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# Theoretical, contemporary observational and palaeo-perspectives on ice sheet hydrology: Processes and products

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

3  
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14  
15 **Abstract**

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

42

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

45

46

## 47 **1. Introduction**

48 Ice sheet hydrology has long been recognised as a crucial component in the understanding of  
49 ice sheets, their behaviour and their evolution. The presence and distribution of water at the  
50 ice sheet bed is widely considered to control ice motion by facilitating basal sliding and  
51 substrate deformation (e.g. Alley et al., 1986; Engelhardt and Kamb, 1997), triggering  
52 enhanced or episodic fast flow via water pressure build-up (e.g. Kamb, 1987; Zwally et al.,  
53 2002), and impeding ice flow due to basal freeze-on, water piracy and bed stiffening (e.g.  
54 Alley et al., 1994; Tulaczyk et al., 2000; Bougamont et al., 2003). Whilst the time domain for  
55 ice sheet-scale behaviour has traditionally been considered to be from 100s – 1000s years  
56 (Alley and Whillans, 1984; Huybrechts and Oerlemans, 1990; Clark, 1994; Kleman et al.,  
57 2006; Kleman and Applegate, 2014; cf. Bamber et al., 2007) meltwater processes are highly  
58 spatially and temporally variable. This difference in conceptual and observational scales of  
59 different glaciological properties presents a problem for our understanding of the physical  
60 processes of meltwater drainage, of ice dynamic responses, of the land-forming footprint of  
61 drainage, and for implementation of these processes within large scale models.

62

63 After decades of studies of meltwater processes at a glacier-scale (e.g. Mathews, 1964;  
64 Stenborg, 1970; Iken and Bindschadler, 1986; Willis et al., 1990; Bingham et al., 2003; Mair  
65 et al., 2003), the last decade has seen a shift towards an ice sheet scale focus, largely out of  
66 concern for the rate of change and the future of the modern-day ice sheets. Pronounced  
67 seasonality of meltwater drainage linked to high summer supraglacial meltwater production  
68 has been observed in the ablation zone of the Greenland Ice Sheet (Zwally et al., 2002;  
69 Bartholomew et al., 2011a). On top of seasonal changes in subglacial efficiency, high-  
70 magnitude supraglacial lake drainage events connect the ice surface to the bed through ice up  
71 to 1000 m thick (Das et al., 2008; Doyle et al., 2013). In Antarctica, repeated drainage of  
72 water between subglacial lakes has been inferred, as well as water transfer towards the ice  
73 margin and interior freeze-on areas (Fricker et al., 2007; Wolovick et al., 2013). Such  
74 growing understanding of ice sheet scale drainage behaviour brings the scale of contemporary  
75 ice sheet research in line with that more typically inhabited by palaeo glaciology. Elements of  
76 palaeo meltwater drainage networks have long been recognised on the beds of Pleistocene ice  
77 sheets (e.g. Mannerfelt, 1945; Prest et al., 1968). Meltwater landforms have been widely used  
78 for reconstructions of past ice geometries and dynamics (e.g. Borgström, 1989; Kleman et al.,  
79 1992; Boulton et al., 2001; Margold et al., 2013a,b), but only limited focus has been directed  
80 towards the processes and function of the palaeo hydrological system.

81

82 In parallel to observational research, developments within the ice sheet modelling community  
83 increasingly allow for credible predictions of ice sheet responses and feedbacks under future  
84 climate scenarios (e.g. Gregory and Huybrechts, 2006; Greve et al., 2011; Applegate et al.,  
85 2012; Barrand et al., 2013; Edwards et al., 2014), and enable Quaternary (and older) ice sheet  
86 behaviour and its coupling to longer term climate regimes to be explored and better  
87 understood (e.g. Huybrechts and Oerlemans, 1988; Payne et al., 1989; Marshall and Clarke,  
88 1997a, b; Tarasov and Peltier, 2004; Pollard and DeConto, 2009; Gregoire et al., 2012). Ice  
89 dynamics are arguably reasonably well captured in the latest generation of ice sheet models,  
90 but a realistic representation of ice sheet hydrology is a vital, missing component which must  
91 be incorporated for more robust simulations of likely ice sheet responses to increased  
92 meltwater production. For this we need a reliable, quantitative description of the processes  
93 and form of ice sheet hydrological drainage systems. We therefore consider it timely to  
94 review ice sheet hydrology as it stands and highlight the crucial issues for future  
95 consideration.

96  
97 Ice sheet hydrology has generally been approached from different perspectives within three  
98 research sub-disciplines: *theoretical glaciology* which describes glacial processes in physical  
99 terms, contemporary *observational glaciology* within which real-time data concerning modern  
100 ice sheets are gathered, and *palaeo glaciology* which reconstructs the dynamics of past ice  
101 sheets. Although each of these fields is informed by the progress of the others, these research  
102 sub-disciplines traditionally work within different domains of space and time, with research  
103 conducted by different communities and developing along different trajectories. A number of  
104 reviews of glacial hydrology provide comprehensive summaries from different research  
105 perspectives (Clark and Walder, 1994; Hubbard and Nienow, 1997; Fountain and Walder,  
106 1998; Benn and Evans, 2010; Bingham et al., 2010; Walder, 2010; Irvine-Fynn et al., 2011;  
107 Cook and Swift, 2012; Chu, 2014; Flowers, 2015). However, there have historically been few  
108 attempts to approach glacial hydrology from a fully cross-disciplinary perspective, leading to  
109 a disconnect of theoretical, contemporary observational and palaeo glaciology in their current  
110 understanding of ice sheet hydrology. With research agendas converging on the ice sheet  
111 scale, there is now the prospect for more integrated research designs that will tackle questions  
112 that have thus far been addressed semi-independently. Our effort here is to identify these  
113 disconnects between the findings and understandings of the sub-disciplines and to assess to  
114 what extent is it possible to merge the advances in each field into a more holistic view.

115  
116 We first review the contributions of the three research sub-disciplines, focussing on the form,  
117 distribution and spatial scale of the hydrological system, and the timescales, rates and fluxes  
118 of meltwater delivery. We do not hope to be exhaustive in these reviews; other dedicated  
119 reviews better meet this demand. Our reviews rather serve as a basis for discussion of the  
120 areas of consensus in understanding and the disconnects, and significance thereof, between  
121 research fields, in order to address questions most urgently demanding attention. The  
122 discussion focusses primarily on subglacial drainage and inputs to subglacial drainage from

123 the supraglacial domain. Here we identify a number of questions for which significant  
 124 uncertainty remains, concerning the prevailing form, controls, variability, and spatial extent of  
 125 subglacial drainage and the significance of high magnitude drainage events compared to long-  
 126 term steady meltwater drainage. With this review, we aim to provide a critical basis for future  
 127 collaborative efforts necessary to further our understanding of ice sheet hydrology and thus  
 128 ice sheet behaviour as a whole.

129

## 130 **2. Ice sheet hydrology: a theoretical perspective**

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

138

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

$$142 \quad \Phi = \Phi_0 + p_w + \rho_w g z \quad (1)$$

143 where  $\Phi_0$  is a (constant) background potential,  $p_w$  is the water pressure,  $\rho_w$  is the density of  
 144 water,  $g$  is acceleration due to gravity and  $z$  is the elevation. For a connected hydraulic system  
 145 water will flow down the negative gradient of  $\Phi$ . In the subglacial case, the gradient of the  
 146 hydraulic potential,  $\nabla\Phi$ , will be due to the ice sheet's geometry (glacier bed,  $z_b$ , and surface,  
 147  $z_s$ ) and due to  $p_e$  as

$$148 \quad \Psi := -\nabla\Phi = -\nabla(p_i - p_e + \rho_w g z_b) \approx -\rho_i g \nabla z_s - (\rho_w - \rho_i) g \nabla z_b + \nabla p_e \quad (2)$$

150

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

158

### 159 *2.1. Englacial water routing*

160 Unless water is stored supraglacially (in lakes or a snow-firn aquifer) or routed away as  
 161 surface runoff, it is fed to the englacial system (Fig. 1). A small amount of water can be  
 162 transported between the boundaries of individual ice grains (Nye and Frank, 1973; Lliboutry,  
 163 1971), but this form of transport is incapable of flow of any significance (Lliboutry, 1971).

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

177  
 178 A Shreve topology assumes a coherent ice mass acting as a hydraulically connected system,  
 179 although this is under steady-state conditions. Due to this limitation, much recent research has  
 180 turned to focus on hydro-fracturing through ice sheets, to better address observations that  
 181 water primarily exploits weaknesses and structures within the ice mass (Fig. 1). Hydrofracture  
 182 theory has also been invoked to explain short-term events such as the drainage of supraglacial  
 183 lakes on Greenland (van der Veen, 2007; Krawczynski et al., 2009). Initially Weertman  
 184 (1973) showed, by treating ice as an elastic material with negligible fracture toughness, that  
 185 the difference in  $\rho_w$  and  $\rho_i$  could cause crevasse propagation through an ice sheet. Later  
 186 studies have used a more sophisticated model in the form of linear elastic fracture mechanics  
 187 (LEFM), which has the possibility to include a parameter for the fracture toughness of ice  
 188 (Smith, 1976; van der Veen, 1998, 2007). The rate at which a water-filled crevasse propagates  
 189 can be approximated by

$$190 \quad d = \left(\frac{\rho_w}{\rho_i}\right)^{\frac{2}{3}} Q_r t \quad (3)$$

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

200  
 201 Fountain and Walder (1998) focussed on the energy dissipation of flowing water at the  
 202 bottom of a crevasse, rather than the initiation or propagation of the crevasse itself. Assuming

203 that the energy is used to melt the surrounding ice, they find the relationship

204

$$205 \quad \dot{d} = c_1 S^{\frac{19}{16}} Q^{\frac{5}{8}} \quad (4)$$

206

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

215

## 216 2.2. Subglacial water routing

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

233

### 234 2.2.1. Fast systems

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

$$238 \quad \frac{dp_w}{dx} = c_2 Q^{\frac{-2}{11}} \left( \frac{p_e}{nA} \right)^{\frac{8n}{11}} \quad (5)$$

239 where  $Q$  is as above,  $n$  and  $A$  are parameters from the flow law (Glen, 1952),  $c_2$  is a constant  
240 dependent on ice rheology, channel roughness/shape and the pressure melting effect, and  $x$  is  
241 the horizontal distance along channel (after Röthlisberger, 1972; Shreve, 1972; Walder,  
242 2010). From Eq. (5) one can see that the change in water pressure decreases with discharge.

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

245

246 A modification to the theory was made by Lliboutry (1983), who claimed that glacier sliding  
247 and bed separation (cavities) were a necessity for channels to exist, and that full channels  
248 (higher  $p_w$  than atmospheric) occur only at short periods of high discharge, in which the  
249 channel grows in size. Lliboutry suggested that the normal state would be water flowing at  
250 atmospheric pressure in a channel that is closing due to the plastic deformation of the  
251 overlying ice, a conclusion that was also reached by Hooke (1984). Furthermore Hooke et al.  
252 (1990) argued that the restriction to a semi-circular shape was unrealistic due to the fact that  
253 stress regimes would alter at the glacier bed interface where friction occurs. Moreover there  
254 would be preferential melt of the side of the channel walls under conditions of open flow.  
255 They showed that if the channel geometry was modified to be low and wide, the closure rates  
256 of a channel could be explained without having to prescribe the ice to be unusually soft (as in  
257 Röthlisberger, 1972). These modifications to the basic Röthlisberger theory predict a multi-  
258 arborescent, widespread system of low, wide channels cut into the basal ice.

259

### 260 2.2.2. *Slow systems*

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

267

$$268 \quad Q = u_b^m \left( \frac{d\phi}{ds} \right)^{\frac{1}{2}} p_e^{-n} \quad (6)$$

269

270 where  $m$  is a constant and  $s$  is the along-flow path coordinate (after Fountain and Walder,  
271 1998). For a given discharge, this system would result in a much lower  $p_e$  than in the fast  
272 system of Eq. (5). Moreover, according to Eq. (6), cavities with higher discharge would have  
273 higher water pressures than cavities with lower discharge. This does not therefore lead to a  
274 capturing mechanism as in the above described channelised systems. A dendritic system does  
275 not develop; rather, the topology of the system is governed by the irregular pre-existing  
276 distribution of bedrock obstacles. Comparing Eqs. (5) and (6) qualitatively, one can conclude  
277 that Röthlisberger channels (R-channels) are more dependent on the energy dissipation of  
278 flowing water while linked cavity systems are dependent on the sliding speed. Both Walder  
279 (1986) and Kamb (1987) noted that increasing the discharge (water pressure) above a critical  
280 level would lead to a situation where cavities unstably increase in size. It was suggested that  
281 this instability could lead to a situation favourable to R-channel initiation. However, the  
282 orientation of the cross section of the orifices/cavities in Kamb's 2-D model (across ice



283 flow/surface gradient) would be perpendicular to the expected orientation of an R-channel  
284 (along ice flow/surface gradient).

285  
286 To explain glacier sliding, Weertman (1962, 1972) had early considered the concept of a  
287 continuous subglacial water film (mm in thickness) at the ice-bed interface. In a series of  
288 papers Lliboutry (see 1979) argued that the resulting stresses needed in Weertman's  
289 formulation to reproduce normal ice velocities would be too large to be supported by either  
290 ice or bedrock. Lliboutry suggested that bed separation on the lee side of bed protrusions must  
291 be a significant process. Later the stability of Weertman's water film was questioned by  
292 Walder (1982). Creyts and Schoof (2009) have recently expanded the concepts originally  
293 presented in Weertman (1972) and Weertman and Birchfield (1983). They showed that, on an  
294 uneven bed with protrusions to partially support the overlying ice, several stable states of  
295 water sheet thickness could exist. This idea is qualitatively similar to the slow system  
296 described by Walder (1986) and Kamb (1987), in that the water sheet would be expected to  
297 increase in height with decreasing effective pressure, but different in that the system can adapt  
298 to changes in hydraulic potential and effective pressure from one steady-state to another.  
299 Therefore, with changes of water sheet thickness, different levels of efficiency can result from  
300 the same basic system.

301  
302 *2.2.3. 'Active beds' – slow systems?*

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

312  
313 Sediment substrates are known to underlie large parts of contemporary and palaeo-ice sheets,  
314 yet how water flows through or interacts with a subglacial sediment layer remains to a large  
315 degree uncertain. It is unlikely that the underlying substrate could be permeable enough to  
316 evacuate all water through porewater, aquifer-type processes (Alley, 1989), even in an ice  
317 sheet setting where all melt is generated at the bed and water volumes are small. Shoemaker  
318 (1986) found that in order for the till to be an efficient aquifer, it must be integrated with a  
319 channel system in which channel spacing must be very close (on the order of  $10^2 - 10^3$  m).  
320 Walder and Fowler (1994) described till as a creep flow material, and showed that, as long as  
321 the creep of till is faster than that of ice, subglacial channels cut into the sediment (canals)  
322 could exist. Due to sediment erosion and transport, the shape of canals would have a low  
323 aspect ratio. For a low ice surface slope typical for ice sheets or ice streams ( $\sin \alpha = 0.001$ ,  
324 where  $\alpha$  is the ice surface slope), sediment canals together with flow through the till would be

325 the stable hydrological configuration. Contrary to R-channels, a system of canals would have  
326  $p_e$  close to zero and would not be arborescent. For fine-grained sediment, canals would have  
327 an approximate depth of 10 cm at a discharge of  $1\text{ m}^3\text{ s}^{-1}$  (Walder and Fowler, 1994). Ng  
328 (2000) expands on the canal concept by including mechanisms for sediment transport and  
329 downstream variability of the canal. The type of inverse relationship between effective  
330 pressure and discharge as presented by Walder and Fowler (1994) is concluded to hold, but  
331 that the discharge and sediment flux are both needed to determine the downstream  
332 distribution of the effective pressure.

