulting from different ice sheet stages over time and therefore also likely a composite of different drainage regimes. Although the glacial meltwater record as a whole might encompass tens to hundreds of thousands of years, individual systems within this likely represent a regime spanning only tens to thousands of years. The time associated with the processes responsible for individual landform creation might be even shorter, on the order of seasons to decades. Deciphering these temporal scales from a composite record of a whole glaciation (e.g. Figs. 3-5) and, furthermore, bridging this information with the timeframes represented in Figure 1, presents a serious challenge.

5. Discussion

A driving agenda for glaciohydrological research is to better understand the dynamic responses of ice flow to hydrological processes, and to better predict likely responses of ice sheets and ice sheet sectors to increased surface melting under atmospheric warming scenarios. To this end, we need a clearer understanding of the dominant form of the meltwater drainage system, how pressure and discharge are related, and under what physical conditions meltwater is delivered through the glacial hydrological system. Based on our preceding reviews, it is clear that the existence of one single mode of meltwater drainage in an ice sheet is implausible. A key objective for the glaciohydrological community must be to determine over which spaces, and which timescales, different drainage modes operate.

Achieving this goal is complicated by the fact that i) the physical processes of water flow in an ice sheet, ii) spatial variability and spatial controls upon water drainage, and iii) temporal variability and controls on drainage, are each uncertain to some degree, may appear to be contradictory between different fields of research, and are all intimately interlinked and interdependent. Our review presents three major challenges to which we draw attention:

1. Contemporary observations and glaciological theory suggest that the occurrence of low-pressure, channelised drainage (cf. Röthlisberger, 1972) should be relatively limited. The palaeo record contains widespread evidence of channelised drainage, of highly varied topologies. Moreover:
    - Long-distance operational connectivity of a channelised system has been interpreted from some palaeo-ice sheet beds and long-distance operational connectivity has also been found in contemporary Antarctica.
    - The apparent lack of a widespread 'distributed' system in the palaeo record may be a product of poor preservation, or poorly developed tools and templates for identification and interpretation of non-channelised drainage.
2. The palaeo record is a cumulative product of formative processes and environments. However, understanding of the timescales of landform development is limited, as is understanding of the potential of hydrological regimes of different magnitudes for landform creation. Conversely, events of different magnitudes and timescales are known from contemporary ice sheets, but their long-term significance is poorly understood.
3. Fundamental scale contrasts between research fields are a major disconnect, both in spatial and temporal scale. A 'large channel' to a contemporary glaciologist might be 10-100 m wide; to a palaeoglaciologist a large channel would be 1-2 km in width.

In the context of these challenges, we consider below the form, spatial extent and variability, and temporal controls on the drainage system. These issues are highly interconnected and cannot wholly be treated apart. We highlight the sensitivity of the spatial extent of subglacial channel maintenance to temporal variability. We emphasise a need to understand geomorphological as well as hydrological processes in order to make robust use of the glaciofluvial landform record, and we consider the degree of consistency of the ice sheet drainage system across geographical and geological domains. Our discussion is stimulated by our varied perspectives and we attempt to present a cross-disciplinary view of the controls, distribution and evolution of subglacial drainage systems.

5.1. What is the form of the drainage system?

Theoretical glaciology, and much of the language employed concerning ice sheet hydrology, conceptualises two contrasting frameworks which describe how water might be delivered through the subglacial domain (e.g. Röthlisberger, 1972; Kamb, 1987): a channelised, or a distributed system. These are respectively associated with efficient, low pressure conduits capturing flow into a dendritic network, and a poorly connected network of water-filled voids at high pressure; simplistically, theory predicts two contrasting drainage topologies.

Support for a widespread, low pressure Röthlisberger channelised system does not appear to be emerging from studies of contemporary ice sheets. Efficient meltwater delivery around the

Greenland ice margin is assumed to reflect channelisation *close* to the margin (Bartholomew et al., 2011a; Chandler et al., 2013), the interior comprising a high pressure domain (Meierbachtol et al., 2013) consistent with the high pressure basal environment inferred to be widespread in Antarctica (Alley et al., 1986; Engelhardt et al., 1990; King et al., 2004). This is in stark contrast to the palaeo record of Quaternary glaciations, in which there is almost ubiquitous evidence of the channelisation of subglacial water. The topologies, however, of both eskers and palaeo-meltwater channels are highly varied (Figs. 3, 4) and it is by no means clear that (all) these landforms should represent low pressure Röthlisberger drainage. Any disconnect between modern and palaeo evidence of subglacial drainage should be perceived in the context of the high degree of uncertainty as to whether observations taken to indicate a Röthlisberger-type network (low pressure, dendritic) should indeed be interpreted as such.

On modern ice sheets, interpretation of the *form* of a drainage network cannot be made with confidence based only on the *efficiency* and speed of transit (cf. Gulley et al., 2012b; Cowton et al., 2013); when all that can be monitored is the input and output, the exact pathways and modes of delivery remain unknown. Indeed, emerging evidence suggests that efficient drainage could be viable in high pressure settings (Fudge et al., 2008). The widespread channelisation of meltwater indicated by the palaeo record comprises highly varied forms and network topologies. It is widely assumed that eskers reflect a Röthlisberger system and a broadly dendritic arrangement is often taken as support, though the theory has no explicit prediction of stream ordering, merely that water should be conveyed from high to low pressure. Modifications to classic Röthlisberger theory (e.g. Hooke et al., 1990) allow for non-dendritic channelised topologies and, indeed, eskers exhibit contrasting topologies depending on the scale at which they are viewed (Fig. 4A). Hewitt’s (2011) implementation of channel theory predicts lateral channel spacing in broad agreement with that of Laurentide Ice Sheet eskers documented by Storrar et al. (2013). Canals and Nye channels have typically been considered part of an inefficient (slow) system of drainage (e.g. Walder and Fowler, 1994; Fountain and Walder, 1998). This could be consistent with the varied topologies of palaeo channels: often discontinuous, branching and chaotic in contrast to the form of esker networks (Fig. 3A, 4A). However, it is difficult to explain large palaeo channels that clearly have been eroded by high-discharge subglacial streams (e.g. Booth and Hallet, 1993; Lowe and Anderson, 2003; Nitsche et al., 2013; Jansen et al., 2014) as part of inefficient, poorly connected drainage, and it is similarly problematic to account for intermittent esker – channel – esker connectivity over long distances.

Further, there is an *apparent* lack of distributed drainage in the palaeo record. This is arguably a consequence of poor preservation under a retreating marginal zone (see Fig. 5), of inadequate templates for its identification and interpretation in the landform record, and of the immense challenges in extending sedimentological observations of distributed drainage to the ice sheet scale. Finally, the appearance of ubiquity is likely to a large degree an artefact of a time composite record. The geological record is, inherently, an amalgam of processes and products over an extended time domain and, in the case of eskers and meltwater channel landforms, arguably a product of ablation and ice sheet deglaciation. Any apparent mismatch in drainage topology between contemporary and palaeo perspectives is therefore, in part, a likely result of the contrasting time domains and 'life-cycle' stages of the respective ice sheets.

Figure 6 summarises a variety of drainage topologies which are supported, to varying degrees, by observations from both modern and palaeo environments and implied by glacial hydrology theory. Whilst these conceptual sketches focus on the 'shape' the drainage takes, topology can be considered a simplified proxy for physical process descriptions (see Section 2.2). We suggest that each of these topologies (and possibly more) may be viable and appropriate in different contexts, and argue that it is more instructive to conceptualise the subglacial system as such a multiplicity of modes, rather than the traditional binary channelised (low pressure, dendritic) or distributed (high pressure, inefficient) modes. The outstanding challenge, then, is to assign a spatial scale to these multiple modes and determine their spatial and temporal occurrence.

5.2. Spatial extent and variability of effective drainage

It is impossible to determine the relative dominance of any of the topologies (and respective drainage processes) illustrated in Figure 6 without a sense of scale. If one were to assume that eskers are the best representative of the dominant subglacial hydrological configuration, and if one were to adopt a regional perspective, one might conclude that topology 6A (Fig. 6) – i.e. a low-order dendritic, low pressure flow-capturing channel system – were the normal mode of drainage. Esker formation, however, is yet to be observed on our modern templates for the system – Greenland and Antarctica – and their formative scales are equivocal. If we alternatively consider the palaeo-meltwater channel record then topology 6B is more typical. Change the spatial scale of focus (see the boxed areas on Figure 4, for example) and topologies 6C or 6D would appear more appropriate.

Given that the full palaeo-ice sheet bed is now exposed, the palaeo record might be best placed to enlighten the spatial scales of different drainage modes. However, any such attempt is thwarted by the difficulties of determining the temporal controls on landform generation, and the time domain represented by the landforms we now observe. There is an order of magnitude difference in inferred operational length of subglacial conduits depending on whether one adopts a continuous tunnel fill model (e.g. Brennand, 1994; Brennand and Shaw, 1996; Brennand, 2000), such as for the Laurentide systems hundreds of kilometres in length, or a time-transgressive beaded esker model (e.g. Hebrand and Åmark, 1989; Mäkinen, 2003) in which channelised drainage is confined to a marginal zone a few 10s km at most. The former invokes a wide ablation zone with high surface melt rates, typical of the last deglaciation (e.g. Carlson et al., 2008, 2009) and which may therefore account for a contrasting length scale to modern observations. However, Livingstone et al. (2015) find that modelled subglacial pathways routed according to Shreve (1972; Section 2, Eqs. 1, 2) provide a robust match with observed esker geometry only up to ~10-20 km from the margin. Beyond this distance, a small minority of eskers match modelled subglacial pathways for >100 km. The majority of apparently lengthy esker networks deviate from modelled channel routing, suggesting ice sheet geometry (hydraulic gradient) has changed and that these long-distance networks must be formed time-transgressively.

Modern glaciological observations and current theory rather support limited channel lengths, with an implied maximum distance of efficient subglacial drainage of about 50 km (Bartholomew et al., 2011a; Chandler et al., 2013), in the marginal zone where there is a sufficiently steep ice surface slope (cf. Clark and Walder, 1994; Walder and Fowler, 1994; Meierbachtol et al., 2013). Recent modelling by Meierbachtol et al. (2013) argued that even with high and sustained meltwater input in the interior (cf. Brennand, 2000; Carlson et al., 2008, 2009), the inhibited melting back of subglacial conduit walls is not sufficient to overcome creep closure; rather, highly pressurised water would be driven into a surrounding distributed system and the conduit would not be sustained. A related approach by Dow et al. (2014) models subglacial conduits under high pressure ~70 km from the Greenland Ice Sheet margin, and finds that conduit maintenance is improbable under these conditions.

The operational length of subglacial channels remains a significant uncertainty. We must consider the potential that deglaciation of mid-latitude ice sheets provided different hydrological conditions to today and that our modern ice sheet observations may, in part, be inappropriate analogues. The physics presented in Röthlisberger (1972; i.e. transient forms of Eq. 5, Section 2.2.1) underlie numerous theoretical developments and numerical explorations of subglacial channelisation (e.g. Hooke et al., 1990; Walder and Fowler, 1994; Clarke, 1996; Schoof, 2010; Meierbachtol et al., 2013). We can explore conceptual scenarios for subglacial channel extent using a simple numerical model (e.g. Meierbachtol et al., 2013) of a straight channel over a hard bed (with no explicit physics for what happens outside the bounds of Röthlisberger channelisation). This provides a ‘favourable case’ for channel maintenance, making no attempt to address how a drainage system evolves under different regimes, only to illustrate at what point conventional channel theory becomes insufficient such that an additional or alternative physical system is needed. Under steady-state conditions with a constant meltwater input, this model shows that subglacial conduit water pressures lie below overburden pressure for any input discharge and *any given distance* from the ice margin under a range of ice surface profiles (Fig. 7; cf Meierbachtol et al., 2013; Helanow et al., 2015; and demonstrated in Supplementary Material). In the strictly steady-state case, a conduit could be maintained 10s-100s km up ice from the margin.

However, even small transient increases in water input drive subglacial water pressures towards ice overburden and beyond the limits of Röthlisberger physics (Meierbachtol et al., 2013); conduit wall melt-back cannot keep pace with an increase in discharge. If a conduit survives at high pressures, a dendritic, flow capturing network is unlikely to develop, since any flow across a pressure gradient would only serve to reverse the differential and drive flow back where it came from. *Temporal* controls on the system therefore appear to be paramount. The same model (see Supplementary Material) indicates that a low surface profile typical of outlets of the Laurentide Ice Sheet (e.g. Beget, 1986; Clark, 1992) may in fact sustain a lower pressure channel than a high profile (e.g. Greenland Ice Sheet) at ~40 km from the ice margin. We could, therefore, anticipate some divergence between our landform and contemporary indications of channel length. At greater distances, however, channel maintenance is significantly more difficult under low surface profiles. This is in partial keeping with the conventional view that channelisation is best developed under higher surface slopes (Röthlisberger, 1972; Walder and Fowler, 1994), but at a certain proximity to the margin, a slow creep closure rate under thin ice becomes significant. We finally draw attention to the role of conduit size, which, for example in this simple Röthlisberger-based model, does not diminish monotonically upstream but rather narrows to a knickpoint part way along its viable (< overburden pressure) length (Fig. 7). The sensitivity of any conduit constrictions to reaching overburden pressure will arguably be most important in governing the conduit's stability and, potentially, the length of a persistent channel that may leave a recognisable landform imprint. We note that with a range of different channel length scenarios (10s-100s km) this knickpoint lies c. 20-50 km upstream from the ice margin.

We consider that the weight of evidence supports a spatial extent of a dendritic channelised topology (i.e. classic Röthlisberger, Fig. 6A) limited to a few 10s km, likely <50km. However, we do not exclude the possibility of significant efficient drainage in some form of channel system at greater spatial scales inward of this limit. Our present understanding of such drainage is that it would be at relatively high pressures and in a likely non-dendritic system. We highlight three outstanding challenges in constraining the spatial extent of different drainage modes:

i. The topology and processes of significant drainage in a high pressure (at/near overburden) setting are poorly understood, encompassing a wide variety of process models (Section 2.2.2; e.g. Weertman, 1972; Kamb, 1987; Walder and Fowler, 1994; Creyts and Schoof, 2009). Targeted, systematic and repeat geophysical monitoring of different hypothesised drainage environments in both Greenland and Antarctica, for example saturated sedimentary domains, regions of supraglacial melt penetration to the bed, and subglacial lake drainage routes, will be profitable in better constraining high pressure drainage modes.

ii. The spatial extent of subglacial drainage must not only consider individual systems, but also the overall distribution of where drainage does, or does not operate at all: i.e. the (subglacial) thermal organisation of the whole ice sheet and the supply of meltwater. The thermal state of an ice sheet bed comprises a complex spatial pattern that evolves over time (Joughin et al 2004; Hubbard et al., 2009; Pattyn, 2010), leading to a time-transgressive composite being preserved in the palaeo record (Kleman and Glasser, 2007). The extent and magnitude of subglacial melt and drainage at any one snapshot in time may be significantly less than the impression given by the time-integrated palaeo-landform record.

iii. Conversely, the palaeo record reflects not only hydrological processes, but also geomorphic and sedimentological. Lack of geomorphic power or sediment starvation may mean a period of ice sheet drainage has no legacy, and our geological record under-represents the palaeo hydrological system (cf Burke et al., 2015).

5.3. Temporal variability and controls on drainage

5.3.1 Transient meltwater regimes

The significance of the time domain has been raised in a number of contexts throughout our reviews and the Discussion so far, both in relation to the temporal variability of meltwater drainage and in relation to the time composite landform record. In its pure form, Röthlisberger theory illustrates that channel maintenance 100s km from the ice margin is feasible under truly steady input regimes (Fig. 7 and Supplementary Material). However, it is difficult to envisage a requisite steady input setting to maintain a conduit system below overburden pressure. Basal melt dominated systems, such as in Antarctica where there is no/little surface input, might be such an environment where short term temporal variability is removed, though the flux is so low (e.g. basal melt of 1-10 mm a^{-1}, Joughin et al., 2004; discharge across the entire Siple Coast 300 m^3s^{-1}, Carter and Fricker, 2012) that it is difficult to explain large-scale channelisation in this context. Schoof (2010) also suggests that below a critical discharge, channels would not be initiated, but rather that the system would stay in a distributed form. Alternatively, retention of surface meltwater in supra-, en- and subglacial stores (Willis et al., 1990; Fountain, 1993; Harper et al., 2010; Cowton et al., 2013) may deliver water to relevant points of the bed at a constant rate. This, together with a surface melt rate so high that diurnal and seasonal contrasts are significantly dampened, could perhaps elicit subglacial conduit development in a pseudo steady-state fashion. Carlson et al. (2009) model palaeo-surface mass balance for the Laurentide Ice Sheet and suggest that at the height of deglaciation, parts of the southern margin were thinning at rates of 9 m yr^{-1}. It is possible that such extreme scenarios are not well covered by current drainage theory, and that we are consequently not even conceptually familiar with a resultant drainage system. It is such mass melting environments with which long esker systems have been traditionally associated (Brennand, 2000).

Most realistic systems are, however, transient: supraglacial meltwater dominated systems vary diurnally, are highly seasonal with an annual winter shut-down and spring restart, and exhibit multi-annual variability. Moreover, both supraglacial and subglacial systems experience episodic events such as lake drainages superimposed on other more regular cycles. Following Meierbachtol et al. (2013), we find that only very low daily and seasonal fluctuations in meltwater input (e.g. 2% daily, 20% seasonal) are able to yield a conduit pressure response akin to a steady regime and stay below overburden pressure (albeit at a high fraction; see Supplementary Material). At 40 km, overburden is reached when daily oscillations in input exceed ~14% of initial discharge; this amplitude threshold for reaching overburden is even lower at greater distances. Sensitivity to transience in meltwater supply is increased, therefore, with distance from the ice margin: there is a dependence of *spatial* extent and spatial stability on the *temporal* properties of the system, and all the more so under low ice surface profiles.

If transience in the supply of meltwater to the subglacial system inevitably drives water pressures towards overburden, then it is clear that one cannot assume that an 'average' of transient fluctuations approximates a steady-state, and that addressing aspects of ice sheet hydrology on palaeo or multi-annual timescales cannot ignore short-term processes. It implies also that an incipient conduit up-ice from the marginal zone needs *some* supply, constantly, to be sustained. Mass melting of the Quaternary ice sheets (e.g. Carlson et al., 2009) could perhaps enable this, but today winter closure would preclude any interior conduit maintenance under this model (cf. Gulley et al., 2009a; Jarosch and Gudmundsson, 2012). This could go some way to offering an explanation for the contrasting evidence of conduits between the landform record of deglaciation and modern observations.