333

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

342

#### 343 *2.2.4. Interactions between subglacial hydraulic systems*

344 It is unlikely that any one hydraulic system would be exclusive to an ice sheet, either spatially  
345 or temporally. Which of multiple coexistent components comes to dominate may depend on  
346 water flux, the ‘permeability’ of different domains, and the system geometry. Walder and  
347 Fowler (1994) argued that a till aquifer can coexist with sediment canals and R-channels, the  
348 dominant component depending on the ice surface slope (low slopes favouring canals).  
349 Hewitt (2011) examined how water may be drawn from a porous medium into channelised  
350 flow, and the spatial scales which may be appropriate in such an integrated system. He  
351 provides an analytical criterion for the position of channel initiation (i.e. channel length:  
352 distance up-ice from the margin) and a spacing relation coupled to this extent criterion. For an  
353 idealised ice sheet (characteristic length 1000 km; thickness 1 km) channel length would be  
354 10-100 km and spacing 7-15 km. Qualitatively, longer channels should develop where there  
355 are steep slopes, a high discharge, and a low permeability in the distributed system. These  
356 same conditions reduce the spacing between channels of a given length. Hewitt (2011)  
357 concludes that channel length and spacing are governed by the (in)efficiency of the distributed  
358 system.

359

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

362

$$363 \quad \frac{dR}{dt} = c_4 Q \Psi + u_b h - c_5 p_e^n R \quad (7)$$

364

365 where  $R$  is the cross-sectional area of the cavity or channel,  $h$  is a characteristic height of bed

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

375

### 376 *2.3. Theoretical ice sheet hydrology: summary*

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

392

## 393 **3. Ice sheet hydrology: contemporary observations**

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

403

### 404 *3.1. Supraglacial and englacial reservoirs and pathways*

405 The ablation zone of ice masses is often characterised by networks of streams running across  
406 the ice surface (Ferguson, 1973). These streams provide a mechanism for efficient transport  
407 of meltwater to the margin, or into supraglacial lakes and moulins, from which water can be

408 delivered into the englacial and subglacial systems. In the snowpack and firn, refreezing and  
409 percolation of meltwater can produce superimposed ice (Fig. 1D), lowering primary  
410 permeability and encouraging meltwater storage within supraglacial ponds and slush zones in  
411 topographic lows (Müller, 1962; Boon and Sharp, 2003; Wright et al., 2007; Irvine-Fynn et  
412 al., 2011). Vertical percolation rates are highly heterogeneous (Gerdel, 1954; Schneider and  
413 Jansson, 2004; Campbell et al., 2006) and meltwater retention in supraglacial and firn zones  
414 buffers ice sheet and sea level response to surface melting (Fig. 1D, Harper et al., 2012;  
415 Humphrey et al., 2012; Forster et al., 2014).

416

### 417 *3.1.1. Supraglacial lakes and their drainage*

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

434

435 Box and Ski (2007) estimate that a moulin with a cross-sectional area of  $10 \text{ m}^2$ , penetrating  
436 through 800 m thick ice, would hold only 0.1% of an 'average' lake's drained volume at any  
437 one time, based on lake dimension statistics for a sample of lake outburst events in west  
438 Greenland. Elevated basal water pressure could therefore be maintained, at least locally, for  
439 hours to days (e.g. Box and Ski, 2007; Doyle et al., 2013). Selmes et al. (2011) report that  
440 13% of Greenland's lakes undergo rapid drainage each year (averaged over a 5-year period).  
441 Of these fast drainage events, 61% occurred in the southwest sector of the ice sheet, in  
442 contrast to only 1% in the southeast. This highlights pronounced regional differences in  
443 supraglacial meltwater drainage processes, in addition to large regional differences in lake  
444 numbers and total lake area (Selmes et al., 2011). The net effect of short-term drainage events  
445 on longer-term ice surface motion and subglacial network efficiency remains unclear. Sundal  
446 et al. (2011) suggest that seasonal, surface-melt induced speed-ups may be relatively  
447 insignificant at an annual timescale, although their data were restricted to marginal regions.  
448 Furthermore, recent modelling indicates that single-event lake drainages, although large,  
449 transfer a much smaller proportion of annual melt to the bed than the more steady flow of

450 meltwater streams through moulins (Clason et al., 2015). Under future climatic warming  
451 scenarios the spatial coverage of supraglacial lakes may extend further into the interior of the  
452 Greenland Ice Sheet (Leeson et al., 2015), and may have potential to deliver meltwater to the  
453 ice sheet bed in areas of widespread inefficient subglacial drainage. However, the extent to  
454 which lakes forming at higher elevation could drain through new surface-to-bed connections  
455 is uncertain. Drainage may be limited by the relative dearth of existing crevasses through  
456 which water can reach the bed, producing tensile stresses in otherwise compressive areas  
457 (Stevens et al., 2015).

458

### 459 *3.1.2. Moulins and englacial conduits*

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

482

483 There has been a recent shift towards a greater focus on englacial hydrology, yet the exact  
484 form of surface-to-bed connectivity remains largely unobserved and unknown. This is  
485 especially true for high magnitude drainage events; a stark contrast to traditional thinking on  
486 long-term steady basal melting. On a macro-scale the englacial system consists of moulins,  
487 crevasses, fractures and enclosed horizontal conduits (Fig. 2). Rather than the dendritic  
488 topology of veins and conduits first hypothesised by Shreve (1972), direct imaging,  
489 exploration and speleology suggest a complex network of fractures and conduits (e.g. Vatne,  
490 2001; Fountain et al., 2005a, b; Benn et al., 2009; Gulley et al., 2009a). Benn et al. (2009)  
491 directly explored englacial drainage systems in Alaska, Nepal and Svalbard, finding that

492 hydrofracturing could lead to englacial system development across a large range of  
493 glaciological regimes (Fig. 2). Fracture formation and geometry was found to be controlled by  
494 supraglacial meltwater ponding and exploitation of resealed fractures and ‘crevasse traces’  
495 (cf. Holmlund, 1988), while repeated formation of conduits and fractures in the same places  
496 was postulated to yield stable locations for subglacial recharge (Benn et al., 2009; Catania and  
497 Neumann, 2010). Fountain et al. (2005a, b) argue that fracture systems dominate englacial  
498 water flow, and Gulley et al. (2009a) found that even in crevasse-free areas of polythermal  
499 glaciers, englacial channels could evolve through downward incision of meltwater channels,  
500 as previously described by Vatne (2001), coining the term ‘cut-and-closure’ channels (Fig.  
501 2C). Gulley et al. (2009a) observed down-cutting rates of up to  $0.33 \text{ m day}^{-1}$ , allowing full  
502 incision to the bed, although channel growth occurred more rapidly via knickpoint migration  
503 upstream. Compiled data on observed englacial conduits indicate lengths ranging from 10s to  
504 1000s metres at depths ~2-100 m below the ice surface (Gulley et al., 2009b). Little evidence  
505 has been found to support convergence of water-filled capillaries into larger conduits or for  
506 flow pathways strongly related to hydraulic gradient (cf. Shreve, 1972). Rather, slow- and  
507 fast-flow englacial components may co-exist. Large, spatially distributed networks of small  
508 fractures (tens of millimetres in width) found throughout the ice column likely operate  
509 simultaneously alongside fast, efficient flow through englacial channels and moulins (cf.  
510 Jansson, 1996; Fountain et al., 2005a, b).

511

### 512 *3.2. Subglacial configuration*

513 Much of our understanding of the configuration of subglacial drainage systems is inferred  
514 from monitoring of other, more accessible parts of the hydrological system or from ice  
515 dynamic proxies. Monitoring of proglacial discharge for its physical and chemical  
516 characteristics and for artificial tracer investigations (e.g. Willis et al., 1990; Nienow et al.,  
517 1996, 1998; Bingham et al., 2005; Cowton et al., 2013), study of boreholes drilled from the  
518 surface (e.g. Fountain, 1994; Hubbard et al., 1995; Mair et al., 2003; Fudge et al., 2008) and  
519 use of surface velocity patterns as a proxy for meltwater drainage events and subglacial  
520 drainage efficiency (e.g. Iken and Bindschadler, 1986; Kamb, 1987; Jansson, 1996; Mair et  
521 al., 2002; Bartholomew et al., 2012) are commonly used to deduce the likely form of the  
522 subglacial drainage network. Water pressure fields, meltwater delivery rates and outflow  
523 properties are related to meltwater routing through the subglacial domain, often then  
524 conceptualised as efficient or inefficient, or a channelised or distributed system (Fig. 1).

525

#### 526 *3.2.1. Insight from small ice masses*

527 Investigation of proglacial discharge and surface-to-bed boreholes has for many years been  
528 implemented on alpine and arctic glaciers, and many major advances in subglacial hydrology  
529 come from this environment (see reviews by Hubbard and Nienow, 1997, and Fountain and  
530 Walder, 1998). Tracer studies widely show seasonal evolution of the subglacial drainage  
531 system (e.g. Seaberg et al., 1988; Willis et al., 1990; Hock and Hooke, 1993; Nienow et al.,  
532 1996, 1998), with dye return curves measured in proglacial outlets typically evolving from a  
533 slow, diffuse and multi-peaked return to a single-peaked fast return as the melt season

534 progresses. Corresponding transition from a distributed to a channelised drainage  
535 configuration is widely assumed, with subsequent enlargement of channels, headward  
536 extension and ‘rationalisation’ of the network (Fig. 1) associated with seasonal up-glacier  
537 extension of surface meltwater production and retreat of the snowline (e.g. Nienow et al.,  
538 1998; Bingham et al., 2005). Borehole measurements of water pressure support a general  
539 model of seasonally enhanced drainage efficiency, with summer low pressure zones and  
540 synchronous fluctuations across boreholes (e.g. diurnal) taken to indicate a connected,  
541 channelised system (e.g. Fountain 1994; Hubbard et al., 1995; Harper et al., 2002; Fudge et  
542 al., 2008).

543  
544 Superimposed on this seasonal model is considerable spatial and temporal heterogeneity.  
545 Punctuated events such as supraglacial lake drainage events (Chu et al., 2009) and outburst  
546 floods or jökulhlaups (Russell, 1989; Russell et al., 1990; Mernild et al., 2008; Werder et al.,  
547 2009; Russell et al., 2011) are detected in proglacial discharge. Large borehole pressure  
548 variations, high gradients and sites disconnected from the main hydraulic system have been  
549 reported across relatively short distances (e.g. Murray and Clarke, 1995; Fudge et al., 2008).  
550 Hubbard et al. (1995) report diurnal interaction between a high pressure (‘distributed’) system  
551 and a low pressure channel, with increasing melt and rising pressure forcing water from the  
552 channel into the surrounding distributed system. While simultaneous variance in water  
553 pressure has been documented in closely clustered boreholes, observations by Fudge et al  
554 (2008) suggest that in late summer across-glacier connectivity is limited, and infer that long  
555 down-glacier water flow is necessary to drive large diurnal pressure fluctuations recorded  
556 within a relatively small area of the bed: the system is channelised, but not classically  
557 arborescent. Interpretations of drainage topology, connectivity and efficiency often make  
558 assumptions that borehole studies have difficulty substantiating; local borehole pressures may  
559 not reflect the pressure conditions of the wider system (Fudge et al., 2008; Meierbachtol et al.,  
560 2013), and limited spatial coverage of boreholes restricts the possibility of connecting with  
561 channelised drainage (Andrews et al., 2014).

562  
563 An observed correspondence between borehole water level fluctuations and surface ice flow  
564 velocities led Iken and Bindshadler (1986) to conclude that sufficient basal water pressures  
565 could locally hydraulically jack a glacier from its bed. Kamb (1987) drew similar conclusions  
566 from monitoring at Variegated Glacier during its 1982-83 surge event, further suggesting that  
567 a switch in subglacial regime from a low pressure, channelised system to a high-pressure  
568 linked-cavity (distributed) system may act as a trigger mechanism for glacier surges and fast  
569 ice flow. Despite uncertainties in linking basal water pressure to drainage topology, that  
570 changes therein can directly precipitate ice dynamic responses has been demonstrated in  
571 numerous settings (e.g. Jansson, 1995; Mair et al., 2003; Bingham et al., 2003; Sugiyama and  
572 Gudmundsson, 2004). On individual Arctic and alpine glaciers, small scale spatial patterns in  
573 surface ice velocity change in accordance with local basal water pressure changes are now  
574 well-established, both in terms of high pressure ‘spring events’ (e.g. Jansson and Hooke,  
575 1989; Mair et al., 2003) and subsequent dampening of velocity responses as the melt season

576 progresses and the drainage network becomes increasingly efficient.

577

### 578 3.2.2 *Scaling up: ice sheets*

579 The last decade has seen a shift of focus to the ice sheet scale when Zwally et al. (2002)  
580 argued that seasonal velocity patterns of outlet glaciers of the Greenland Ice Sheet are linked  
581 to seasonal drainage changes in a similar way to alpine contexts. The mode of surface  
582 meltwater delivery to the bed via moulins and crevasses modulates basal sliding (Colgan et  
583 al., 2011), and may further influence ice dynamics via cryo-hydrologic warming (Phillips et  
584 al., 2010; 2013). Seasonal meltwater-forced velocity behaviour of some of Greenland's outlet  
585 glaciers is now well-documented (e.g. van de Wal et al., 2008; Shepherd et al., 2009;  
586 Bartholomew et al., 2010, 2011a, b; Sundal et al., 2011) and similar behaviour is exhibited by  
587 both terrestrial and marine-terminating outlets (Sole et al., 2011). However, uncertainty  
588 remains on the extent to which inland expansion of supraglacial meltwater penetration to the  
589 bed can offset a reduction in mean annual ice velocities caused by self-regulation of the  
590 subglacial drainage system and decreased winter velocities (Sundal et al., 2011; Sole et al.,  
591 2013; Doyle et al., 2014; van de Wal et al., 2015).