5.3.2 Transience in geomorphology

How transience in meltwater input is manifested in the landform record of the subglacial system is far from clear. The relative importance of different timescales of drainage for landform building is largely unknown from either a process or palaeo perspective. An attempt to better constrain the timeframes represented by glaciofluvial landforms, and whether a certain landform type or network therefore reflects a more common or rare state of the palaeo-drainage system, is urgently required. A model for long, continuous esker systems (e.g. Brennand, 2000) would call for gradual sedimentation along their whole length. Röthlisberger theory indicates that such a conduit is very unstable, would exist at or close to overburden pressure and is only possible under a permanent, high flux as close to steady as possible (cf Fig. 7). Eskers would then reflect one single ice sheet state and stability of the hydrological system for the duration of esker building. Repeated exploitation of the same supra- and englacial pathways season after season (Fountain et al., 2005a, b; Catania et al., 2008; Benn et al., 2009; Catania and Neumann, 2010) make such stability a possibility, though these observations are not from great distances from the ice margin. However, development of a dendritic network, the very characteristic often used to support an extensive operational network, does not seem physically plausible under the predicted high pressure setting. The transience of a realistic meltwater input is more consistent with the alternative end-member model of esker formation, whereby sedimentation is confined to short (a few hundred metres long), thick beads at the ice front (e.g. Mäkinen, 2003). Sediment architecture in such segments argues for highly seasonal drainage, with high spring and summer sedimentation rates, as well as positional stability of the drainage pathways over successive seasons (Mäkinen, 2003; Ahokangas and Mäkinen, 2014). An outstanding question relates to the source of these high sediment volumes: a transport mechanism (by water?) is required beyond the confines of the esker deposit itself, implying greater extent of the hydrological system than manifested by the depositional landform.

Ultimately we must determine under which meltwater regimes landform development occurs. The extent to which we can attribute landform creation to stable seasonal drainage evolution, to annually occurring or repeated episodic events, to semi-independent series of inter-linked events, to isolated catastrophic outburst events, or even to multiple glaciations is palpably unclear (cf. Ó Cofaigh, 1996; Kehew et al., 2012b). We may expect higher magnitude events to have stronger geomorphic power (cf. Hjulström, 1935); the question is whether infrequent events of high magnitude are more geomorphologically effective than those of more moderate magnitude and recurrence (cf. Wolman and Miller, 1960). Could our landform record of palaeo-ice sheets in fact reflect only short-term, one-off high magnitude events (Figs. 8, 9)? Supraglacial lake drainages with discharge magnitudes on the order of 10 000 $m^3 s^{-1}$ (Das et al., 2008) are both seasonal and episodic in occurrence, following the seasonal up-ice expansion of ablation, whilst the timing and frequency of individual lake drainages varies inter-annually (Fitzpatrick et al., 2014). Antarctica's subglacial lakes are known to burst episodically, some draining and re-filling over sub-decadal timescales (e.g. Wingham et al., 2006; Fricker et al., 2007) but with any cyclicity likely independent of wider climatic or seasonal forcing. Meltwater landform corridors containing plunge pools and point-sourced eskers (e.g. Rampton, 2000; Burke et al., 2012a, b) and deeply incised bedrock gorges (e.g. Jansen et al., 2014) attest to sudden fluxes of water to and along the bed, and have been attributed to supraglacial events. Burke et al. (2008, 2010) interpret distinct esker sediment packages as products of subglacial outburst flooding, whilst large palaeo-channel systems such as the Labyrinth channels have been linked to catastrophic subglacial outbursts (Denton and Sugden, 2005; Lewis et al., 2006). The landform record of palaeo-ice sheets undoubtedly contains traces of high magnitude events. What, then, in our landform record is the imprint of the exceptional, and what of the normal?

If we were to suppose that the 'normal' condition of subglacial hydrology would not form landforms, but these were identifiable as a product of cumulative exceptional events, then our landform record would not be representative of the dominant mode of ice sheet drainage. We may envisage a scenario where, despite the apparent ubiquity in the palaeo record of some kind of channelised drainage, a topology somewhat akin to Fig. 6E (largely distributed, disconnected and small-scale) dominates, but our palaeo record is skewed by the higher landform creation potential of systems looking more like 6A-D which occur only temporarily and sporadically. Figure 9 conceptualises the uncertainty regarding landforming potential in response to meltwater drainage. Not only can various different drainage regimes (e.g. marked i-iii on Fig. 9) deliver a long term discharge that is difficult to discern from a stable regime (iv), but it is uncertain where, within such drainage regimes, most geomorphic work may take place. Does most landform creation (red dashed lines on Fig. 9) only occur during infrequent high discharge events (e.g. scenario E, Fig. 9) or during more regular, moderate discharge cycles (e.g. P_{max}, P_{min})? Or, alternatively, does the landform record simply represent an ‘average’ condition (scenario A)? The uncertainty in the process-product relationship makes interpretation of the timescales represented by the landform imprint challenging. These uncertainties are further compounded if we consider the opposing directions of geomorphic work: erosion or deposition. In using an esker as a proxy for palaeo drainage, our product only represents the depositional phase of that drainage, and we assume that its distribution is limited only by the operation of the hydrological system and not by the geomorphic. Sediment starvation would preclude depositional landform development (cf Burke et al., 2015) just as effectively as a drainage regime either too weak to drive conduit development or too powerful to deposit its sediment load. Our geomorphic products are indirect proxies that are more spatially and temporally restricted than their parent hydrological system. Resolving the timeframes and discharge conditions responsible for geomorphic work is paramount to our ability to use the palaeo record as proxies of palaeo-ice sheet hydrology.

5.4. Scales of internal consistency and variability

At local scale, the landform record displays varied and irregular patterns. This leads us to ask whether the fundamental modes of drainage vary on these spatial scales, or whether it is simply the landform expression that changes. Varied assemblages of meltwater landforms occur within small areas (e.g. Hättestrand, 1998; Hättestrand and Clark, 2006; Margold et al., 2013a; Atkinson et al., 2014; Turner et al., 2014), with esker–channel–esker downstream sequences (Johansson, 2003; Hättestrand and Clark, 2006; Greenwood, 2008; Hughes, 2009; Margold et al., 2013a; Atkinson et al., 2014), eskers on the floor of large channels (Hättestrand and Clark, 2006; Kehew et al., 2012b), or meltwater corridors containing an assortment of landforms (Rampton, 2000; Burke et al., 2012b). In light of this, the topologies, or modes of drainage, represented in Figure 6 appear to be gross simplifications. We could draw two contrasting conclusions from the varied expression of the landform record regarding system functionality. Local controls such as micro-topography, substrate properties or discharge availability could drive a switch to an alternative hydrological system: the downstream change in landforms would reflect downstream (or temporal) changes in the fundamental mode of meltwater drainage. Alternatively, the *same* hydrological system passing over/through varying external boundary controls would simply be manifested in different ways in the landform imprint.

Physically, one might expect there to be fundamentally different systems on different substrates (cf Clark and Walder, 1994), i.e. the former of the two above scenarios. If different landforms and landform configurations reflect different levels of efficiency (e.g. eskers as ‘fast drainage’; palaeo channels as ‘slow systems’), then in a continuous functional network (assumed) how would water be conveyed through an inefficient system situated in between two efficient systems? Alternatively, different landforms should be interpreted as a reflection of geomorphological rather than hydrological efficacy. Recent theoretical efforts attempt to integrate systems of different efficiencies, but the focus of theory has not hitherto been to investigate the influence of different boundary controls and the effect of boundary transitions. Thus far we lack clear understanding of the independent role of separate boundary conditions (e.g. sediment, soft bedrock or hard bedrock). It is therefore difficult to say from a theoretical point of view how these components might integrate within a unifying or overriding system.

Regional boundary controls can be expected to impact both glaciohydrological processes and the efficacy and form of geomorphic work, therefore dictating both local and short-term operability and the regional-to-continental scale organisation of drainage within an ice sheet. Substrate geology and topography (e.g. Clark and Walder, 1994), sediment fluxes (e.g. Burke et al., 2015), subglacial thermal organisation (e.g. Joughin et al., 2004; Kleman and Glasser, 2007), climatically driven surface melt availability (e.g. Carlson et al., 2009; Storrar et al., 2014a) and the mode and stability of regional ice surface to bed connectivity undoubtedly influence the functionality of the subglacial system, its large-scale spatial variability, and our ability to use landforms as proxies for palaeo-ice sheet hydrology. To some degree, boundary controls are specific to an individual ice sheet's geography or stage of evolution. However, the large scale patterns of the palaeo record attest to some element of self-organisation of the ice sheet hydrological system. At an ice sheet-wide scale, the spatial distribution, arrangement and topology of esker and palaeo-channel networks are similar between the different northern hemisphere palaeo-ice sheet beds (Fig. 10) suggesting that they can be treated similarly. Different landsystem expressions of the hydrological system may correspond to different regional controls but each unit still displays a degree of internal organisation or regularity, and the spatial scale of this organisation is qualitatively similar (Fig. 10). There thus appears to be a level of system control, regardless of the local external boundary controls. Whether the hydrological system is truly self-organising, or is perhaps rather a response to or reflection of self-organisation of ice flow, taps into one of the fundamental uncertainties in present understanding of ice sheet behaviour.

5.5. Subglacial systems: controls, distribution and evolution

In attempting to understand the processes, controls upon, variability of and products of the glaciohydrological system, there are clearly numerous domains we must bridge: spatial and temporal domains, supraglacial and subglacial environments, palaeo-ice sheet records and modern ice sheet observations. There is considerable difficulty in doing so: contrast again the timescales, the research logistics and research approaches represented in Figures 1 and 5. Figure 11 conceptualises an ice sheet's first-order hydrological characteristics in both space and time domains. A high pressure environment is found throughout the ice sheet, and may take on several forms with different levels of efficiency (Fig. 6); moderately efficient drainage could occur through a system close to overburden pressure. Röthlisberger-type channelised drainage develops in an outer zone (C in Fig. 11) whose inward limit is governed by ice surface slope, discharge and the temporal variability of surface melt delivery to the bed, and whose wider spatial distribution must reflect the thermal state of the ice sheet and patterns of organisation therein. Pockets of efficient delivery may occur within the interior high pressure zone (cf. Antarctic drainage between lake basins) whilst short-term high magnitude events occur throughout the ice sheet's history and likely increase through deglaciation, though it is not yet clear how far and for how long these could expand a zone of channelised drainage.