592

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

604

605 However, these discrete proxy measurements provide relatively little insight to the actual  
606 configuration of subglacial drainage system. This is largely inferred, based on an assumption  
607 that 'efficient' drainage can be correlated with a channelised topology and delayed or diffuse  
608 delivery reflects a high pressure distributed system. This assumption is increasingly  
609 challenged. Harper et al. (2010) argue that basal crevasses act as subglacial meltwater stores  
610 and yield an irregular distribution of basal water pressure. Furthermore, measurement of  
611 borehole pressures in southwest Greenland, twinned with numerical modelling of channel  
612 maintenance, do not support the presence of stable systems of relatively low pressure more  
613 than 17 km inland of the margin (Meierbachtol et al., 2013), in contrast to Chandler et al.  
614 (2013); perhaps pointing to a different type of subglacial network. High ice-velocity events  
615 were observed by Cowton et al. (2013) up to 14 km from the Greenland Ice Sheet margin,  
616 even after the inferred channelisation of the subglacial system, perhaps due to meltwater  
617 inputs exceeding the drainage system capacity. Based on direct channel exploration in



618 conjunction with dye tracing analyses, Gulley et al. (2012a, b) suggest that elevated pressure-  
619 induced sliding can occur in both channelised and distributed systems, and emphasise that dye  
620 breakthrough curves therefore are problematic for reliably determining changes in subglacial  
621 system configuration. Indeed, changes in efficiency of the non-channelised drainage system  
622 may account for observed velocity changes during the late melt season (Andrews et al., 2014).  
623 Finally, our knowledge of subglacial channel size is very much limited to those seen  
624 emerging at glacier fronts, which can be up to a few 10s of metres in width and may already  
625 have been subject to partial collapse, and to the very few direct observations beneath glaciers  
626 (Gulley et al., 2012a; Fig. 2D,E). Away from the most marginal outlets of the Greenland Ice  
627 Sheet, we have very little understanding of ice sheet subglacial hydrology where ice is thicker  
628 and overburden pressure higher. This has important implications for inferring the form, length  
629 scales and connectivity of subglacial systems.

630  
631 In Antarctica the vast scales, limited surface meltwater availability and the logistical  
632 challenges of mounting field campaigns have prompted the application of alternative research  
633 strategies, which have in turn revealed rather different elements of the subglacial hydrological  
634 system. Many studies of Antarctic subglacial hydrology assume dominance of distributed  
635 drainage (Ashmore and Bingham, 2014). Geophysical techniques complemented by borehole  
636 studies revealed saturated sediments and high basal water pressures at the bed of the Whillans  
637 ice stream, stimulating a widely adopted conceptual model of fast ice stream flow facilitated  
638 by saturated soft beds (Alley et al., 1986; Blankenship et al., 1986; Engelhardt et al., 1990;  
639 Engelhardt and Kamb, 1997). A widely explored corollary of this model is that loss of  
640 meltwater from the subglacial sediment body, via freeze-on or via re-routing ('water piracy'),  
641 drives slow-down, sticking or stoppage of fast ice flow (Alley et al., 1994; Tulaczyk et al  
642 2000; Christoffersen and Tulaczyk, 2003; Bougamont et al., 2003; Vaughan et al., 2008,  
643 Beem et al., 2014). In the Siple Coast sector, a highly heterogeneous pattern of basal melting  
644 and freezing has been detected and modelled (e.g. Vogel et al., 2003, 2005; Joughin et al.,  
645 2004; Jacobel et al., 2009), with a peak combined discharge across the whole Siple Coast  
646 grounding line ( $\sim 300 \text{ m}^3 \text{ s}^{-1}$ , Carter and Fricker, 2012) comparable with the much smaller but  
647 supraglacially-fed Leverett Glacier in west Greenland (Bartholomew et al., 2011b). Beneath  
648 the Rutford Ice Stream, ice-penetrating radar and seismic methods identify a patchy mosaic  
649 (kilometre-scale) of saturated deforming sediments and ponded water bodies (e.g. King et al.,  
650 2004; Smith et al., 2007; Murray et al., 2008). The arrangement of the inferred ponded 'free  
651 water' is consistent with small cavities or with broad, shallow canals, in the order of 10 m  
652 wide and 0.1 m deep, incised into the soft bed (King et al., 2004; Murray et al., 2008).

653  
654 There are recent indications that, in the marginal zone, subglacial channels may in fact be  
655 developed in Antarctica despite virtually no contributions of surface melt to the subglacial  
656 environment. Distributed canals and subglacial channels are interpreted under Thwaites  
657 Glacier based on radar scattering signatures (Schroeder et al, 2013), and a transition zone  
658 between distributed and channelised drainage is interpreted around 50 km from the grounding

659 line. Le Brocq et al. (2013) similarly interpret grounding zone channelised meltwater, argued  
660 to be responsible for sub-ice shelf channels via grounding line meltwater plumes at the  
661 Filchner-Ronne Ice Shelf. Radio echo-sounding and satellite laser altimetry reveal hundreds  
662 of ponded, subglacial lakes under both the West and East Antarctic Ice Sheets, with typical  
663 length-scales of 5-10 km (Siegert et al., 2005; Carter et al., 2007; Smith et al., 2009). Repeat  
664 imaging has shown that these lakes periodically drain and fill, episodically delivering large  
665 volumes of water through the subglacial system of the ice sheet interior (Gray et al., 2005;  
666 Wingham et al., 2006; Fricker et al., 2007). Delivery is inferred to be via channels with  
667 network lengths of 10s-100s km (Wingham et al., 2006; Wright et al., 2012; Wolovick et al.,  
668 2013), though with no confirmed topology. Drainage occurs from lake basin to basin with  
669 timescales for individual events of months to years and associated maximum discharge  
670 estimates of  $40\text{-}50\text{ m}^3\text{ s}^{-1}$  (Wingham et al., 2006; Fricker et al., 2007). Wolovick et al. (2013)  
671 describe subglacial lake networks terminating in regions of basal accretion ice (Bell et al.,  
672 2011a) which indicate stability of an internally confined hydrological system over 10s kyr.

673  
674 While Bell et al. (2011b) have identified regions of accretion ice beneath Petermann Glacier,  
675 in Greenland, similar extensive interior hydrological systems to those in Antarctica have yet  
676 to be identified beneath the Greenland Ice Sheet. Recent detection of *subglacial* lakes in  
677 Greenland via radio echo-sounding and remotely sensed ice surface elevation changes have  
678 been linked to the *supraglacial* meltwater system as sources of lake recharge via  
679 hydrofracture and englacial routing, and as triggers for lake drainage (Palmer et al., 2013;  
680 Howat et al., 2015; Willis et al., 2015). Willis et al. (2015) observed the recharge of a  
681 subglacial lake in conjunction with surface meltwater inflow into crevasses bordering the lake  
682 basin. Howat et al. (2015) theorise an increase in subglacial drainage efficiency in response to  
683 increased surface melting as a trigger for lake drainage. The abundance of surface meltwater  
684 and its supply to the bed of the Greenland Ice Sheet may then explain the relative dearth of  
685 subglacial lakes in comparison to the Antarctic Ice Sheets, such that a larger proportion of  
686 Greenland is underlain by efficient subglacial drainage, restricting the formation and  
687 maintenance of subglacial lakes (Palmer et al., 2013). The steeper ice surface profile of the  
688 Greenland Ice Sheet in comparison to the Antarctic Ice Sheet may also hamper subglacial lake  
689 formation through its influence on hydraulic potential (Livingstone et al., 2013).

### 690 691 *3.3. Contemporary ice sheet hydrology: summary*

692 Seasonal changes in drainage efficiency on the Greenland Ice Sheet appear to be similar to the  
693 seasonal development of an efficient channelised system seen in alpine glaciers. The extent of  
694 channelisation inferred from observations in west Greenland is, however, limited to around 50  
695 km inland from the ice sheet margin. Hence, exporting knowledge of alpine glacier hydrology  
696 to larger systems may introduce a bias into interpretation of observations, since a clear  
697 correlation between drainage system form, pressure distribution and drainage efficiency has  
698 not yet been established. Structural controls on the form of the englacial system and on the  
699 initiation of surface-to-bed connections, including the drainage of supraglacial lakes, raise the

700 possibility of stable locations for subglacial recharge on an inter-seasonal timescale. A  
701 growing proportion of the Greenland Ice Sheet is characterised by supraglacial lakes, the  
702 drainage of which influences ice dynamics and subglacial hydrology on at least a local scale.  
703 The physical form of the subglacial drainage system in response to such delivery, and its  
704 capacity to leave a landform imprint, remains unknown. Sequential drainage of subglacial  
705 lakes in Antarctica, however, hints at an internal, non-climatic control on the mode and timing  
706 of subglacial drainage, and the relative roles of the internal glacial system versus surface  
707 meltwater production in governing drainage processes on contemporary ice sheets remain  
708 poorly understood.

709

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

711 Studying the geological products of past ice sheet hydrological systems offers a broader-scale  
712 and longer-term view compared to contemporary observations. Glaciated landscapes also  
713 provide direct ‘access’ to the palaeo-subglacial domain. Meltwater landforms have long been  
714 documented at range of scales (Jamieson, 1863; Högbom, 1885, 1892; Mannerfelt, 1945;  
715 Ives, 1958; Soyez, 1974; Borgström, 1989; Dyke, 1993; Kleman et al., 1992; Russell et al.,  
716 2001; Mäkinen, 2003, Burke et al., 2008, Livingstone et al., 2010; Margold et al., 2011). They  
717 have widely been used to infer past ice sheet geometries and dynamics, based on relatively  
718 simple assumptions regarding their formation. Meltwater landforms and sediments have,  
719 however, been less often used to infer properties of the (palaeo) *hydrological* system (though  
720 see, for example, Shreve, 1985; Brennand and Shaw, 1994; Piotrowski 1997; Lowe and  
721 Anderson 2002; Shaw, 2002; Boulton et al., 2007a, b; Piotrowski et al., 2009; Boulton et al.,  
722 2009; Nitsche et al., 2013; Phillips and Lee, 2013; Storrar et al., 2014a; Burke et al., 2015;  
723 Lee et al., 2015; Livingstone et al., 2015).

724

725 Meltwater in supra-, sub- and proglacial environments has the capacity to create landforms;  
726 these subsequently have different preservation potential depending on the domain and the  
727 glacial phase (advance or retreat) in which they formed (see Mannerfelt, 1945; Sugden and  
728 John, 1976; Evans, 2003; Bennett and Glasser, 2009; Benn and Evans, 2010, for overview).  
729 Subaerially, meltwater channels are carved at the surface and margins of glaciers as well as in  
730 the glacier fore-field, and shorelines are eroded by wave action and accompanying processes  
731 in glacial lakes. Sediments are deposited where water loses the capacity to carry them farther:  
732 in kame terraces, on outwash plains and in deltas. In en- and subglacial environments, the  
733 pressure field is expected to influence the routing of meltwater and associated erosion and  
734 deposition, in turn determining the spatial distribution of meltwater landforms (Shreve, 1972,  
735 1985; see Section 2.2). Subglacial streams erode meltwater channels, deposit eskers (these  
736 might also form in en- or supraglacial channels; Fitzsimons, 1991; Russell et al., 2001, Burke  
737 et al., 2008, 2009), or create more complex and composite features like meltwater corridors  
738 (Rampton, 2000). Subglacial meltwater channels and tunnel valleys are often seen to be  
739 evidence of Nye-channels, whereas eskers are taken to represent the sedimentary infill of  
740 former R-channels (Clark and Walder, 1994).

741

#### 742 *4.1. Ice-marginal meltwater landforms*

743 Lateral and submarginal meltwater channels document drainage of supraglacial meltwater  
744 along the glacier margin, where ice has been pinned against a topographic slope. These  
745 channels tend to be relatively small (metre-scale in cross-section) and occur in well-developed  
746 series on high ground (Kleman et al., 1992; Dyke, 1993; Greenwood et al., 2007; Margold et  
747 al., 2011, 2013b). With increasing size of the channels (tens of metres in cross-section) their  
748 sinuosity increases and connections to the bed in the form of subglacial chutes or short esker  
749 segments appear; these are usually considered to be submarginal, at the lateral margin but  
750 beneath the ice surface (Mannerfelt, 1949; Greenwood et al., 2007; Syverson and Mickelson,  
751 2009; Lovell et al., 2011; Margold et al., 2011, 2013b). It has traditionally been inferred that  
752 lateral channels form where the ice margin was cold based, preventing downward percolation  
753 (Kleman et al., 1992; Dyke, 1993; Kleman and Borgström, 1996). Where lateral meltwater  
754 channels are a dominant landform (parts of Scandinavia, the Canadian Arctic and Cordillera),  
755 cold-based or polythermal ice can be expected to have prevailed during deglaciation  
756 (Borgström, 1989; Dyke, 1990; Kleman et al., 1992; Dyke, 1993; Sollid and Sørbel, 1994).  
757 However, lateral meltwater channels have been reported to form at the margins of decaying  
758 warm-based Alaskan glaciers (Syverson and Mickelson, 2009) and our palaeo-glacial  
759 interpretations must thus be made with caution.

760  
761 Historically, much attention has been paid to glacial lakes reconstructed from fossil shorelines  
762 and perched deltas (Jamieson, 1863; Gavelin and Högbom, 1910; Frödin, 1925; Charlesworth,  
763 1955; Ives, 1960; Jansson, 2003). However, the focus of these studies has been on  
764 reconstructing past ice margin geometry and few attempts have been made to link the volume  
765 and evolution of glacial lakes to the rest of the ice sheet hydrological system. This is true also  
766 of other ice-marginal landforms such as outwash plains or kames (Clague and Evans, 1993;  
767 Evans et al., 1999; Livingstone et al., 2010; Kehew et al., 2012a).

#### 768 769 *4.2. Subglacial meltwater landforms*

770 Glaciohydrological theory and observations at the margins of contemporary ice masses tell us  
771 that subglacial meltwater must be channelised, at least in parts. Palaeo-subglacial meltwater  
772 channels (also termed tunnel valleys or tunnel channels) are widely detected in the landform  
773 record (e.g. Sissons, 1958; Walder and Hallet, 1979; Sugden et al., 1991; Huuse and Lykke-  
774 Andersen, 2000; Rains et al., 2002; Lowe and Anderson, 2003; Greenwood, 2008; Hughes,  
775 2009; Kehew et al., 2012b; Fig. 3), though they can be difficult to distinguish from their  
776 proglacial counterparts and extensions (Benn and Evans, 2010, p. 296). Often characterised  
777 by an anastomosing topology and undulating long-profiles, individual channel segments may  
778 be 100s metres – kilometres in length, whilst connecting to form more extensive networks  
779 (Greenwood et al., 2007; Kehew et al., 2012b). The scale envelope of subglacial channels is  
780 vast, the smallest of decimetre-metre scale, and their morphology, topology and their  
781 exposure in sediment sections indicate ephemeral, braiding systems in relatively small,  
782 disconnected patches (e.g. Walder and Hallet, 1979; Sharp et al., 1989; Piotrowski, 1999;  
783 Greenwood, 2008; Hughes, 2009). In contrast, subglacial channels on Antarctic palaeo-ice

784 stream beds reach hundreds of metres depth (Nitsche et al., 2013), similar to the tunnel  
785 valleys 2-3 km wide, 100s m deep and up to 100 km long associated with near marginal  
786 drainage of the Fennoscandian and Laurentide Ice Sheets (Cutler et al., 2002; Sandersen et al.,  
787 2009; Stewart and Lonergan, 2011; Stewart et al., 2013). Whilst the large Antarctic channels  
788 cut into crystalline bedrock, the tunnel valleys at the margins of the palaeo-ice sheets of the  
789 northern hemisphere are mostly incised in sediments. The latter case might be associated with  
790 'canals' theorised by Walder and Fowler (1994) and Ng (2000) while subglacial meltwater  
791 channels cut into bedrock might be more closely associated with Nye channels (see Section  
792 2.2.3.).