These are likely to be especially capable of geomorphic work, but the most effective discharge regimes for erosional and depositional landform creation, and the degree to which these are influenced by local boundary controls, remain uncertain.

The outer, channelised zone (C) occupies a similar subglacial envelope as does the ablation zone in the supraglacial domain; systems of the channelised domain are generally considered supraglacial melt dominated (Bartholomew et al., 2011b; Cowton et al., 2013). Whilst the ice surface slope and stress field are fundamental in meltwater routing, many theoretical formulations argue that discharge regimes are a critical parameter in drainage system evolution (see Section 2), and will inevitably govern the efficacy of the geomorphic processes of erosion and deposition. In the supraglacial melt zone, both the background discharge and superimposed event frequency are high. However, what actually happens to water at its point of entry to the bed is still little understood. What is the immediate response of subglacial pressure and drainage topology to supraglacial water input? If conduit growth is inhibited under overburden pressures then stable channelised systems will be restricted to the near-marginal zone irrespective of surface melt discharges: enhanced areal supply of meltwater through deglaciation would not drive a commensurate expansion of the subglacial channelised zone. Alternatively, if efficient drainage can operate at or close to overburden conditions, then supraglacial control over the subglacial domain may remain significant throughout the ice sheet cycle. In the Antarctic case, where supraglacial melting is negligible, the subglacial system must be controlled by a different suite of mechanisms and parameters to those of supraglacially-dominated systems. A minimum, continual supply of water to the bed appears to be necessary to keep a channel open and operational (Fig. 7C). Antarctica may be better represented by an almost entirely high pressure zone (absence of C-zone; cf. Fig. 6E), with some internal pockets of focussed drainage, likely also at high pressure.

We stress, finally, that there are stark differences in the climatic, topographic, 'life-cycle stage' and ice dynamic settings of the Greenland, Antarctic and northern hemisphere palaeo-ice sheets. Meltwater of supraglacial origin likely constitutes a significant portion of meltwater drained through both the present day Greenland Ice Sheet and the deglaciating Quaternary ice sheets, dominating discharge regimes, while negligible surface melt in Antarctica likely produces a distinctly different hydrology. Furthermore, the hydrology of the ephemeral Pleistocene ice sheets may have changed substantially during their lifetime, resembling present-day Antarctica during the cold climates of growth and Last Glacial Maximum phases, whereas the conditions during deglaciation may have resembled an extreme version of what can nowadays be observed in SW Greenland. Since the palaeo record is a time-integrated composite, it is possible there is a legacy of all these states, and it remains possible that none is an appropriate analogue for today's remnant interglacial ice sheets. Moreover, it is clear that the scales exhibited within the palaeo record of ice sheet drainage are in considerable excess of those evident today, in terms of landform size (cross-section, amplitude), spatial extent and in the temporal domain. The significance of the scale mismatch, and the processes responsible for the palaeo record, are yet to be determined. We argue that there is a degree of self-organisation of the drainage system and this holds between different ice sheets (Section 5.4), albeit on the continental spatial scale and a glacial cycle temporal scale. The field can and should advance by learning from palaeo and modern observational analogues for ice sheet hydrology, but should treat these analogues with care and caution.

6. Conclusions

Based on our reviews of theoretical, contemporary observational and palaeo glaciology, we have attempted to highlight how understanding of ice sheet hydrology can be advanced by a cross-disciplinary approach. Isolated study by different communities working with different perspectives risks that objectives become niche and divorced from one another. We have attempted to identify and discuss some of the main deficiencies in the contemporary understanding of ice sheet hydrology and we offer our key conclusions here.

Drainage topologies and inferred pressure relationships in both modern and palaeo-ice sheet environments differ from a traditional binary conceptualisation of low pressure, dendritic channels versus a high pressure distributed system. Instead, we consider drainage topology to comprise a multiplicity of forms in an amalgam of drainage modes and processes, each of which may be viable and appropriate in different contexts and at different scales. Greenland and Antarctic observations indicate efficient drainage at high pressure, and we suggest that particular attention should be turned to determining topologies, drainage processes and geomorphological products of these regimes.

These same field observations and numerical modelling suggest that low pressure settings are confined close to the margin (< ~50 km), and that the spatial extent of a classical channelised drainage system may be highly sensitive to the temporal variability in meltwater supply. One cannot, therefore, neglect short-term processes even when considering a long-term perspective. A further implication is that transient esker sedimentation likely reflects a position close to the ice margin.

Opposing interpretative models for glacial landforms, apparent contrasts with modern ice sheet analogues, and the composite, cumulative nature of the record of Pleistocene ice sheets hinder robust interpretations of palaeo-hydrological systems. It remains poorly understood which scales, magnitudes and frequencies of drainage events are most important for geomorphic work in the glacial hydrological system, and therefore which elements of the palaeo system are imprinted in our landform record: the normal or the exceptional. Furthermore, we must determine whether different (subglacial) landforms represent fundamentally different hydrological modes, or whether they are simply a different landform expression of a unifying hydrological system. The uniformity of spatial patterns across palaeo-ice sheet beds argues for a degree of self-organisation of the ice sheet hydrological system independent of any local or regional boundary conditions. There is a real need for further geomorphic process research, with a physical process basis and which considers the bed as an active participant in glacial hydrology, independent of the palaeo-ice sheet reconstruction paradigms with which glaciofluvial landforms have often been associated.

Glacial hydrology is influenced by the climatic, topographic, 'life-cycle stage' and ice dynamic settings of each particular ice sheet. Full physical description of meltwater drainage modes and regimes, and of their interactivity, is needed for the evaluation of ice sheet behaviour on different temporal scales, both in interpretation of palaeo ice dynamics and for predicting ice sheet evolution in a future warming world. Linking the processes to the products of ice sheet hydrology remains challenging, but vital to these endeavours.

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Figure captions

Figure 1. *(Portrait, full page width)*

Conceptual diagram of a glacial hydrological system, comprising the supraglacial, englacial and subglacial environments. Note that not all components may be applicable in all ice bodies. Insets depict: A) Water-driven crevasse propagation following van der Veen (2007). B) Proposed form of Röthlisberger, Hooke and Nye subglacial channels, where darker colours represent higher discharge. C) Main physical factors influencing the opening or closure of a subglacial channel, where p_i and p_w are the pressure of ice and water, respectively, and Q is discharge through the channel. D) Cross-section of accumulation zone facies (following Nolin and Payne, 2007).

Figure 2. *(Portrait, full page width)*
Contemporary observations of englacial and subglacial drainage systems. A) Moulin entrance on Leverett Glacier, southwest Greenland Ice Sheet, with inflow of supraglacial meltwater into the moulin. B) Englacial conduit c.60 m below the ice surface originating from a longitudinal crevasse on Matanuska Glacier, Alaska. Note the channel wall roughness. C) Cut and closure englacial conduit on Longyearbreen, Svalbard at c. 15 m below the ice surface, where the upper channel walls undergo closure due to ice pressure. D) Subglacial (Röthlisberger-style) channel c. 60 m below the ice surface of Hansbreen, Svalbard. Note the large-scale roughness of channel ice walls. E) Subglacial (Nye-style) channel under Hansbreen, Svalbard at a depth of c. 110 m. Note uneven sediment deposition and roughness of the channel bed. Photo credit for A: Caroline Clason; B-E: Jason Gulley; B published in Gulley et al. (2009b).

Figure 3. *(Portrait, column width)*
Meltwater systems in the palaeo-record: meltwater channels. (A) Distribution of meltwater channels on the bed of the British-Irish Ice Sheet (after Greenwood, 2008; Hughes, 2009). Approximate southern terrestrial limit of the ice sheet in Great Britain during the last glaciation (Devensian; in purple, dashed) is drawn after Clark et al. (2012). (B) Subglacial meltwater channels of the British-Irish Ice Sheet, in the Vale of Eden, NW England (after Hughes, 2009). Note the contrasting topologies compared to Fig. 4A: whereas eskers display a low-order dendritic pattern at a regional scale, subglacial meltwater channels form anastomosing networks. Note also the difference in scale between esker and palaeo-channel systems. Location of panel B is marked by black box in panel A.

Figure 4. *(Portrait, full page width)*
Meltwater systems in the palaeo-record: eskers. (A) Eskers in Keewatin, north-central Canada (adapted from Storrar et al., 2013). Note the correspondence with the retreating margin of the Laurentide Ice Sheet between 11 and 6.5 ka (drawn after Dyke et al., 2003, with isochrons in ^{14}C years). The limit of the Canadian Shield is marked by a yellow line. Rare examples of bifurcation or anastomosing are marked by arrows. Boxed areas are discussed in the text. (B) Ice sheet wide distribution of eskers on the bed of the North American Ice Sheet Complex. CIS – Cordilleran Ice Sheet, LIS – Laurentide Ice Sheet, Last Glacial Maximum limit after Kleman et al. 2010. Esker networks (marked in red, adapted from Storrar et al., 2013, are best developed on the Canadian Shield (in green, from Wheeler et al., 1996). Location of panel A and Fig. 10A is shown by black rectangles.

Figure 5. *(Portrait, full page width)*
Conceptual diagram of a palaeo-ice sheet hydrological system and its geological imprint. A) Meltwater channels incised into the bed, and esker deposits upon the bed, are both taken as geomorphic products of a channelised meltwater drainage system. Porewater flow through a till aquifer may be speculated upon, but there are few templates for interpreting distributed drainage. Through time, the snapshot shown here is repeatedly imprinted and overprinted by each successive change in ice sheet configuration or drainage behaviour, during both advance and retreat of an ice sheet. In a landform (plane) view (B), a composite landform product represents different drainage regimes and/or different ice sheet configurations over a timeframe of 10^1-10^4 years. In a stratigraphic (cross-section) view (C), these different regimes and/or ice sheet configurations must be deciphered from a stack of units. The timeframes represented by individual landforms or sedimentological traces, to a coherent unit, to a sequence of units, to a whole palaeo-ice sheet bed encompass the whole spectrum from seasons to tens of thousands of years.

Figure 6. *(Portrait, full page width)*
A variety of different drainage system topologies may be inferred from observations and theory. A) Dendritic channel system which, following classic R-theory, captures flow across a pressure gradient and is conceptually considered to drain water ‘efficiently’ in low-pressure channels. Observed esker systems strongly resemble this form, at a variety of scales (Figs. 4, 10). B) A non-dendritic channel network is commonly observed among subglacial meltwater channels in the palaeo record, at a variety of scales (Figs. 4, 10). Considered ‘inefficient’ systems, these are conceptually consistent with canal theory or linked cavity networks. C) Flow captured by a main low pressure channel, but locally feeders are non-dendritic (cf. pressure observations by Fudge et al., 2008, at glacier scale). D) Theoretical and modelling attempts to bridge channelised and distributed models (e.g. Hewitt, 2011) use a simplified topology. E) Contemporary observations point to a high pressure environment, non-channelised (Meierbachtol et al., 2013), everywhere other than immediate marginal zone. On all panels dark blue marks channel conduits at relatively low pressure (with line thickness scaling with discharge where this can be safely assumed), red conduits at relatively high pressure, light blue dashes indicate drainage areas, and a non-channelised high pressure system is represented by red dotted zones. Ice margin in black.

Figure 7. *(Portrait, column width)*
Idealised consequences of A) steady and B) variable meltwater input to a subglacial channel, following Röthlisberger (1972). In steady input scenario A), a conduit water level can remain below overburden pressure at distances 100s km inward from the ice margin. Conduit size (cross-sectional area) is smallest close to the ice margin. B) When input discharge varies (e.g. diurnally, seasonally), the conduit invariably is driven to overburden pressure; conduit pressure remains below overburden pressure for only small windows of time (e.g. at the very

start of a rise in discharge, and when discharge begins to fall before creep-closure re-adjusts to a lower input).

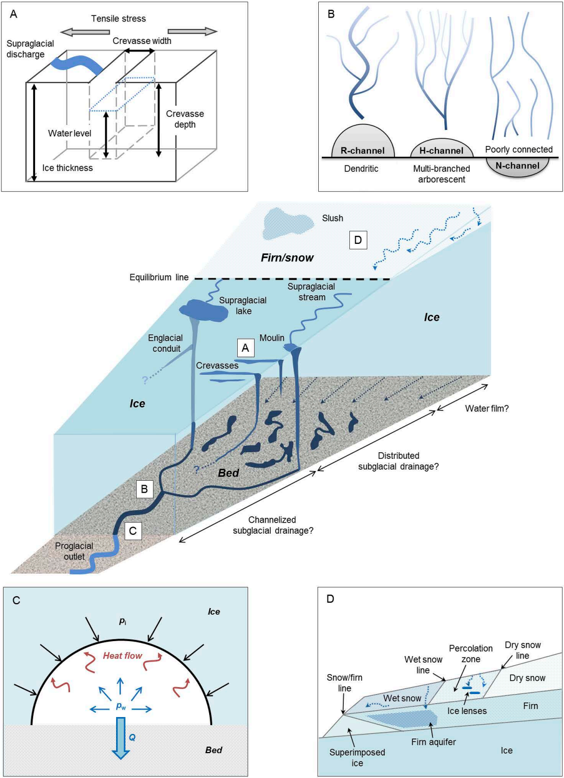
Figure 8. *(Portrait, column width)*
Peak discharge circles (scaled with circle area) for the Glacial Lake Missoula floods (Hanson et al., 2012), the Glacial Lake Agassiz outburst flood (Clarke et al., 2004), the Labyrinth channels (Lewis et al., 2006), average discharge of the Amazon River (Vörösmarty et al., 1989), Skeiðarárjökull jökulhlaup (Russell et al., 2006), a west Greenland supraglacial lake drainage (Das et al., 2008), the Bering Glacier outburst flood (Burke et al., 2010), 2009 peak discharge for the Leverett Glacier proglacial outlet (Bartholomew et al., 2011b), total peak discharge across the Siple Coast grounding line (Carter and Fricker, 2012) and the Adventure trench subglacial lake drainage (Wingham et al., 2006). For reference, the discharge of Glacial Lake Missoula is 340,000 times greater than the Adventure trench subglacial lake drainage; comparable with the difference in discharge between the Amazon River and a small brook.

Figure 9. *(Portrait, column width)*
Uncertainty in the correspondence between drainage regime and landform formation. Four different drainage regimes (grey lines, i-iv) may all have a similar average state (discharge over time), for example equivalent to drainage regime iv, when considered over a long timescale. It is uncertain where, with respect to these drainage regimes, most geomorphic work takes place. Landforms may represent the 'average' condition (A), may represent geomorphic work towards the peaks or troughs of periodic discharge regimes (P_{max} or P_{min}, respectively) or may represent geomorphic work which only takes place under few, isolated, high magnitude events (E).

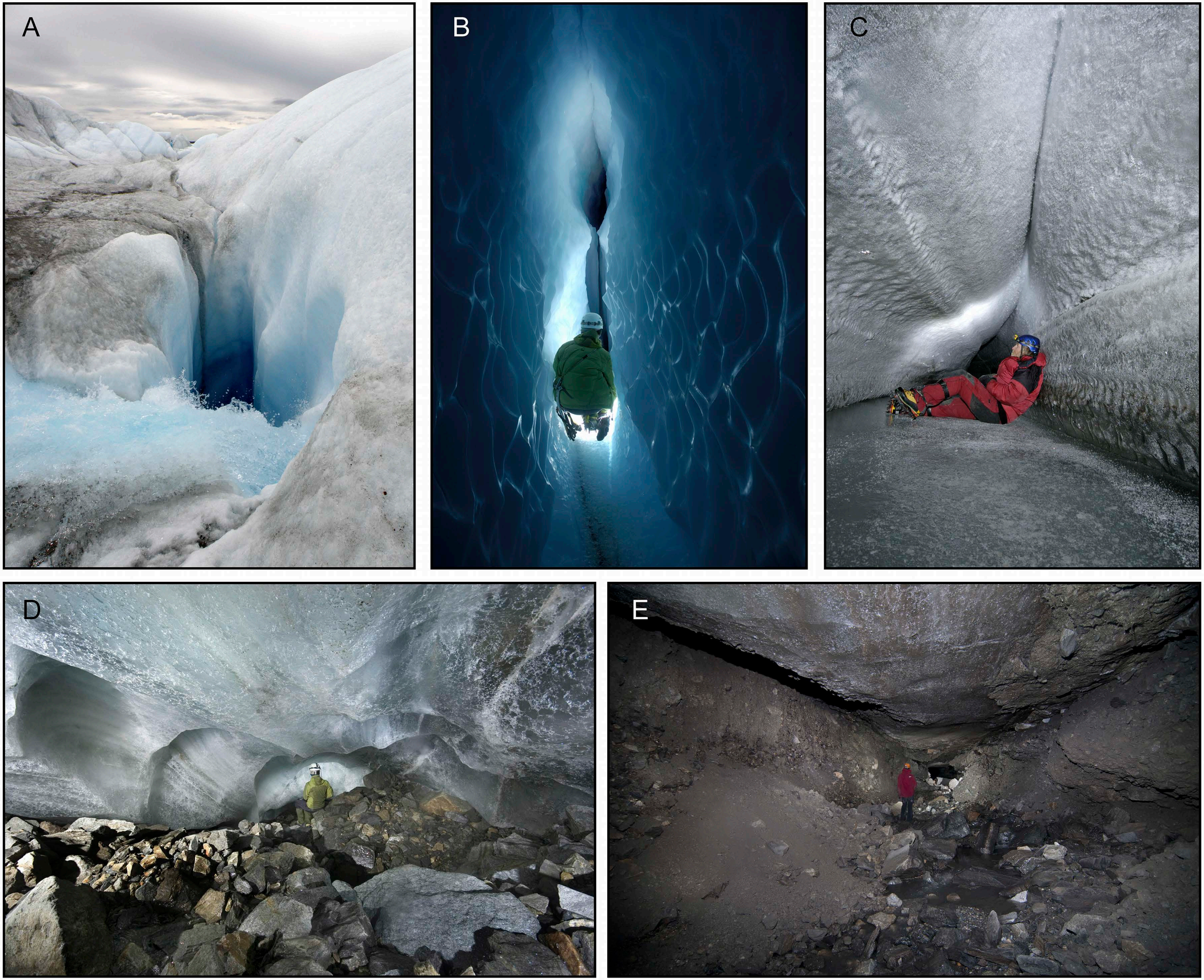
Figure 10. *(Portrait, column width)*
Despite different continental ice sheets, different climate and ice sheet zones, different substrate geologies, indeed, different marine isotope stage ice sheets, the scaling and organisation of contrasting landform systems in the palaeo-record is broadly similar. (A) Eskers in Ontario are marked in red (with filled gaps between individual esker segments, adapted from Storrar et al., 2013). Bathymetry of Lake Superior is shown in shades of blue (data from NOAA-NGDC) with networks of tunnel valleys visible on the lake floor. Note the correspondence with the retreating margin of the Laurentide Ice Sheet between 11 and 8.5 ka (drawn after Dyke et al., 2003, with isochrons in ^{14}C years). (B) Esker networks in Sweden (with filled gaps between individual esker segments) are marked in red (provided courtesy of Clas Hättestrand) and Elsterian tunnel valleys in Germany and Poland are marked in blue (adapted from Huuse and Lykke-Andersen, 2000). Both panels are drawn at the same scale. The black dashed line marks the boundary between the hard crystalline rocks of the Canadian (A) and the Baltic (B) shields (upper parts of the panels) and softer sedimentary rocks.