793  
794 Given the hidden nature of the contemporary subglacial environment and the range of scales  
795 concerned, it is difficult to find appropriate modern analogues or to witness active formation  
796 of the types of meltwater channels observed in the palaeo records. As a consequence, the  
797 mode of formation of these landforms remains uncertain (Ó Cofaigh, 1996; Kehew et al.,  
798 2012b), whether a product of steady-state down-cutting, a single high magnitude or multiple  
799 lower magnitude erosional events. Given such uncertainties, meltwater channels in the  
800 landform record have often been used as a simple indicator of palaeo-flow directions, or  
801 proxy for ice surface slope direction, rather than for specific hydrological processes.

802  
803 At an intermediate scale, a number of sculpted forms (referred to as p-forms, plastically  
804 sculpted forms, Dahl, 1965; or s-forms, sculpted forms, Kor et al., 1991) appear to be the  
805 products of erosion by water flow, in cases pressurised (e.g., Gray, 1981; Shaw, 1988; Glasser  
806 and Nicholson, 1998). However, there is considerable uncertainty in these processes (Glasser  
807 and Bennett, 2004; Bennett and Glasser, 2009, p. 146-147; Benn and Evans, 2010, p. 272-  
808 277), and whether a primarily meltwater or ice flow origin is responsible.

809  
810 Esker ridges are typically 10s-100s metres wide and 1-10s m high (Aylsworth and Shilts,  
811 1989; Hebrand and Åmark, 1989; Syverson et al., 1994; Huddart et al., 1999). They occur in  
812 segments or 'beads' a few kilometres long, sometimes with interlocking ridges and often  
813 terminating in deltaic sediments (Warren and Ashley, 1994; Hebrand and Åmark, 1989;  
814 Mäkinen, 2003). Esker segments line up into systems that stretch – apparently continuously –  
815 for hundreds of km over the beds of the former northern hemispheric Pleistocene ice sheets  
816 (Prest et al., 1968; Aylsworth and Shilts, 1989; Clark et al., 2000; Boulton et al., 2001; Storrar  
817 et al., 2013; Storrar et al., 2014b; Fig. 4). Whether this reflects an extensive, continuous  
818 drainage network under large portions of the ice sheet, or incremental (time-transgressive)  
819 formation shortly behind a retreating margin, is a matter of some uncertainty.

820  
821 Sediment architecture, esker geometry and distribution, and relationships to channel theory  
822 have been widely used to inform this debate. In the central parts of the Laurentide Ice Sheet,  
823 continuous ridges 10s-100s km in length have been observed, with fans or deltas only at their  
824 distal ends (Brennand, 1994; Brennand and Shaw, 1996; Brennand, 2000). Sediment clast  
825 characteristics documenting long transport distances and low palaeo-flow variability

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

840

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

853

854 Esker properties have perhaps provided the closest links between the palaeo record and  
855 developing theory of ice sheet hydrological systems (e.g. Shreve, 1985; Clark and Walder,  
856 1994; Boulton et al., 2007a,b). Despite incomplete understanding of their formation  
857 processes, eskers have traditionally been the landform of choice for broad-scale inferences on  
858 past ice sheet retreat patterns (De Geer, 1929; Dyke and Prest, 1987; Kleman et al., 1997;  
859 Boulton et al., 2001; Putkinen and Lunkka, 2008; Greenwood and Clark 2009a, b; Clark et al.,  
860 2012; Margold et al., 2013a, b).

861

862 Subglacial lakes are expected to leave a geological trace, but research into their identification  
863 in the palaeo-record is in its infancy. Livingstone et al. (2012) provide criteria, largely  
864 sedimentological, for palaeo-subglacial lake identification, and recent modelling efforts  
865 attempt to predict their location (Evatt et al., 2006; Livingstone et al., 2013). Such predictions  
866 test earlier hypothesised subglacial lake locations (e.g. Shoemaker, 1999; Munro-Stasiuk,

867 2003; Christoffersen et al., 2008; Lesemann and Brennand, 2009) and offer an improved  
868 picture of an integrated palaeo-glaciohydrological system.

869

#### 870 *4.3. Spatial distribution*

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

876

##### 877 *4.3.1. Controls exerted by subglacial thermal organisation and ice dynamics*

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

888

889 That the meltwater record appears to be nearly ubiquitous at ice sheet scale points to the  
890 superimposition of deglacial mass melting signatures upon the earlier thermal organisation of  
891 the ice sheet. At a regional scale, the thermal history of the ice sheet may have a more  
892 significant role. Palaeo-ice streams on the Canadian Shield, identified on the basis of mega-  
893 scale glacial lineations and shear-margin moraines, operated during deglaciation in what had  
894 previously been a cold-based interior of the Laurentide Ice Sheet. Here, eskers have been  
895 found to be less frequent than in the surrounding areas (Kleman and Glasser, 2007; Storrar et  
896 al., 2013; Margold et al., 2015). It has been suggested that distributed meltwater drainage  
897 within the ice streaming zones may explain the lower frequency of eskers in these areas  
898 (Storrar et al., 2014b; Livingstone et al., 2015). Ice dynamics might affect not only the  
899 frequency of esker occurrence, but also their spatial pattern. While coherent and uniformly  
900 oriented esker networks, such as those in Keewatin, Labrador or parts of Scandinavia, have  
901 generally been interpreted as indicating frontally retreating ice sheet margins (though see  
902 Sections 4.2, 5.2 and 5.3), more chaotic esker patterns have variously been explained as a  
903 result of (1) a complex ice streaming history in a given area (Victoria Island; Storrar et al.,  
904 2014b), (2) interaction of a low-profile ice sheet with a variable topography (Canadian Arctic  
905 Archipelago; Storrar et al., 2014b), (3) subglacial drainage governed by local potentiometric  
906 surfaces in stagnant ice or subglacial drainage in stagnant ice fed into broader subglacial  
907 channels (southern Alberta, Brennand, 2000), or (4) by a transition from a coherent active ice

908 surface to a more complex and stagnant ice body (Margold et al., 2013b).

909  
910 Frozen areas of ice sheet beds have traditionally been identified by the presence of relict  
911 surfaces and landscapes, and an *absence* of landforms associated with subglacial melt  
912 delivery and soft-sediment deformation. Such sectors occupy considerable areas of ice sheets  
913 (Kleman and Glasser, 2007) and can be taken to indicate that we should not expect a  
914 (sub)glacial hydrological system to be widely developed in an ice sheet. Lateral meltwater  
915 channels have been considered indicative of cold-based or polythermal ice (Borgström, 1989;  
916 Dyke, 1990; Kleman et al., 1992; Dyke, 1993; Sollid and Sørbel, 1994). However, they have  
917 been reported to form at the margins of decaying warm-based Alaskan glaciers (Syverson and  
918 Mickelson, 2009), and surface-to-bed downcutting of a supraglacial dendritic network  
919 through the ‘cut-and-closure’ process (Vatne, 2001; Gulley et al., 2009) has been observed  
920 under cold-based glaciers in Svalbard (Naegeli et al., 2014). How the thermal condition of an  
921 ice sheet generates, directs and diverts the hydrological system is a fundamental question, but  
922 is non-trivial to extract from the landform record.

#### 923 924 *4.3.2. Geological controls*

925 At a regional to ice sheet scale, eskers have been correlated with areas of hard bed lithologies,  
926 whilst incised channels occur where ice sheets covered less resistant bedrock or thick  
927 sedimentary sequences (Clark and Walder, 1994; Boulton et al., 2009). Soft, deformable beds  
928 are argued to favour distributed drainage through a network of shallow ‘canals’. As a  
929 generality, these associations hold true for glaciated terrains of the northern hemisphere and,  
930 indeed, it has been argued that channel development at a local scale intimately follows local  
931 substrate geological variability (e.g. Sandersen and Jørgensen, 2012) and may be strongly  
932 controlled by the interaction with the local groundwater system (Piotrowski, 1997; Piotrowski  
933 et al., 2009). However, numerous local landform arrangements display complexity beyond the  
934 general landform–substrate associations. Many channel systems incise hard bedrock  
935 (Clapperton, 1968; Sugden et al., 1991; Nitsche et al., 2013; Jansen et al., 2014) whilst well-  
936 developed esker networks occur locally over thick sediments and sedimentary bedrock (Flint,  
937 1930; Greenwood, 2008; Storrar et al., 2014b; Burke et al 2015). Furthermore, esker chains  
938 may transform into deeply cut channels, and vice-versa (e.g. Johansson, 1995, 2003;  
939 Hättestrand and Clark, 2006; Greenwood, 2008; Hughes, 2009; Margold et al., 2013a;  
940 Atkinson et al., 2014), whilst eskers are commonly observed on the floors of large tunnel  
941 valleys (see Kehew et al., 2012b). The landform record points to conditions conducive to both  
942 esker and channel formation within the same inter-connected system; the length-scale of  
943 synchronously operational networks is, however, a crucial unknown.

944  
945 A different landsystem model has been advanced for the Antarctic continental shelves (e.g.  
946 Wellner et al., 2001, 2006; Graham et al., 2009), where the beds of former ice streams display  
947 a common landform succession from subglacial meltwater channels cut deeply into crystalline  
948 bedrock (Lowe and Anderson, 2002; Nitsche et al., 2013) to increasingly elongate glacial



949 lineations across and beyond the seaward transition to sedimentary bedrock. Here, any traces  
950 of subglacial drainage disappear and it is hypothesised that meltwater is delivered through  
951 small canals or a deforming till aquifer (Ó Cofaigh et al., 2002; Graham et al., 2009;  
952 Noormets et al., 2009; cf. Murray et al., 2008). A single reported meltwater channel occurs in  
953 the outer shelf sedimentary environment of the Ross Sea (Alonso et al., 1992; Wellner et al.,  
954 2006) and eskers are notably absent. Meltwater landforms are not especially common in other  
955 marine glaciated environments, with the exception of some rare examples of eskers (e.g. Todd  
956 et al., 2007; Todd and Shaw, 2012; Feldens et al., 2013) and the tunnel valleys of the North  
957 and Barents Seas (Huuse and Lykke-Andersen, 2000; Stewart and Lonergan, 2011;  
958 Bjarnadóttir et al., 2012; Stewart et al., 2013). It should, however, be noted that that much  
959 research in marine environments has focused on palaeo-ice stream troughs, where fast  
960 flowing ice may impede channelised drainage. In contrast, the inter-ice stream banks are  
961 typically heavily scoured by icebergs, which might have destroyed any evidence of drainage.  
962

#### 963 *4.4. Temporal properties*

964 At an ice sheet scale, meltwater landforms in the palaeo-record are typically interpreted as a  
965 product of stable flow, and thereby used to reconstruct palaeoglaciological regimes  
966 corresponding to years, decades or longer (e.g. Kleman et al., 1992; Dyke 1993; Kleman et  
967 al., 1997; Clark et al., 2000; Boulton et al., 2001; Clark et al., 2012; Margold et al., 2013a, b).  
968 However there remains much uncertainty over the timescales of landform creation (Kehew et  
969 al., 2012b). Meltwater channel erosion may be a product of incremental but steady erosion  
970 (Boulton et al., 2009), single high-magnitude events (Brennand and Shaw 1994; Cutler et al.,  
971 2002; Hooke and Jennings, 2006), or repeated erosional episodes over time periods from ~100  
972 years to multiple glaciations (Jørgensen and Sandersen, 2006; Sandersen et al., 2009; Nitsche  
973 et al., 2013). Sedimentological and morphological arguments are used to support each  
974 alternative and it is possible that all are viable models in different contexts. Contrary to the  
975 typical long-term view of palaeo glaciology, englacial eskers have been observed to form in  
976 single jökulhlaup events in Iceland (Russell et al., 2001; Burke et al., 2008). On the basis of  
977 its single, point source and its specific internal architecture, Burke et al. (2012a) interpret an  
978 esker in southern British Columbia as a possible evidence of supraglacial lake drainage,  
979 whilst the ‘Labyrinth’ network of massive channels and other high-magnitude flow traces in  
980 the Dry Valleys, Antarctica, have been interpreted as a result of subglacial floods with  
981 estimated discharges of  $1.6\text{--}2.2 \times 10^6 \text{ m}^3 \text{ s}^{-1}$  and velocities of  $11\text{--}15 \text{ m s}^{-1}$  (Denton and  
982 Sugden, 2005; Lewis et al., 2006). Short duration and high-magnitude drainage events are  
983 likely to leave a geological and geomorphological product. Reliably distinguishing between  
984 such landforms and those formed under steady flow conditions remains a challenge for the  
985 palaeoglacial community.  
986

#### 987 *4.5. Palaeo-ice sheet hydrology: summary*

988 Thanks to the extensive occurrence of the meltwater landforms on the past ice sheet beds, the  
989 meltwater record has been widely used for palaeoglaciological reconstructions of Pleistocene  
990 ice sheets, but mainly as a simple indicator of palaeo-flow directions, ice surface slope

991 directions, and ice margin positions. The mode of formation, especially for landforms of  
992 subglacial origin (subglacial meltwater channels, eskers), remains uncertain and different  
993 interpretative models have been developed for different regions subjected to past glaciations  
994 (cf. Brennand, 2000 and Mäkinen, 2003).

995  
996 Meltwater landforms display a vast range of scales, contrasting topologies and arrangements  
997 (extensive esker networks, esker-channel-esker successions, chaotic distribution of channels).  
998 Landform creation is subject to processes beyond the glaciohydrological: processes of  
999 sediment mobilisation and deposition. Understanding of the operation of these processes and  
1000 related controls on the landform distribution, such as the underlying geology (i.e. 'boundary  
1001 conditions' in theoretical and numerical modelling approach) is incomplete. Similarly, the  
1002 temporal and spatial scales of formation remain unclear: whether single or multiple drainage  
1003 events, or gradual steady-state drainage; and how spatially connected an operational drainage  
1004 network may be. Nevertheless, the palaeo record offers a spatial context which is difficult to  
1005 extract from the contemporary systems: the whole ice sheet bed is exposed with the resulting  
1006 landforms often easily accessible, which is in stark contrast to the indirect character of study  
1007 of contemporary ice sheet hydrology.

1008  
1009 Figure 5 summarises an ice sheet's hydrological system from a palaeoglaciological  
1010 perspective, in terms of the geological record that we have today to decipher. This record is a  
1011 composite resulting from different ice sheet stages over time and therefore also likely a  
1012 composite of different drainage regimes. Although the glacial meltwater record as a whole  
1013 might encompass tens to hundreds of thousands of years, individual systems within this likely  
1014 represent a regime spanning only tens to thousands of years. The time associated with the  
1015 processes responsible for individual landform creation might be even shorter, on the order of  
1016 seasons to decades. Deciphering these temporal scales from a composite record of a whole  
1017 glaciation (e.g. Figs. 3-5) and, furthermore, bridging this information with the timeframes  
1018 represented in Figure 1, presents a serious challenge.

1019

## 1020 **5. Discussion**

1021 A driving agenda for glaciohydrological research is to better understand the dynamic  
1022 responses of ice flow to hydrological processes, and to better predict likely responses of ice  
1023 sheets and ice sheet sectors to increased surface melting under atmospheric warming  
1024 scenarios. To this end, we need a clearer understanding of the dominant form of the meltwater  
1025 drainage system, how pressure and discharge are related, and under what physical conditions  
1026 meltwater is delivered through the glacial hydrological system. Based on our preceding  
1027 reviews, it is clear that the existence of one single mode of meltwater drainage in an ice sheet  
1028 is implausible. A key objective for the glaciohydrological community must be to determine  
1029 over which spaces, and which timescales, different drainage modes operate.

1030

1031 Achieving this goal is complicated by the fact that i) the physical processes of water flow in  
1032 an ice sheet, ii) spatial variability and spatial controls upon water drainage, and iii) temporal

1033 variability and controls on drainage, are each uncertain to some degree, may appear to be  
1034 contradictory between different fields of research, and are all intimately interlinked and  
1035 interdependent. Our review presents three major challenges to which we draw attention:  
1036 1. Contemporary observations and glaciological theory suggest that the occurrence of low-  
1037 pressure, channelised drainage (cf. Röthlisberger, 1972) should be relatively limited. The  
1038 palaeo record contains widespread evidence of channelised drainage, of highly varied  
1039 topologies. Moreover:

- 1040 - Long-distance operational connectivity of a channelised system has been  
1041 interpreted from some palaeo-ice sheet beds and long-distance operational  
1042 connectivity has also been found in contemporary Antarctica.
- 1043 - The apparent lack of a widespread ‘distributed’ system in the palaeo record may  
1044 be a product of poor preservation, or poorly developed tools and templates for  
1045 identification and interpretation of non-channelised drainage.

1046 2. The palaeo record is a cumulative product of formative processes and environments.  
1047 However, understanding of the timescales of landform development is limited, as is  
1048 understanding of the potential of hydrological regimes of different magnitudes for  
1049 landform creation. Conversely, events of different magnitudes and timescales are known  
1050 from contemporary ice sheets, but their long-term significance is poorly understood.

1051 3. Fundamental scale contrasts between research fields are a major disconnect, both in  
1052 spatial and temporal scale. A ‘large channel’ to a contemporary glaciologist might be 10-  
1053 100 m wide; to a palaeoglaciologist a large channel would be 1-2 km in width.

1054  
1055 In the context of these challenges, we consider below the form, spatial extent and variability,  
1056 and temporal controls on the drainage system. These issues are highly interconnected and  
1057 cannot wholly be treated apart. We highlight the sensitivity of the spatial extent of subglacial  
1058 channel maintenance to temporal variability. We emphasise a need to understand  
1059 geomorphological as well as hydrological processes in order to make robust use of the  
1060 glaciofluvial landform record, and we consider the degree of consistency of the ice sheet  
1061 drainage system across geographical and geological domains. Our discussion is stimulated by  
1062 our varied perspectives and we attempt to present a cross-disciplinary view of the controls,  
1063 distribution and evolution of subglacial drainage systems.

1064  
1065 *5.1. What is the form of the drainage system?*  
1066 Theoretical glaciology, and much of the language employed concerning ice sheet hydrology,  
1067 conceptualises two contrasting frameworks which describe how water might be delivered  
1068 through the subglacial domain (e.g. Röthlisberger, 1972; Kamb, 1987): a channelised, or a  
1069 distributed system. These are respectively associated with efficient, low pressure conduits  
1070 capturing flow into a dendritic network, and a poorly connected network of water-filled voids  
1071 at high pressure; simplistically, theory predicts two contrasting drainage topologies.

1072  
1073 Support for a widespread, low pressure Röthlisberger channelised system does not appear to  
1074 be emerging from studies of contemporary ice sheets. Efficient meltwater delivery around the

1075 Greenland ice margin is assumed to reflect channelisation *close* to the margin (Bartholomew  
1076 et al., 2011a; Chandler et al., 2013), the interior comprising a high pressure domain  
1077 (Meierbachtol et al., 2013) consistent with the high pressure basal environment inferred to be  
1078 widespread in Antarctica (Alley et al., 1986; Engelhardt et al., 1990; King et al., 2004). This  
1079 is in stark contrast to the palaeo record of Quaternary glaciations, in which there is almost  
1080 ubiquitous evidence of the channelisation of subglacial water. The topologies, however, of  
1081 both eskers and palaeo-meltwater channels are highly varied (Figs. 3, 4) and it is by no means  
1082 clear that (all) these landforms should represent low pressure Røthlisberger drainage. Any  
1083 disconnect between modern and palaeo evidence of subglacial drainage should be perceived  
1084 in the context of the high degree of uncertainty as to whether observations taken to indicate a  
1085 Røthlisberger-type network (low pressure, dendritic) should indeed be interpreted as such.

1086

1087 On modern ice sheets, interpretation of the *form* of a drainage network cannot be made with  
1088 confidence based only on the *efficiency* and speed of transit (cf. Gulley et al., 2012b; Cowton  
1089 et al., 2013); when all that can be monitored is the input and output, the exact pathways and  
1090 modes of delivery remain unknown. Indeed, emerging evidence suggests that efficient  
1091 drainage could be viable in high pressure settings (Fudge et al., 2008). The widespread  
1092 channelisation of meltwater indicated by the palaeo record comprises highly varied forms and  
1093 network topologies. It is widely assumed that eskers reflect a Røthlisberger system and a  
1094 broadly dendritic arrangement is often taken as support, though the theory has no explicit  
1095 prediction of stream ordering, merely that water should be conveyed from high to low  
1096 pressure. Modifications to classic Røthlisberger theory (e.g. Hooke et al., 1990) allow for  
1097 non-dendritic channelised topologies and, indeed, eskers exhibit contrasting topologies  
1098 depending on the scale at which they are viewed (Fig. 4A). Hewitt's (2011) implementation  
1099 of channel theory predicts lateral channel spacing in broad agreement with that of Laurentide  
1100 Ice Sheet eskers documented by Storrar et al. (2013). Canals and Nye channels have typically  
1101 been considered part of an inefficient (slow) system of drainage (e.g. Walder and Fowler,  
1102 1994; Fountain and Walder, 1998). This could be consistent with the varied topologies of  
1103 palaeo channels: often discontinuous, branching and chaotic in contrast to the form of esker  
1104 networks (Fig. 3A, 4A). However, it is difficult to explain large palaeo channels that clearly  
1105 have been eroded by high-discharge subglacial streams (e.g. Booth and Hallet, 1993; Lowe  
1106 and Anderson, 2003; Nitsche et al., 2013; Jansen et al., 2014) as part of inefficient, poorly  
1107 connected drainage, and it is similarly problematic to account for intermittent esker – channel  
1108 – esker connectivity over long distances.

1109

1110 Further, there is an *apparent* lack of distributed drainage in the palaeo record. This is arguably  
1111 a consequence of poor preservation under a retreating marginal zone (see Fig. 5), of  
1112 inadequate templates for its identification and interpretation in the landform record, and of the  
1113 immense challenges in extending sedimentological observations of distributed drainage to the  
1114 ice sheet scale. Finally, the appearance of ubiquity is likely to a large degree an artefact of a  
1115 time composite record. The geological record is, inherently, an amalgam of processes and

1116 products over an extended time domain and, in the case of eskers and meltwater channel  
1117 landforms, arguably a product of ablation and ice sheet deglaciation. Any apparent mismatch  
1118 in drainage topology between contemporary and palaeo perspectives is therefore, in part, a  
1119 likely result of the contrasting time domains and ‘life-cycle’ stages of the respective ice  
1120 sheets.

1121  
1122 Figure 6 summarises a variety of drainage topologies which are supported, to varying degrees,  
1123 by observations from both modern and palaeo environments and implied by glacial hydrology  
1124 theory. Whilst these conceptual sketches focus on the ‘shape’ the drainage takes, topology can  
1125 be considered a simplified proxy for physical process descriptions (see Section 2.2). We  
1126 suggest that each of these topologies (and possibly more) may be viable and appropriate in  
1127 different contexts, and argue that it is more instructive to conceptualise the subglacial system  
1128 as such a multiplicity of modes, rather than the traditional binary channelised (low pressure,  
1129 dendritic) or distributed (high pressure, inefficient) modes. The outstanding challenge, then, is  
1130 to assign a spatial scale to these multiple modes and determine their spatial and temporal  
1131 occurrence.

### 1132 1133 *5.2. Spatial extent and variability of effective drainage*

1134 It is impossible to determine the relative dominance of any of the topologies (and respective  
1135 drainage processes) illustrated in Figure 6 without a sense of scale. If one were to assume that  
1136 eskers are the best representative of the dominant subglacial hydrological configuration, and  
1137 if one were to adopt a regional perspective, one might conclude that topology 6A (Fig. 6) –  
1138 i.e. a low-order dendritic, low pressure flow-capturing channel system – were the normal  
1139 mode of drainage. Esker formation, however, is yet to be observed on our modern templates  
1140 for the system – Greenland and Antarctica – and their formative scales are equivocal. If we  
1141 alternatively consider the palaeo-meltwater channel record then topology 6B is more typical.  
1142 Change the spatial scale of focus (see the boxed areas on Figure 4, for example) and  
1143 topologies 6C or 6D would appear more appropriate.

1144  
1145 Given that the full palaeo-ice sheet bed is now exposed, the palaeo record might be best  
1146 placed to enlighten the spatial scales of different drainage modes. However, any such attempt  
1147 is thwarted by the difficulties of determining the temporal controls on landform generation,  
1148 and the time domain represented by the landforms we now observe. There is an order of  
1149 magnitude difference in inferred operational length of subglacial conduits depending on  
1150 whether one adopts a continuous tunnel fill model (e.g. Brennand, 1994; Brennand and Shaw,  
1151 1996; Brennand, 2000), such as for the Laurentide systems hundreds of kilometres in length,  
1152 or a time-transgressive beaded esker model (e.g. Hebrand and Åmark, 1989; Mäkinen, 2003)  
1153 in which channelised drainage is confined to a marginal zone a few 10s km at most. The  
1154 former invokes a wide ablation zone with high surface melt rates, typical of the last  
1155 deglaciation (e.g. Carlson et al., 2008, 2009) and which may therefore account for a  
1156 contrasting length scale to modern observations. However, Livingstone et al. (2015) find that  
1157 modelled subglacial pathways routed according to Shreve (1972; Section 2, Eqs. 1, 2) provide

1158 a robust match with observed esker geometry only up to ~10-20 km from the margin. Beyond  
1159 this distance, a small minority of eskers match modelled subglacial pathways for >100 km.  
1160 The majority of apparently lengthy esker networks deviate from modelled channel routing,  
1161 suggesting ice sheet geometry (hydraulic gradient) has changed and that these long-distance  
1162 networks must be formed time-transgressively.

1163

1164 Modern glaciological observations and current theory rather support limited channel lengths,  
1165 with an implied maximum distance of efficient subglacial drainage of about 50 km  
1166 (Bartholomew et al., 2011a; Chandler et al., 2013), in the marginal zone where there is a  
1167 sufficiently steep ice surface slope (cf. Clark and Walder, 1994; Walder and Fowler, 1994;  
1168 Meierbachtol et al., 2013). Recent modelling by Meierbachtol et al. (2013) argued that even  
1169 with high and sustained meltwater input in the interior (cf. Brennand, 2000; Carlson et al.,  
1170 2008, 2009), the inhibited melting back of subglacial conduit walls is not sufficient to  
1171 overcome creep closure; rather, highly pressurised water would be driven into a surrounding  
1172 distributed system and the conduit would not be sustained. A related approach by Dow et al.  
1173 (2014) models subglacial conduits under high pressure ~70 km from the Greenland Ice Sheet  
1174 margin, and finds that conduit maintenance is improbable under these conditions.

1175

1176 The operational length of subglacial channels remains a significant uncertainty. We must  
1177 consider the potential that deglaciation of mid-latitude ice sheets provided different  
1178 hydrological conditions to today and that our modern ice sheet observations may, in part, be  
1179 inappropriate analogues. The physics presented in Röthlisberger (1972; i.e. transient forms of  
1180 Eq. 5, Section 2.2.1) underlie numerous theoretical developments and numerical explorations  
1181 of subglacial channelisation (e.g. Hooke et al., 1990; Walder and Fowler, 1994; Clarke, 1996;  
1182 Schoof, 2010; Meierbachtol et al., 2013). We can explore conceptual scenarios for subglacial  
1183 channel extent using a simple numerical model (e.g. Meierbachtol et al., 2013) of a straight  
1184 channel over a hard bed (with no explicit physics for what happens outside the bounds of  
1185 Röthlisberger channelisation). This provides a ‘favourable case’ for channel maintenance,  
1186 making no attempt to address how a drainage system evolves under different regimes, only to  
1187 illustrate at what point conventional channel theory becomes insufficient such that an  
1188 additional or alternative physical system is needed. Under steady-state conditions with a  
1189 constant meltwater input, this model shows that subglacial conduit water pressures lie below  
1190 overburden pressure for any input discharge and *any given distance* from the ice margin under  
1191 a range of ice surface profiles (Fig. 7; cf Meierbachtol et al., 2013; Helanow et al., 2015; and  
1192 demonstrated in Supplementary Material). In the strictly steady-state case, a conduit could be  
1193 maintained 10s-100s km up ice from the margin.

1194

1195 However, even small transient increases in water input drive subglacial water pressures  
1196 towards ice overburden and beyond the limits of Röthlisberger physics (Meierbachtol et al.,  
1197 2013); conduit wall melt-back cannot keep pace with an increase in discharge. If a conduit  
1198 survives at high pressures, a dendritic, flow capturing network is unlikely to develop, since

1199 any flow across a pressure gradient would only serve to reverse the differential and drive flow  
1200 back where it came from. *Temporal* controls on the system therefore appear to be paramount.  
1201 The same model (see Supplementary Material) indicates that a low surface profile typical of  
1202 outlets of the Laurentide Ice Sheet (e.g. Beget, 1986; Clark, 1992) may in fact sustain a lower  
1203 pressure channel than a high profile (e.g. Greenland Ice Sheet) at ~40 km from the ice margin.  
1204 We could, therefore, anticipate some divergence between our landform and contemporary  
1205 indications of channel length. At greater distances, however, channel maintenance is  
1206 significantly more difficult under low surface profiles. This is in partial keeping with the  
1207 conventional view that channelisation is best developed under higher surface slopes  
1208 (Röthlisberger, 1972; Walder and Fowler, 1994), but at a certain proximity to the margin, a  
1209 slow creep closure rate under thin ice becomes significant. We finally draw attention to the  
1210 role of conduit size, which, for example in this simple Röthlisberger-based model, does not  
1211 diminish monotonically upstream but rather narrows to a knickpoint part way along its viable  
1212 (< overburden pressure) length (Fig. 7). The sensitivity of any conduit constrictions to  
1213 reaching overburden pressure will arguably be most important in governing the conduit's  
1214 stability and, potentially, the length of a persistent channel that may leave a recognisable  
1215 landform imprint. We note that with a range of different channel length scenarios (10s-100s  
1216 km) this knickpoint lies c. 20-50 km upstream from the ice margin.