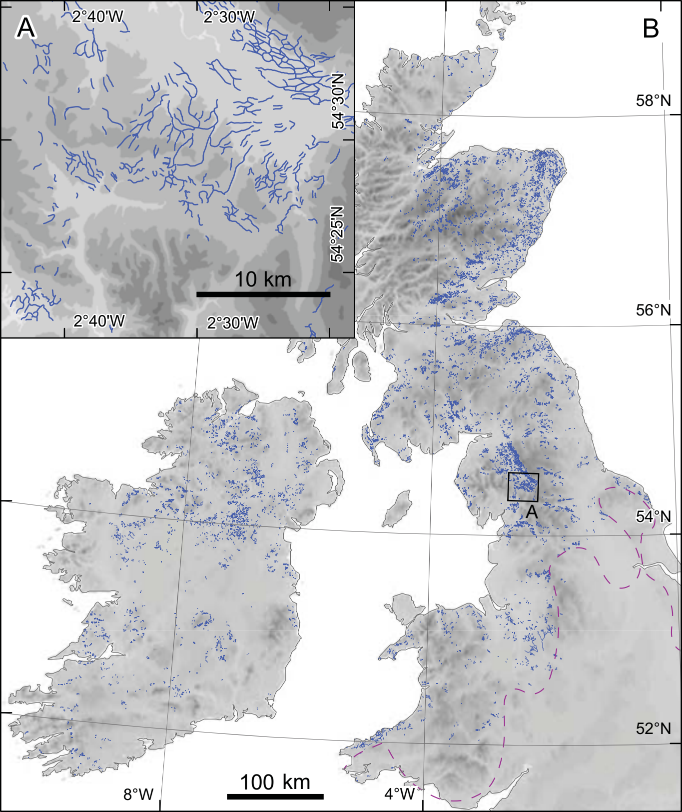
Figure 11. *(Portrait, column width)*

First order hydrological regimes of an ice sheet through space and time. X axis represents ice sheet span from centre to margin; Time (t) on vertical axis from maximum ice sheet extent to deglaciation. A high pressure environment (H, red dots) is found throughout the ice sheet. Röthlisberger-type channelised drainage (C) dissects the outer marginal zone, pockets of efficient delivery may be found inward of this, and short-term high magnitude drainage events (pale blue lines) are superimposed throughout the ice sheet history. The inward limit of the C zone is uncertain: it reflects changing discharge, ice thickness, drainage event frequency (amplitude of meltwater input variability). For further explanation refer to the main text.

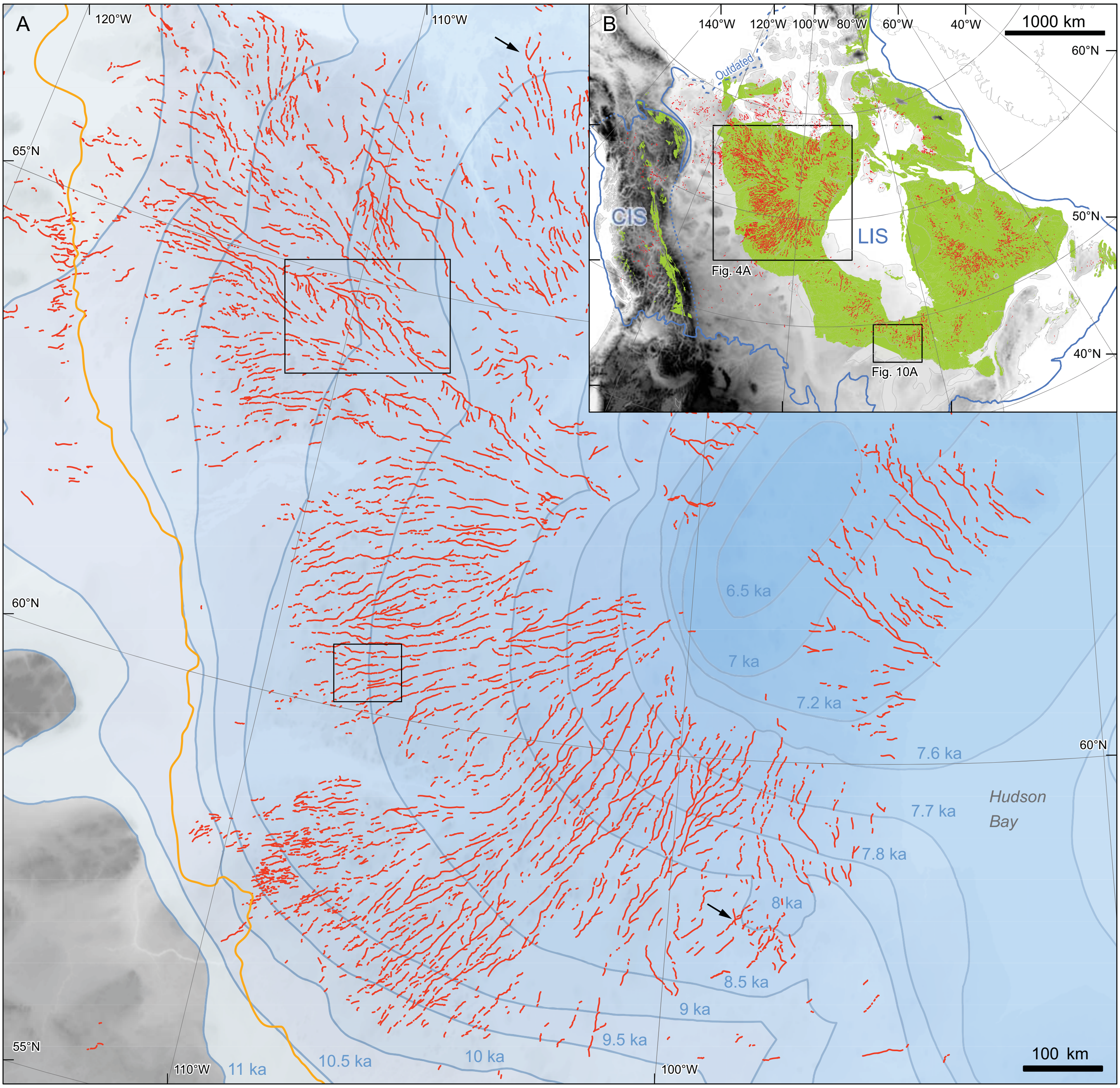


A
Tensile stress
Supraglacial discharge
Crevasse width
Crevasse depth
Water level
Ice thickness
B
R-channel
Dendritic
H-channel
Multi-branched arborescent
Poorly connected
N-channel
Slush
D
Firn/snow
Equilibrium line
Supraglacial lake
Supraglacial stream
Ice
Englacial conduit
Moulin
A
Crevasses
?
Ice
Water film?
Distributed subglacial drainage?
?
Bed
B
C
Channelized subglacial drainage?
Proglacial outlet
C
Ice
p_i
Heat flow
p_w
Q
Bed
D
Percolation zone
Dry snow line
Wet snow line
Dry snow
Snow/firn line
Wet snow
Firn
Ice lenses
Firn aquifer
Superimposed ice
Ice