1217

1218 We consider that the weight of evidence supports a spatial extent of a dendritic channelised  
1219 topology (i.e. classic Röthlisberger, Fig. 6A) limited to a few 10s km, likely <50km.

1220 However, we do not exclude the possibility of significant efficient drainage in some form of  
1221 channel system at greater spatial scales inward of this limit. Our present understanding of  
1222 such drainage is that it would be at relatively high pressures and in a likely non-dendritic  
1223 system. We highlight three outstanding challenges in constraining the spatial extent of  
1224 different drainage modes:

1225 i. The topology and processes of significant drainage in a high pressure (at/near  
1226 overburden) setting are poorly understood, encompassing a wide variety of process  
1227 models (Section 2.2.2; e.g. Weertman, 1972; Kamb, 1987; Walder and Fowler, 1994;  
1228 Creyts and Schoof, 2009). Targeted, systematic and repeat geophysical monitoring of  
1229 different hypothesised drainage environments in both Greenland and Antarctica, for  
1230 example saturated sedimentary domains, regions of supraglacial melt penetration to  
1231 the bed, and subglacial lake drainage routes, will be profitable in better constraining  
1232 high pressure drainage modes.

1233 ii. The spatial extent of subglacial drainage must not only consider individual systems,  
1234 but also the overall distribution of where drainage does, or does not operate at all: i.e.  
1235 the (subglacial) thermal organisation of the whole ice sheet and the supply of  
1236 meltwater. The thermal state of an ice sheet bed comprises a complex spatial pattern  
1237 that evolves over time (Joughin et al 2004; Hubbard et al., 2009; Pattyn, 2010),  
1238 leading to a time-transgressive composite being preserved in the palaeo record  
1239 (Kleman and Glasser, 2007). The extent and magnitude of subglacial melt and

1240 drainage at any one snapshot in time may be significantly less than the impression  
1241 given by the time-integrated palaeo-landform record.  
1242 iii. Conversely, the palaeo record reflects not only hydrological processes, but also  
1243 geomorphic and sedimentological. Lack of geomorphic power or sediment starvation  
1244 may mean a period of ice sheet drainage has no legacy, and our geological record  
1245 under-represents the palaeo hydrological system (cf Burke et al., 2015).

### 1247 *5.3. Temporal variability and controls on drainage*

#### 1248 *5.3.1 Transient meltwater regimes*

1249 The significance of the time domain has been raised in a number of contexts throughout our  
1250 reviews and the Discussion so far, both in relation to the temporal variability of meltwater  
1251 drainage and in relation to the time composite landform record. In its pure form, Röthlisberger  
1252 theory illustrates that channel maintenance 100s km from the ice margin is feasible under  
1253 truly steady input regimes (Fig. 7 and Supplementary Material). However, it is difficult to  
1254 envisage a requisite steady input setting to maintain a conduit system below overburden  
1255 pressure. Basal melt dominated systems, such as in Antarctica where there is no/little surface  
1256 input, might be such an environment where short term temporal variability is removed,  
1257 though the flux is so low (e.g. basal melt of 1-10 mm a<sup>-1</sup>, Joughin et al., 2004; discharge  
1258 across the entire Siple Coast 300 m<sup>3</sup> s<sup>-1</sup>, Carter and Fricker, 2012) that it is difficult to explain  
1259 large-scale channelisation in this context. Schoof (2010) also suggests that below a critical  
1260 discharge, channels would not be initiated, but rather that the system would stay in a  
1261 distributed form. Alternatively, retention of surface meltwater in supra-, en- and subglacial  
1262 stores (Willis et al., 1990; Fountain, 1993; Harper et al., 2010; Cowton et al., 2013) may  
1263 deliver water to relevant points of the bed at a constant rate. This, together with a surface melt  
1264 rate so high that diurnal and seasonal contrasts are significantly dampened, could perhaps  
1265 elicit subglacial conduit development in a pseudo steady-state fashion. Carlson et al. (2009)  
1266 model palaeo-surface mass balance for the Laurentide Ice Sheet and suggest that at the height  
1267 of deglaciation, parts of the southern margin were thinning at rates of 9 m yr<sup>-1</sup>. It is possible  
1268 that such extreme scenarios are not well covered by current drainage theory, and that we are  
1269 consequently not even conceptually familiar with a resultant drainage system. It is such mass  
1270 melting environments with which long esker systems have been traditionally associated  
1271 (Brennand, 2000).

1272  
1273 Most realistic systems are, however, transient: supraglacial meltwater dominated systems vary  
1274 diurnally, are highly seasonal with an annual winter shut-down and spring restart, and exhibit  
1275 multi-annual variability. Moreover, both supraglacial and subglacial systems experience  
1276 episodic events such as lake drainages superimposed on other more regular cycles. Following  
1277 Meierbachtol et al. (2013), we find that only very low daily and seasonal fluctuations in  
1278 meltwater input (e.g. 2% daily, 20% seasonal) are able to yield a conduit pressure response  
1279 akin to a steady regime and stay below overburden pressure (albeit at a high fraction; see  
1280 Supplementary Material). At 40 km, overburden is reached when daily oscillations in input  
1281 exceed ~14% of initial discharge; this amplitude threshold for reaching overburden is even



1282 lower at greater distances. Sensitivity to transience in meltwater supply is increased,  
1283 therefore, with distance from the ice margin: there is a dependence of *spatial* extent and  
1284 spatial stability on the *temporal* properties of the system, and all the more so under low ice  
1285 surface profiles.

1286

1287 If transience in the supply of meltwater to the subglacial system inevitably drives water  
1288 pressures towards overburden, then it is clear that one cannot assume that an ‘average’ of  
1289 transient fluctuations approximates a steady-state, and that addressing aspects of ice sheet  
1290 hydrology on palaeo or multi-annual timescales cannot ignore short-term processes. It implies  
1291 also that an incipient conduit up-ice from the marginal zone needs *some* supply, constantly, to  
1292 be sustained. Mass melting of the Quaternary ice sheets (e.g. Carlson et al., 2009) could  
1293 perhaps enable this, but today winter closure would preclude any interior conduit maintenance  
1294 under this model (cf. Gulley et al., 2009a; Jarosch and Gudmundsson, 2012). This could go  
1295 some way to offering an explanation for the contrasting evidence of conduits between the  
1296 landform record of deglaciation and modern observations.

1297

### 1298 *5.3.2 Transience in geomorphology*

1299 How transience in meltwater input is manifested in the landform record of the subglacial  
1300 system is far from clear. The relative importance of different timescales of drainage for  
1301 landform building is largely unknown from either a process or palaeo perspective. An attempt  
1302 to better constrain the timeframes represented by glaciofluvial landforms, and whether a  
1303 certain landform type or network therefore reflects a more common or rare state of the palaeo-  
1304 drainage system, is urgently required. A model for long, continuous esker systems (e.g.  
1305 Brennand, 2000) would call for gradual sedimentation along their whole length. Röthlisberger  
1306 theory indicates that such a conduit is very unstable, would exist at or close to overburden  
1307 pressure and is only possible under a permanent, high flux as close to steady as possible (cf  
1308 Fig. 7). Eskers would then reflect one single ice sheet state and stability of the hydrological  
1309 system for the duration of esker building. Repeated exploitation of the same supra- and  
1310 englacial pathways season after season (Fountain et al., 2005a, b; Catania et al., 2008; Benn et  
1311 al., 2009; Catania and Neumann, 2010) make such stability a possibility, though these  
1312 observations are not from great distances from the ice margin. However, development of a  
1313 dendritic network, the very characteristic often used to support an extensive operational  
1314 network, does not seem physically plausible under the predicted high pressure setting. The  
1315 transience of a realistic meltwater input is more consistent with the alternative end-member  
1316 model of esker formation, whereby sedimentation is confined to short (a few hundred metres  
1317 long), thick beads at the ice front (e.g. Mäkinen, 2003). Sediment architecture in such  
1318 segments argues for highly seasonal drainage, with high spring and summer sedimentation  
1319 rates, as well as positional stability of the drainage pathways over successive seasons  
1320 (Mäkinen, 2003; Ahokangas and Mäkinen, 2014). An outstanding question relates to the  
1321 source of these high sediment volumes: a transport mechanism (by water?) is required beyond  
1322 the confines of the esker deposit itself, implying greater extent of the hydrological system

1323 than manifested by the depositional landform.

1324

1325 Ultimately we must determine under which meltwater regimes landform development occurs.  
1326 The extent to which we can attribute landform creation to stable seasonal drainage evolution,  
1327 to annually occurring or repeated episodic events, to semi-independent series of inter-linked  
1328 events, to isolated catastrophic outburst events, or even to multiple glaciations is palpably  
1329 unclear (cf. Ó Cofaigh, 1996; Kehew et al., 2012b). We may expect higher magnitude events  
1330 to have stronger geomorphic power (cf. Hjulström, 1935); the question is whether infrequent  
1331 events of high magnitude are more geomorphologically effective than those of more moderate  
1332 magnitude and recurrence (cf. Wolman and Miller, 1960). Could our landform record of  
1333 palaeo-ice sheets in fact reflect only short-term, one-off high magnitude events (Figs. 8, 9)?  
1334 Supraglacial lake drainages with discharge magnitudes on the order of  $10\,000\text{ m}^3\text{ s}^{-1}$  (Das et  
1335 al., 2008) are both seasonal and episodic in occurrence, following the seasonal up-ice  
1336 expansion of ablation, whilst the timing and frequency of individual lake drainages varies  
1337 inter-annually (Fitzpatrick et al., 2014). Antarctica's subglacial lakes are known to burst  
1338 episodically, some draining and re-filling over sub-decadal timescales (e.g. Wingham et al.,  
1339 2006; Fricker et al., 2007) but with any cyclicity likely independent of wider climatic or  
1340 seasonal forcing. Meltwater landform corridors containing plunge pools and point-sourced  
1341 eskers (e.g. Rampton, 2000; Burke et al., 2012a, b) and deeply incised bedrock gorges (e.g.  
1342 Jansen et al., 2014) attest to sudden fluxes of water to and along the bed, and have been  
1343 attributed to supraglacial events. Burke et al. (2008, 2010) interpret distinct esker sediment  
1344 packages as products of subglacial outburst flooding, whilst large palaeo-channel systems  
1345 such as the Labyrinth channels have been linked to catastrophic subglacial outbursts (Denton  
1346 and Sugden, 2005; Lewis et al., 2006). The landform record of palaeo-ice sheets undoubtedly  
1347 contains traces of high magnitude events. What, then, in our landform record is the imprint of  
1348 the exceptional, and what of the normal?

1349

1350 If we were to suppose that the 'normal' condition of subglacial hydrology would not form  
1351 landforms, but these were identifiable as a product of cumulative exceptional events, then our  
1352 landform record would not be representative of the dominant mode of ice sheet drainage. We  
1353 may envisage a scenario where, despite the apparent ubiquity in the palaeo record of some  
1354 kind of channelised drainage, a topology somewhat akin to Fig. 6E (largely distributed,  
1355 disconnected and small-scale) dominates, but our palaeo record is skewed by the higher  
1356 landform creation potential of systems looking more like 6A-D which occur only temporarily  
1357 and sporadically. Figure 9 conceptualises the uncertainty regarding landforming potential in  
1358 response to meltwater drainage. Not only can various different drainage regimes (e.g. marked  
1359 i-iii on Fig. 9) deliver a long term discharge that is difficult to discern from a stable regime  
1360 (iv), but it is uncertain where, within such drainage regimes, most geomorphic work may take  
1361 place. Does most landform creation (red dashed lines on Fig. 9) only occur during infrequent  
1362 high discharge events (e.g. scenario E, Fig. 9) or during more regular, moderate discharge  
1363 cycles (e.g.  $P_{\max}$ ,  $P_{\min}$ )? Or, alternatively, does the landform record simply represent an

1364 ‘average’ condition (scenario A)? The uncertainty in the process-product relationship makes  
1365 interpretation of the timescales represented by the landform imprint challenging. These  
1366 uncertainties are further compounded if we consider the opposing directions of geomorphic  
1367 work: erosion or deposition. In using an esker as a proxy for palaeo drainage, our product  
1368 only represents the depositional phase of that drainage, and we assume that its distribution is  
1369 limited only by the operation of the hydrological system and not by the geomorphic. Sediment  
1370 starvation would preclude depositional landform development (cf Burke et al., 2015) just as  
1371 effectively as a drainage regime either too weak to drive conduit development or too powerful  
1372 to deposit its sediment load. Our geomorphic products are indirect proxies that are more  
1373 spatially and temporally restricted than their parent hydrological system. Resolving the  
1374 timeframes and discharge conditions responsible for geomorphic work is paramount to our  
1375 ability to use the palaeo record as proxies of palaeo-ice sheet hydrology.  
1376

#### 1377 *5.4. Scales of internal consistency and variability*

1378 At local scale, the landform record displays varied and irregular patterns. This leads us to ask  
1379 whether the fundamental modes of drainage vary on these spatial scales, or whether it is  
1380 simply the landform expression that changes. Varied assemblages of meltwater landforms  
1381 occur within small areas (e.g. Hättestrand, 1998; Hättestrand and Clark, 2006; Margold et al.,  
1382 2013a; Atkinson et al., 2014; Turner et al., 2014), with esker–channel–esker downstream  
1383 sequences (Johansson, 2003; Hättestrand and Clark, 2006; Greenwood, 2008; Hughes, 2009;  
1384 Margold et al., 2013a; Atkinson et al., 2014), eskers on the floor of large channels  
1385 (Hättestrand and Clark, 2006; Kehew et al., 2012b), or meltwater corridors containing an  
1386 assortment of landforms (Rampton, 2000; Burke et al., 2012b). In light of this, the topologies,  
1387 or modes of drainage, represented in Figure 6 appear to be gross simplifications. We could  
1388 draw two contrasting conclusions from the varied expression of the landform record regarding  
1389 system functionality. Local controls such as micro-topography, substrate properties or  
1390 discharge availability could drive a switch to an alternative hydrological system: the  
1391 downstream change in landforms would reflect downstream (or temporal) changes in the  
1392 fundamental mode of meltwater drainage. Alternatively, the *same* hydrological system  
1393 passing over/through varying external boundary controls would simply be manifested in  
1394 different ways in the landform imprint.  
1395

1396 Physically, one might expect there to be fundamentally different systems on different  
1397 substrates (cf Clark and Walder, 1994), i.e. the former of the two above scenarios. If different  
1398 landforms and landform configurations reflect different levels of efficiency (e.g. eskers as  
1399 ‘fast drainage’; palaeo channels as ‘slow systems’), then in a continuous functional network  
1400 (assumed) how would water be conveyed through an inefficient system situated in between  
1401 two efficient systems? Alternatively, different landforms should be interpreted as a reflection  
1402 of geomorphological rather than hydrological efficacy. Recent theoretical efforts attempt to  
1403 integrate systems of different efficiencies, but the focus of theory has not hitherto been to  
1404 investigate the influence of different boundary controls and the effect of boundary transitions.  
1405 Thus far we lack clear understanding of the independent role of separate boundary conditions

1406 (e.g. sediment, soft bedrock or hard bedrock). It is therefore difficult to say from a theoretical  
1407 point of view how these components might integrate within a unifying or overriding system.