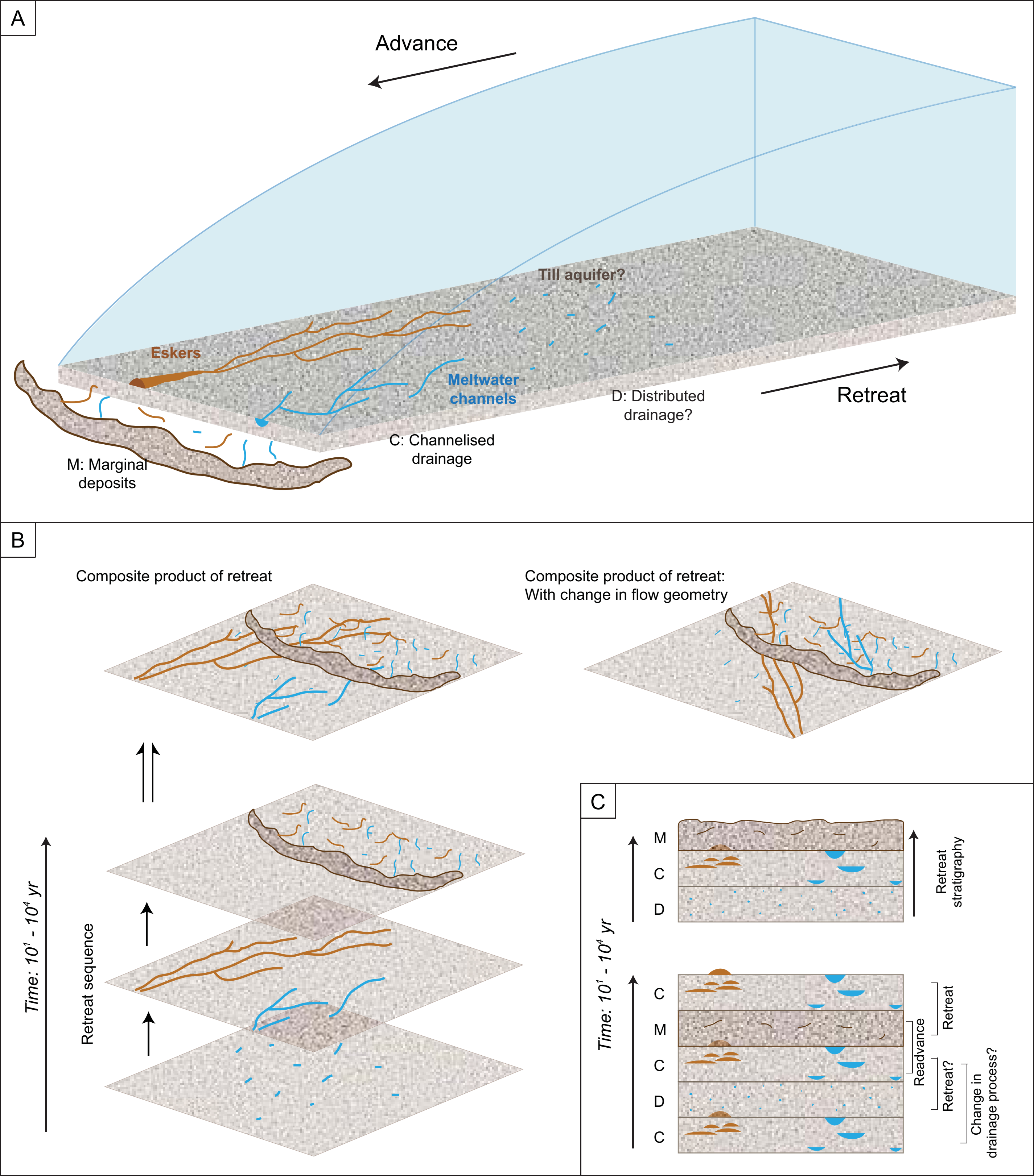

A
B
C
D
E

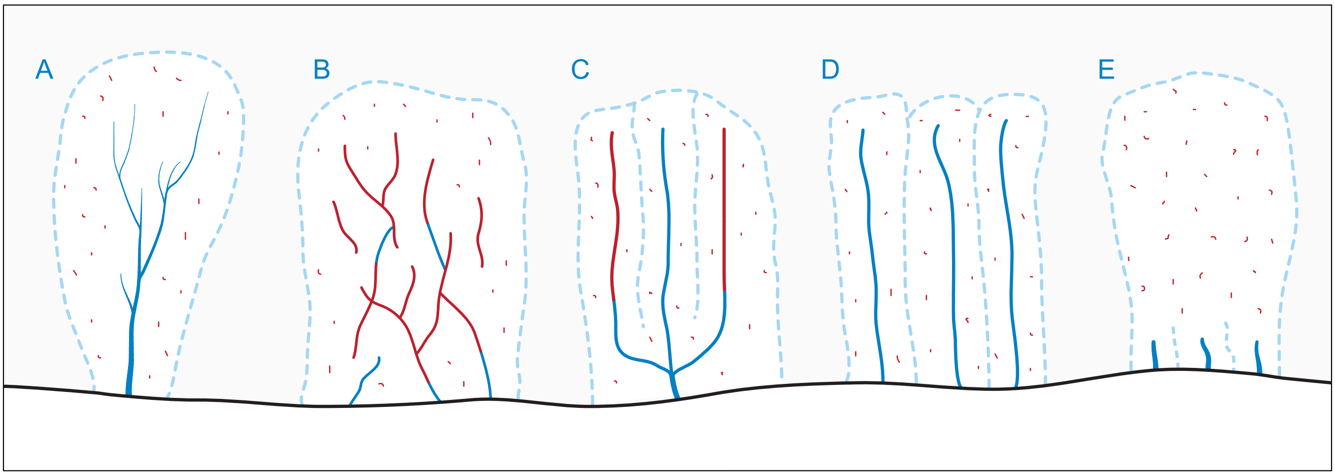


A
2°40'W
2°30'W
54°30'N
54°25'N
10 km
2°40'W
2°30'W
B
58°N
56°N
54°N
52°N
A
8°W
4°W
100 km

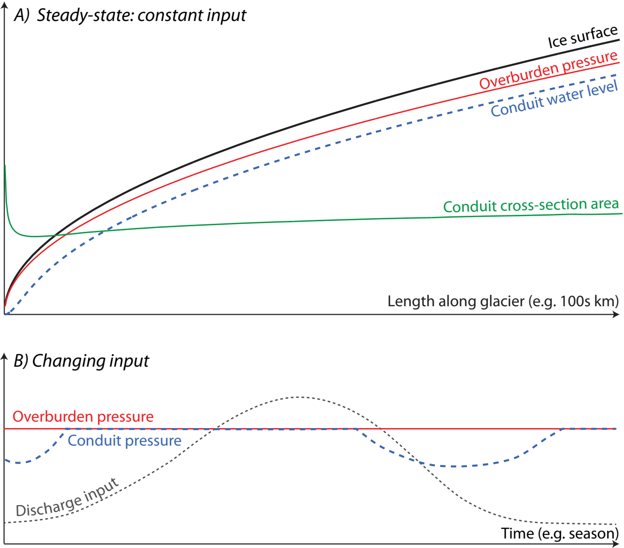


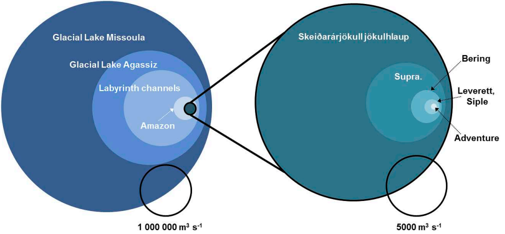
A
120°W
110°W
65°N
60°N
55°N
110°W
100°W
6.5 ka
7 ka
7.2 ka
7.6 ka
60°N
7.7 ka
Hudson
Bay
7.8 ka
8 ka
8.5 ka
9 ka
9.5 ka
10 ka
10.5 ka
11 ka
100 km
B
140°W
120°W
100°W
80°W
60°W
40°W
1000 km
60°N
50°N
40°N
Outdated
CIS
LIS
Fig. 4A
Fig. 10A


A
Advance
Till aquifer?
Eskers
Meltwater channels
Retreat
D: Distributed drainage?
C: Channelised drainage
M: Marginal deposits
B
Composite product of retreat
Composite product of retreat: With change in flow geometry
Time: 10^1 - 10^4 yr
Retreat sequence
C
M
C
D
Retreat stratigraphy
Time: 10^1 - 10^4 yr
C
M
C
D
C
Retreat
Readvance
Retreat?
Change in drainage process?

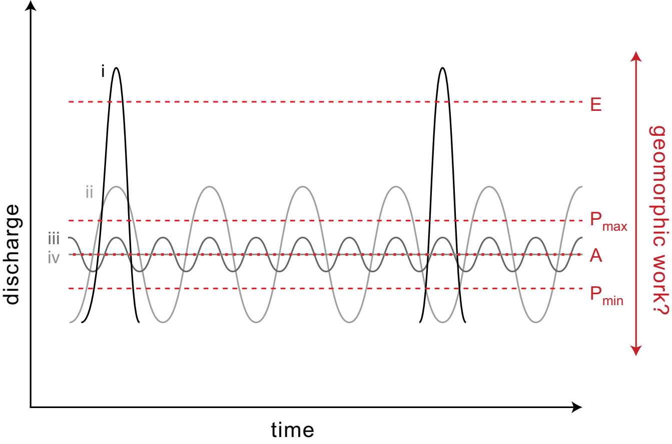


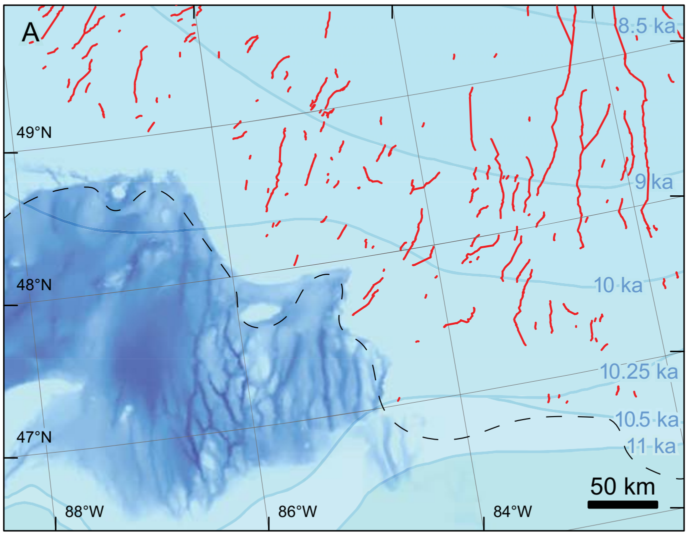
A
B
C
D
E

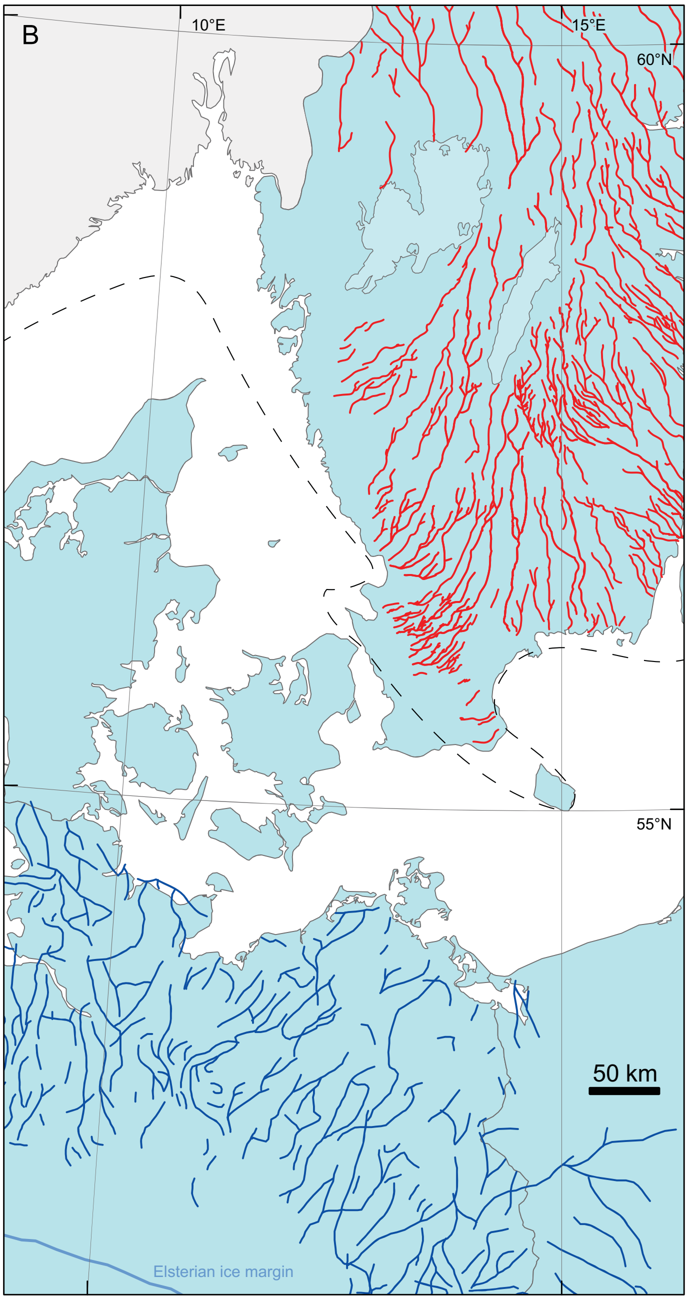

A) Steady-state: constant input
Ice surface
Overburden pressure
Conduit water level
Conduit cross-section area
Length along glacier (e.g. 100s km)
B) Changing input
Overburden pressure
Conduit pressure
Discharge input
Time (e.g. season)