1408

1409 Regional boundary controls can be expected to impact both glaciohydrological processes and  
1410 the efficacy and form of geomorphic work, therefore dictating both local and short-term  
1411 operability and the regional-to-continental scale organisation of drainage within an ice sheet.  
1412 Substrate geology and topography (e.g. Clark and Walder, 1994), sediment fluxes (e.g. Burke  
1413 et al., 2015), subglacial thermal organisation (e.g. Joughin et al., 2004; Kleman and Glasser,  
1414 2007), climatically driven surface melt availability (e.g. Carlson et al., 2009; Storrar et al.,  
1415 2014a) and the mode and stability of regional ice surface to bed connectivity undoubtedly  
1416 influence the functionality of the subglacial system, its large-scale spatial variability, and our  
1417 ability to use landforms as proxies for palaeo-ice sheet hydrology. To some degree, boundary  
1418 controls are specific to an individual ice sheet's geography or stage of evolution. However,  
1419 the large scale patterns of the palaeo record attest to some element of self-organisation of the  
1420 ice sheet hydrological system. At an ice sheet-wide scale, the spatial distribution, arrangement  
1421 and topology of esker and palaeo-channel networks are similar between the different northern  
1422 hemisphere palaeo-ice sheet beds (Fig. 10) suggesting that they can be treated similarly.  
1423 Different landsystem expressions of the hydrological system may correspond to different  
1424 regional controls but each unit still displays a degree of internal organisation or regularity,  
1425 and the spatial scale of this organisation is qualitatively similar (Fig. 10). There thus appears  
1426 to be a level of system control, regardless of the local external boundary controls. Whether the  
1427 hydrological system is truly self-organising, or is perhaps rather a response to or reflection of  
1428 self-organisation of ice flow, taps into one of the fundamental uncertainties in present  
1429 understanding of ice sheet behaviour.

1430

### 1431 *5.5. Subglacial systems: controls, distribution and evolution*

1432 In attempting to understand the processes, controls upon, variability of and products of the  
1433 glaciohydrological system, there are clearly numerous domains we must bridge: spatial and  
1434 temporal domains, supraglacial and subglacial environments, palaeo-ice sheet records and  
1435 modern ice sheet observations. There is considerable difficulty in doing so: contrast again the  
1436 timescales, the research logistics and research approaches represented in Figures 1 and 5.  
1437 Figure 11 conceptualises an ice sheet's first-order hydrological characteristics in both space  
1438 and time domains. A high pressure environment is found throughout the ice sheet, and may  
1439 take on several forms with different levels of efficiency (Fig. 6); moderately efficient drainage  
1440 could occur through a system close to overburden pressure. Røthlisberger-type channelised  
1441 drainage develops in an outer zone (C in Fig. 11) whose inward limit is governed by ice  
1442 surface slope, discharge and the temporal variability of surface melt delivery to the bed, and  
1443 whose wider spatial distribution must reflect the thermal state of the ice sheet and patterns of  
1444 organisation therein. Pockets of efficient delivery may occur within the interior high pressure  
1445 zone (cf. Antarctic drainage between lake basins) whilst short-term high magnitude events  
1446 occur throughout the ice sheet's history and likely increase through deglaciation, though it is  
1447 not yet clear how far and for how long these could expand a zone of channelised drainage.

1448 These are likely to be especially capable of geomorphic work, but the most effective  
1449 discharge regimes for erosional and depositional landform creation, and the degree to which  
1450 these are influenced by local boundary controls, remain uncertain.

1451  
1452 The outer, channelised zone (C) occupies a similar subglacial envelope as does the ablation  
1453 zone in the supraglacial domain; systems of the channelised domain are generally considered  
1454 supraglacial melt dominated (Bartholomew et al., 2011b; Cowton et al., 2013). Whilst the ice  
1455 surface slope and stress field are fundamental in meltwater routing, many theoretical  
1456 formulations argue that discharge regimes are a critical parameter in drainage system  
1457 evolution (see Section 2), and will inevitably govern the efficacy of the geomorphic processes  
1458 of erosion and deposition. In the supraglacial melt zone, both the background discharge and  
1459 superimposed event frequency are high. However, what actually happens to water at its point  
1460 of entry to the bed is still little understood. What is the immediate response of subglacial  
1461 pressure and drainage topology to supraglacial water input? If conduit growth is inhibited  
1462 under overburden pressures then stable channelised systems will be restricted to the near-  
1463 marginal zone irrespective of surface melt discharges: enhanced areal supply of meltwater  
1464 through deglaciation would not drive a commensurate expansion of the subglacial channelised  
1465 zone. Alternatively, if efficient drainage can operate at or close to overburden conditions, then  
1466 supraglacial control over the subglacial domain may remain significant throughout the ice  
1467 sheet cycle. In the Antarctic case, where supraglacial melting is negligible, the subglacial  
1468 system must be controlled by a different suite of mechanisms and parameters to those of  
1469 supraglacially-dominated systems. A minimum, continual supply of water to the bed appears  
1470 to be necessary to keep a channel open and operational (Fig. 7C). Antarctica may be better  
1471 represented by an almost entirely high pressure zone (absence of C-zone; cf. Fig. 6E), with  
1472 some internal pockets of focussed drainage, likely also at high pressure.

1473  
1474 We stress, finally, that there are stark differences in the climatic, topographic, ‘life-cycle  
1475 stage’ and ice dynamic settings of the Greenland, Antarctic and northern hemisphere palaeo-  
1476 ice sheets. Meltwater of supraglacial origin likely constitutes a significant portion of  
1477 meltwater drained through both the present day Greenland Ice Sheet and the deglaciating  
1478 Quaternary ice sheets, dominating discharge regimes, while negligible surface melt in  
1479 Antarctica likely produces a distinctly different hydrology. Furthermore, the hydrology of the  
1480 ephemeral Pleistocene ice sheets may have changed substantially during their lifetime,  
1481 resembling present-day Antarctica during the cold climates of growth and Last Glacial  
1482 Maximum phases, whereas the conditions during deglaciation may have resembled an  
1483 extreme version of what can nowadays be observed in SW Greenland. Since the palaeo record  
1484 is a time-integrated composite, it is possible there is a legacy of all these states, and it remains  
1485 possible that none is an appropriate analogue for today’s remnant interglacial ice sheets.  
1486 Moreover, it is clear that the scales exhibited within the palaeo record of ice sheet drainage  
1487 are in considerable excess of those evident today, in terms of landform size (cross-section,  
1488 amplitude), spatial extent and in the temporal domain. The significance of the scale mismatch,  
1489 and the processes responsible for the palaeo record, are yet to be determined. We argue that

1490 there is a degree of self-organisation of the drainage system and this holds between different  
1491 ice sheets (Section 5.4), albeit on the continental spatial scale and a glacial cycle temporal  
1492 scale. The field can and should advance by learning from palaeo and modern observational  
1493 analogues for ice sheet hydrology, but should treat these analogues with care and caution.  
1494

## 1495 **6. Conclusions**

1496 Based on our reviews of theoretical, contemporary observational and palaeo glaciology, we  
1497 have attempted to highlight how understanding of ice sheet hydrology can be advanced by a  
1498 cross-disciplinary approach. Isolated study by different communities working with different  
1499 perspectives risks that objectives become niche and divorced from one another. We have  
1500 attempted to identify and discuss some of the main deficiencies in the contemporary  
1501 understanding of ice sheet hydrology and we offer our key conclusions here.  
1502

1503 Drainage topologies and inferred pressure relationships in both modern and palaeo-ice sheet  
1504 environments differ from a traditional binary conceptualisation of low pressure, dendritic  
1505 channels versus a high pressure distributed system. Instead, we consider drainage topology to  
1506 comprise a multiplicity of forms in an amalgam of drainage modes and processes, each of  
1507 which may be viable and appropriate in different contexts and at different scales. Greenland  
1508 and Antarctic observations indicate efficient drainage at high pressure, and we suggest that  
1509 particular attention should be turned to determining topologies, drainage processes and  
1510 geomorphological products of these regimes.  
1511

1512 These same field observations and numerical modelling suggest that low pressure settings are  
1513 confined close to the margin ( $< \sim 50$  km), and that the spatial extent of a classical channelised  
1514 drainage system may be highly sensitive to the temporal variability in meltwater supply. One  
1515 cannot, therefore, neglect short-term processes even when considering a long-term  
1516 perspective. A further implication is that transient esker sedimentation likely reflects a  
1517 position close to the ice margin.  
1518

1519 Opposing interpretative models for glacial landforms, apparent contrasts with modern ice  
1520 sheet analogues, and the composite, cumulative nature of the record of Pleistocene ice sheets  
1521 hinder robust interpretations of palaeo-hydrological systems. It remains poorly understood  
1522 which scales, magnitudes and frequencies of drainage events are most important for  
1523 geomorphic work in the glacial hydrological system, and therefore which elements of the  
1524 palaeo system are imprinted in our landform record: the normal or the exceptional.

1525 Furthermore, we must determine whether different (subglacial) landforms represent  
1526 fundamentally different hydrological modes, or whether they are simply a different landform  
1527 expression of a unifying hydrological system. The uniformity of spatial patterns across  
1528 palaeo-ice sheet beds argues for a degree of self-organisation of the ice sheet hydrological  
1529 system independent of any local or regional boundary conditions. There is a real need for  
1530 further geomorphic process research, with a physical process basis and which considers the

1531 bed as an active participant in glacial hydrology, independent of the palaeo-ice sheet  
1532 reconstruction paradigms with which glaciofluvial landforms have often been associated.

1533

1534 Glacial hydrology is influenced by the climatic, topographic, ‘life-cycle stage’ and ice  
1535 dynamic settings of each particular ice sheet. Full physical description of meltwater drainage  
1536 modes and regimes, and of their interactivity, is needed for the evaluation of ice sheet  
1537 behaviour on different temporal scales, both in interpretation of palaeo ice dynamics and for  
1538 predicting ice sheet evolution in a future warming world. Linking the processes to the  
1539 products of ice sheet hydrology remains challenging, but vital to these endeavours.

1540

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1554

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2453 222.

2454

## 2455 **Figure captions**

2456 **Figure 1.** (*Portrait, full page width*)

2457 Conceptual diagram of a glacial hydrological system, comprising the supraglacial, englacial  
2458 and subglacial environments. Note that not all components may be applicable in all ice

2459 bodies. Insets depict: A) Water-driven crevasse propagation following van der Veen (2007).  
2460 B) Proposed form of Röthlisberger, Hooke and Nye subglacial channels, where darker colours  
2461 represent higher discharge. C) Main physical factors influencing the opening or closure of a  
2462 subglacial channel, where  $p_i$  and  $p_w$  are the pressure of ice and water, respectively, and  $Q$  is  
2463 discharge through the channel. D) Cross-section of accumulation zone facies (following Nolin  
2464 and Payne, 2007).

2465

2466 **Figure 2.** (*Portrait, full page width*)

2467 Contemporary observations of englacial and subglacial drainage systems. A) Moulin entrance  
2468 on Leverett Glacier, southwest Greenland Ice Sheet, with inflow of supraglacial meltwater  
2469 into the moulin. B) Englacial conduit c.60 m below the ice surface originating from a  
2470 longitudinal crevasse on Matanuska Glacier, Alaska. Note the channel wall roughness. C) Cut  
2471 and closure englacial conduit on Longyearbreen, Svalbard at c. 15 m below the ice surface,  
2472 where the upper channel walls undergo closure due to ice pressure. D) Subglacial  
2473 (Röthlisberger-style) channel c. 60 m below the ice surface of Hansbreen, Svalbard. Note the  
2474 large-scale roughness of channel ice walls. E) Subglacial (Nye-style) channel under  
2475 Hansbreen, Svalbard at a depth of c. 110 m. Note uneven sediment deposition and roughness  
2476 of the channel bed. Photo credit for A: Caroline Clason; B-E: Jason Gulley; B published in  
2477 Gulley et al. (2009b).

2478

2479 **Figure 3.** (*Portrait, column width*)

2480 Meltwater systems in the palaeo-record: meltwater channels. (A) Distribution of meltwater  
2481 channels on the bed of the British-Irish Ice Sheet (after Greenwood, 2008; Hughes, 2009).  
2482 Approximate southern terrestrial limit of the ice sheet in Great Britain during the last  
2483 glaciation (Devensian; in purple, dashed) is drawn after Clark et al. (2012). (B) Subglacial  
2484 meltwater channels of the British-Irish Ice Sheet, in the Vale of Eden, NW England (after  
2485 Hughes, 2009). Note the contrasting topologies compared to Fig. 4A: whereas eskers display  
2486 a low-order dendritic pattern at a regional scale, subglacial meltwater channels form  
2487 anastomosing networks. Note also the difference in scale between esker and palaeo-channel  
2488 systems. Location of panel B is marked by black box in panel A.

2489

2490 **Figure 4.** (*Portrait, full page width*)

2491 Meltwater systems in the palaeo-record: eskers. (A) Eskers in Keewatin, north-central Canada  
2492 (adapted from Storrar et al., 2013). Note the correspondence with the retreating margin of the  
2493 Laurentide Ice Sheet between 11 and 6.5 ka (drawn after Dyke et al., 2003, with isochrons  
2494 in  $^{14}\text{C}$  years). The limit of the Canadian Shield is marked by a yellow line. Rare examples of  
2495 bifurcation or anastomosing are marked by arrows. Boxed areas are discussed in the text. (B)  
2496 Ice sheet wide distribution of eskers on the bed of the North American Ice Sheet Complex.  
2497 CIS – Cordilleran Ice Sheet, LIS – Laurentide Ice Sheet, Last Glacial Maximum limit after  
2498 Kleman et al. 2010. Esker networks (marked in red, adapted from Storrar et al., 2013, are best  
2499 developed on the Canadian Shield (in green, from Wheeler et al., 1996). Location of panel A  
2500 and Fig. 10A is shown by black rectangles.