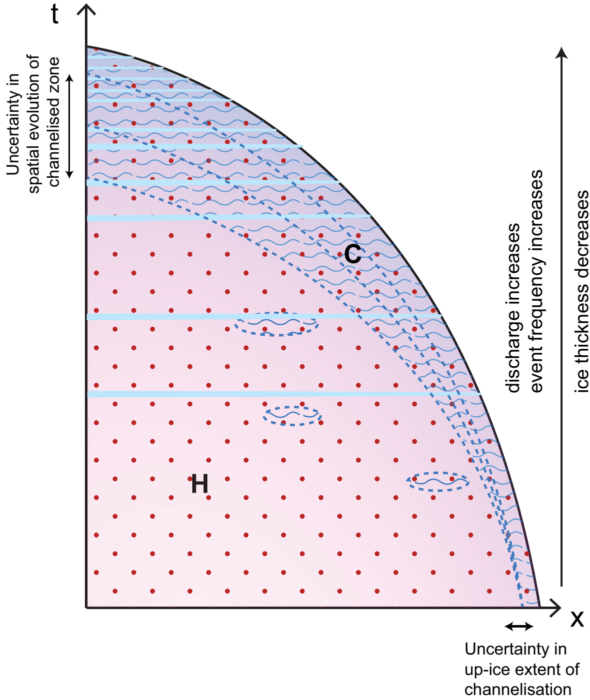


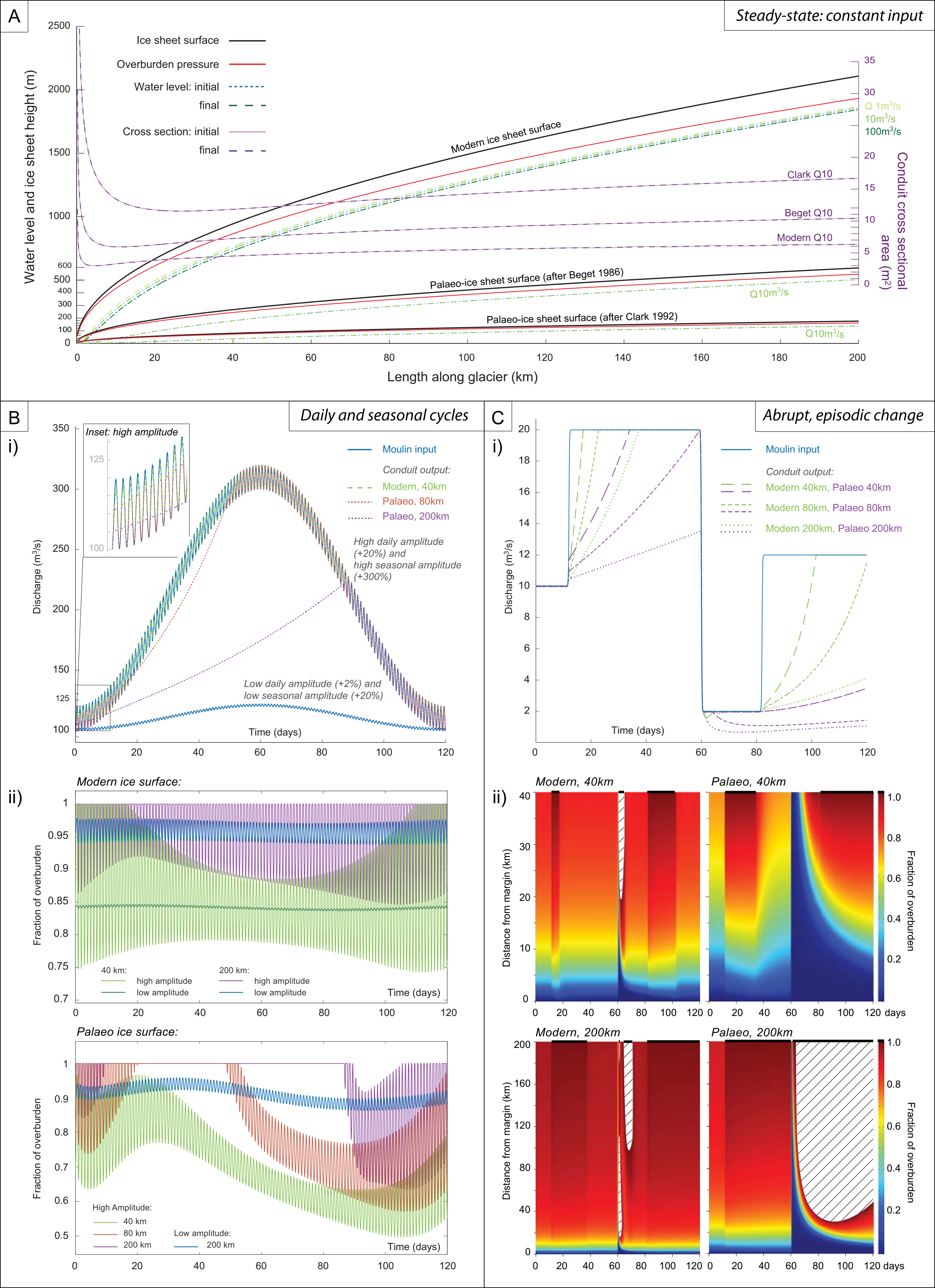
Event	Reference	Discharge ($m^3\ s^{-1}$)
Glacial Lake Missoula floods (Cordilleran Ice Sheet)	Hanson et al. (2012)	17 000 000
Glacial Lake Agassiz outburst flood (Laurentide Ice Sheet)	Clarke et al. (2004)	5 000 000
Labyrinth channels (East Antarctic Ice Sheet)	Lewis et al. (2006)	2 200 000
Amazon River	Vörösmarty et al. (1989)	207 000
Skeiðarárjökull jökulhlaup (Iceland)	Russell et al. (2006)	53 000
Supra.: Western Greenland supraglacial lake drainage (Greenland Ice Sheet)	Das et al. (2008)	8700
Bering glacier outburst flood (Alaska)	Burke et al. (2010)	1500
Leverett proglacial river peak discharge (Greenland Ice Sheet)	Bartholomew et al. (2011a)	317
Siple coast grounding line total peak discharge (West Antarctic Ice Sheet)	Carter and Fricker (2012)	300
Adventure subglacial trench lake drainage (East Antarctic Ice Sheet)	Wingham et al. (2006)	50


discharge
time
i
ii
iii
iv
E
P max
A
P min
geomorphic work?


A
8.5 ka
49°N
9 ka
48°N
10 ka
10.25 ka
10.5 ka
47°N
11 ka
50 km
88°W
86°W
84°W


B
10°E
15°E
60°N
55°N
50 km
Elsterian ice margin


t
Uncertainty in spatial evolution of channelised zone
C
H
discharge increases
event frequency increases
ice thickness decreases
x
Uncertainty in up-ice extent of channelisation



A
Steady-state: constant input
Ice sheet surface
Overburden pressure
Water level: initial
final
Cross section: initial
final
Modern ice sheet surface
Clark Q10
Beget Q10
Modern Q10
Palaeo-ice sheet surface (after Beget 1986)
Palaeo-ice sheet surface (after Clark 1992)
Q 1m^3/s
10m^3/s
100m^3/s
Q10m^3/s
Q10m^3/s
Water level and ice sheet height (m)
Conduit cross sectional area (m^2)
Length along glacier (km)
B
Daily and seasonal cycles
i)
Inset: high amplitude
Moulin input
Conduit output:
Modern, 40km
Palaeo, 80km
Palaeo, 200km
High daily amplitude (+20%) and high seasonal amplitude (+300%)
Low daily amplitude (+2%) and low seasonal amplitude (+20%)
Discharge (m^3/s)
Time (days)
ii)
Modern ice surface:
40 km:
high amplitude
low amplitude
200 km:
high amplitude
low amplitude
Fraction of overburden
Time (days)
Palaeo ice surface:
High Amplitude:
40 km
80 km
200 km
Low amplitude:
200 km
Fraction of overburden
Time (days)
C
Abrupt, episodic change
i)
Moulin input
Conduit output:
Modern 40km, Palaeo 40km
Modern 80km, Palaeo 80km
Modern 200km, Palaeo 200km
Discharge (m^3/s)
Time (days)
ii)
Modern, 40km
Palaeo, 40km
Modern, 200km
Palaeo, 200km
Distance from margin (km)
Fraction of overburden
days