2501

2502 **Figure 5.** (*Portrait, full page width*)

2503 Conceptual diagram of a palaeo-ice sheet hydrological system and its geological imprint. A)  
2504 Meltwater channels incised into the bed, and esker deposits upon the bed, are both taken as  
2505 geomorphic products of a channelised meltwater drainage system. Porewater flow through a  
2506 till aquifer may be speculated upon, but there are few templates for interpreting distributed  
2507 drainage. Through time, the snapshot shown here is repeatedly imprinted and overprinted by  
2508 each successive change in ice sheet configuration or drainage behaviour, during both advance  
2509 and retreat of an ice sheet. In a landform (plane) view (B), a composite landform product  
2510 represents different drainage regimes and/or different ice sheet configurations over a  
2511 timeframe of  $10^1$ - $10^4$  years. In a stratigraphic (cross-section) view (C), these different regimes  
2512 and/or ice sheet configurations must be deciphered from a stack of units. The timeframes  
2513 represented by individual landforms or sedimentological traces, to a coherent unit, to a  
2514 sequence of units, to a whole palaeo-ice sheet bed encompass the whole spectrum from  
2515 seasons to tens of thousands of years.

2516

2517 **Figure 6.** (*Portrait, full page width*)

2518 A variety of different drainage system topologies may be inferred from observations and  
2519 theory. A) Dendritic channel system which, following classic R-theory, captures flow across a  
2520 pressure gradient and is conceptually considered to drain water ‘efficiently’ in low-pressure  
2521 channels. Observed esker systems strongly resemble this form, at a variety of scales (Figs. 4,  
2522 10). B) A non-dendritic channel network is commonly observed among subglacial meltwater  
2523 channels in the palaeo record, at a variety of scales (Figs. 4, 10). Considered ‘inefficient’  
2524 systems, these are conceptually consistent with canal theory or linked cavity networks. C)  
2525 Flow captured by a main low pressure channel, but locally feeders are non-dendritic (cf.  
2526 pressure observations by Fudge et al., 2008, at glacier scale). D) Theoretical and modelling  
2527 attempts to bridge channelised and distributed models (e.g. Hewitt, 2011) use a simplified  
2528 topology. E) Contemporary observations point to a high pressure environment, non-  
2529 channelised (Meierbachtol et al., 2013), everywhere other than immediate marginal zone. On  
2530 all panels dark blue marks channel conduits at relatively low pressure (with line thickness  
2531 scaling with discharge where this can be safely assumed), red conduits at relatively high  
2532 pressure, light blue dashes indicate drainage areas, and a non-channelised high pressure  
2533 system is represented by red dotted zones. Ice margin in black.

2534

2535 **Figure 7.** (*Portrait, column width*)

2536 Idealised consequences of A) steady and B) variable meltwater input to a subglacial channel,  
2537 following Röthlisberger (1972). In steady input scenario A), a conduit water level can remain  
2538 below overburden pressure at distances 100s km inward from the ice margin. Conduit size  
2539 (cross-sectional area) is smallest close to the ice margin. B) When input discharge varies (e.g.  
2540 diurnally, seasonally), the conduit invariably is driven to overburden pressure; conduit  
2541 pressure remains below overburden pressure for only small windows of time (e.g. at the very

2542 start of a rise in discharge, and when discharge begins to fall before creep-closure re-adjusts  
2543 to a lower input).

2544

2545 **Figure 8.** (*Portrait, column width*)

2546 Peak discharge circles (scaled with circle area) for the Glacial Lake Missoula floods (Hanson  
2547 et al., 2012), the Glacial Lake Agassiz outburst flood (Clarke et al., 2004), the Labyrinth  
2548 channels (Lewis et al., 2006), average discharge of the Amazon River (Vörösmarty et al.,  
2549 1989), Skeiðarárjökull jökulhlaup (Russell et al., 2006), a west Greenland supraglacial lake  
2550 drainage (Das et al., 2008), the Bering Glacier outburst flood (Burke et al., 2010), 2009 peak  
2551 discharge for the Leverett Glacier proglacial outlet (Bartholomew et al., 2011b), total peak  
2552 discharge across the Siple Coast grounding line (Carter and Fricker, 2012) and the Adventure  
2553 trench subglacial lake drainage (Wingham et al., 2006). For reference, the discharge of  
2554 Glacial Lake Missoula is 340,000 times greater than the Adventure trench subglacial lake  
2555 drainage; comparable with the difference in discharge between the Amazon River and a small  
2556 brook.

2557

2558 **Figure 9.** (*Portrait, column width*)

2559 Uncertainty in the correspondence between drainage regime and landform formation. Four  
2560 different drainage regimes (grey lines, i-iv) may all have a similar average state (discharge  
2561 over time), for example equivalent to drainage regime iv, when considered over a long  
2562 timescale. It is uncertain where, with respect to these drainage regimes, most geomorphic  
2563 work takes place. Landforms may represent the ‘average’ condition (A), may represent  
2564 geomorphic work towards the peaks or troughs of periodic discharge regimes ( $P_{\max}$  or  $P_{\min}$ ,  
2565 respectively) or may represent geomorphic work which only takes place under few, isolated,  
2566 high magnitude events (E).

2567

2568 **Figure 10.** (*Portrait, column width*)

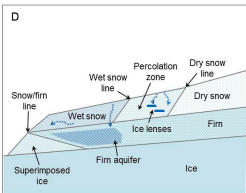
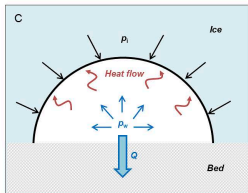
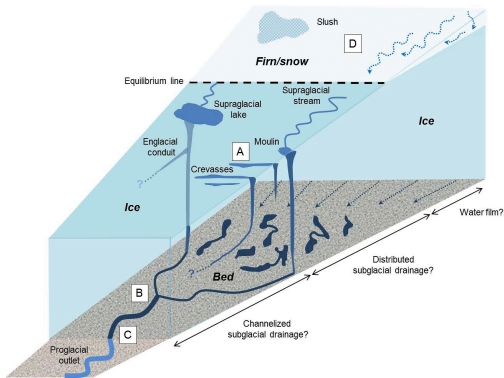
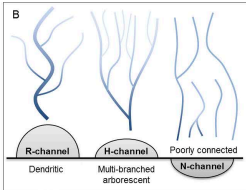
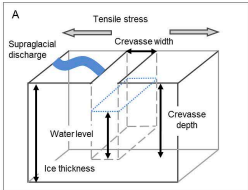
2569 Despite different continental ice sheets, different climate and ice sheet zones, different  
2570 substrate geologies, indeed, different marine isotope stage ice sheets, the scaling and  
2571 organisation of contrasting landform systems in the palaeo-record is broadly similar. (A)  
2572 Eskers in Ontario are marked in red (with filled gaps between individual esker segments,  
2573 adapted from Storrar et al., 2013). Bathymetry of Lake Superior is shown in shades of blue  
2574 (data from NOAA-NGDC) with networks of tunnel valleys visible on the lake floor. Note the  
2575 correspondence with the retreating margin of the Laurentide Ice Sheet between 11 and 8.5 ka  
2576 (drawn after Dyke et al., 2003, with isochrons in  $^{14}\text{C}$  years). (B) Esker networks in Sweden  
2577 (with filled gaps between individual esker segments) are marked in red (provided courtesy of  
2578 Clas Hättestrand) and Elsterian tunnel valleys in Germany and Poland are marked in blue  
2579 (adapted from Huuse and Lykke-Andersen, 2000). Both panels are drawn at the same scale.  
2580 The black dashed line marks the boundary between the hard crystalline rocks of the Canadian  
2581 (A) and the Baltic (B) shields (upper parts of the panels) and softer sedimentary rocks.

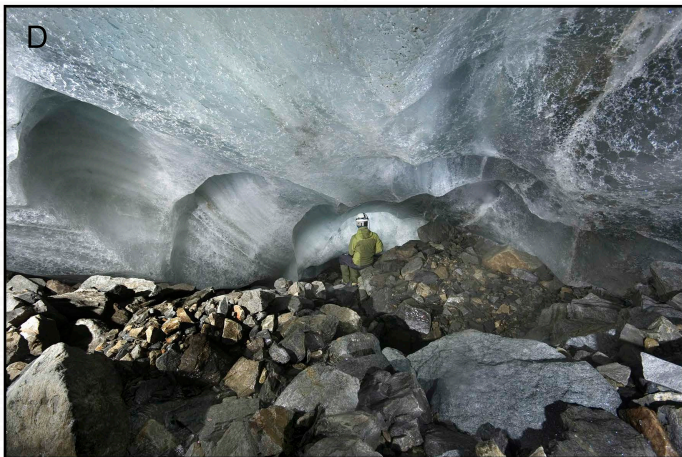
2582

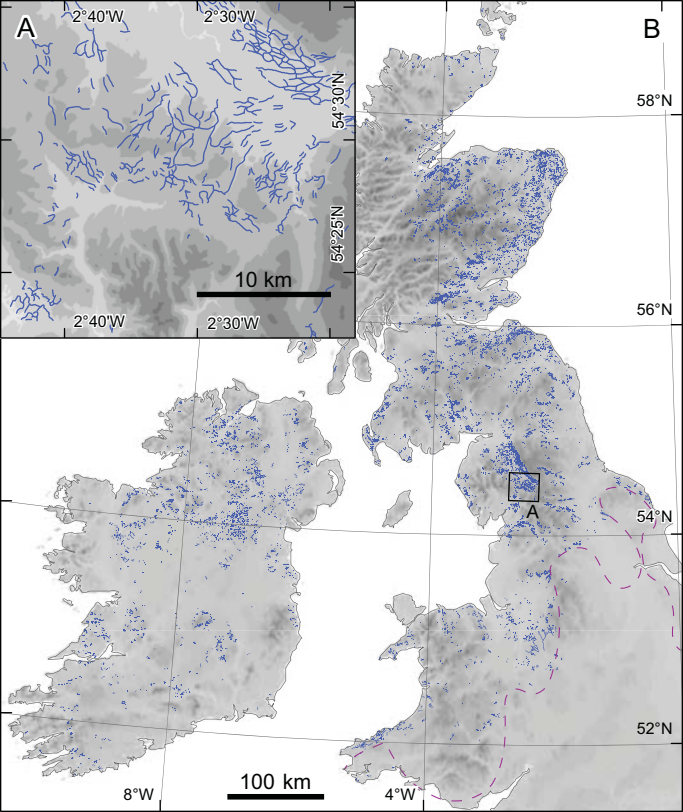
2583 **Figure 11.** (*Portrait, column width*)

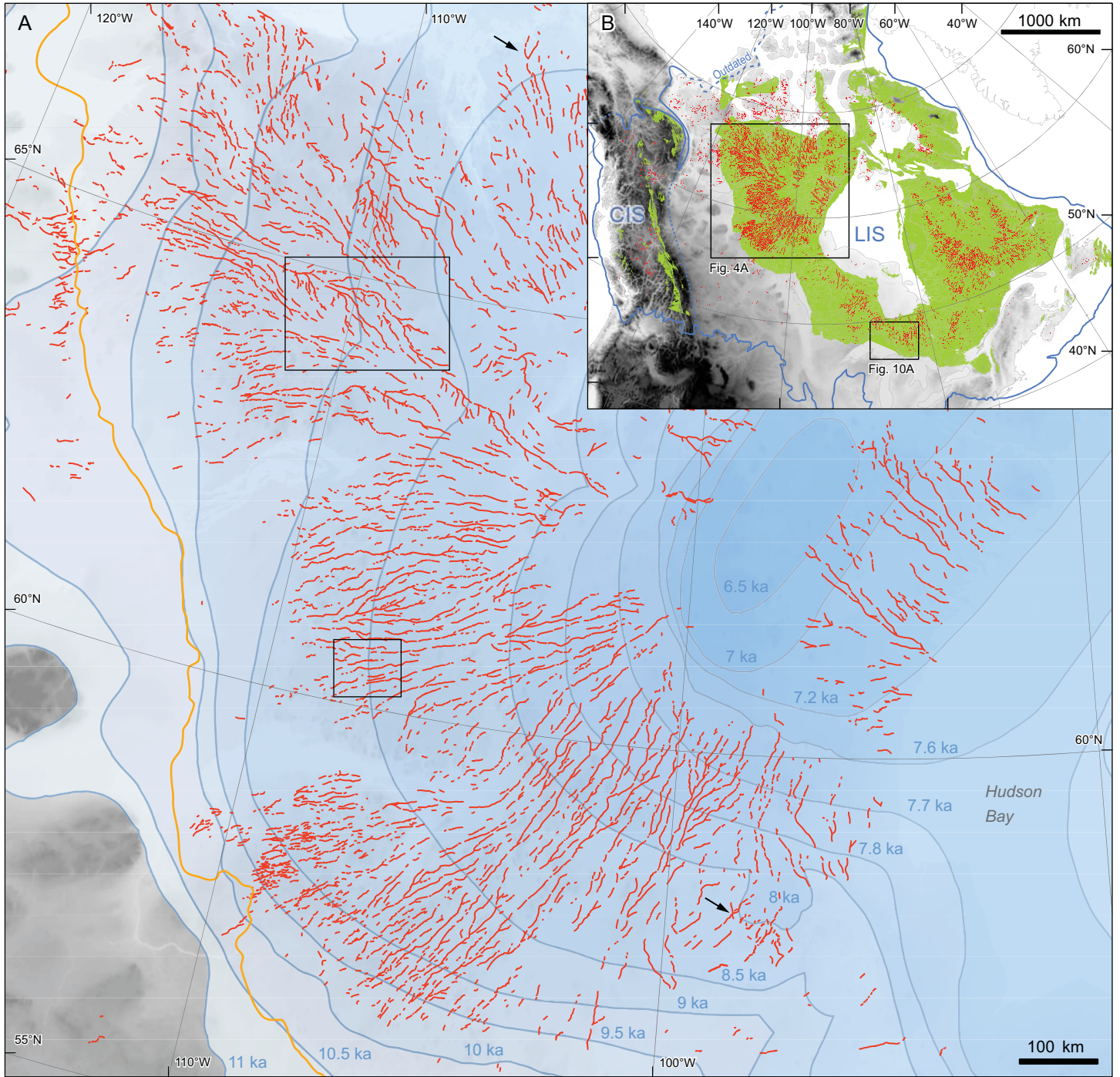


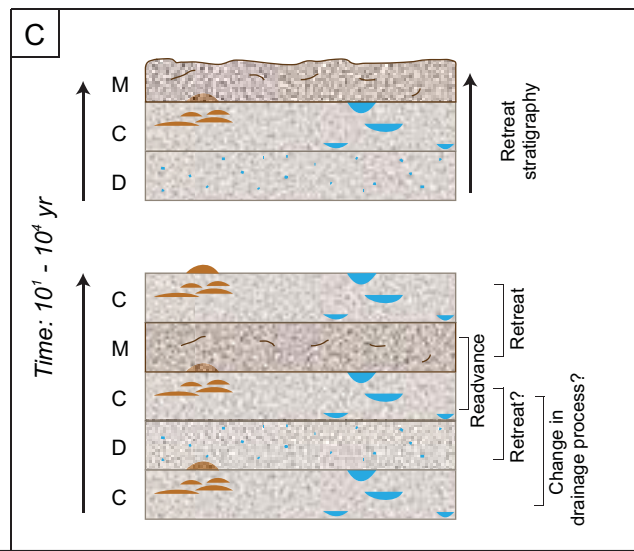
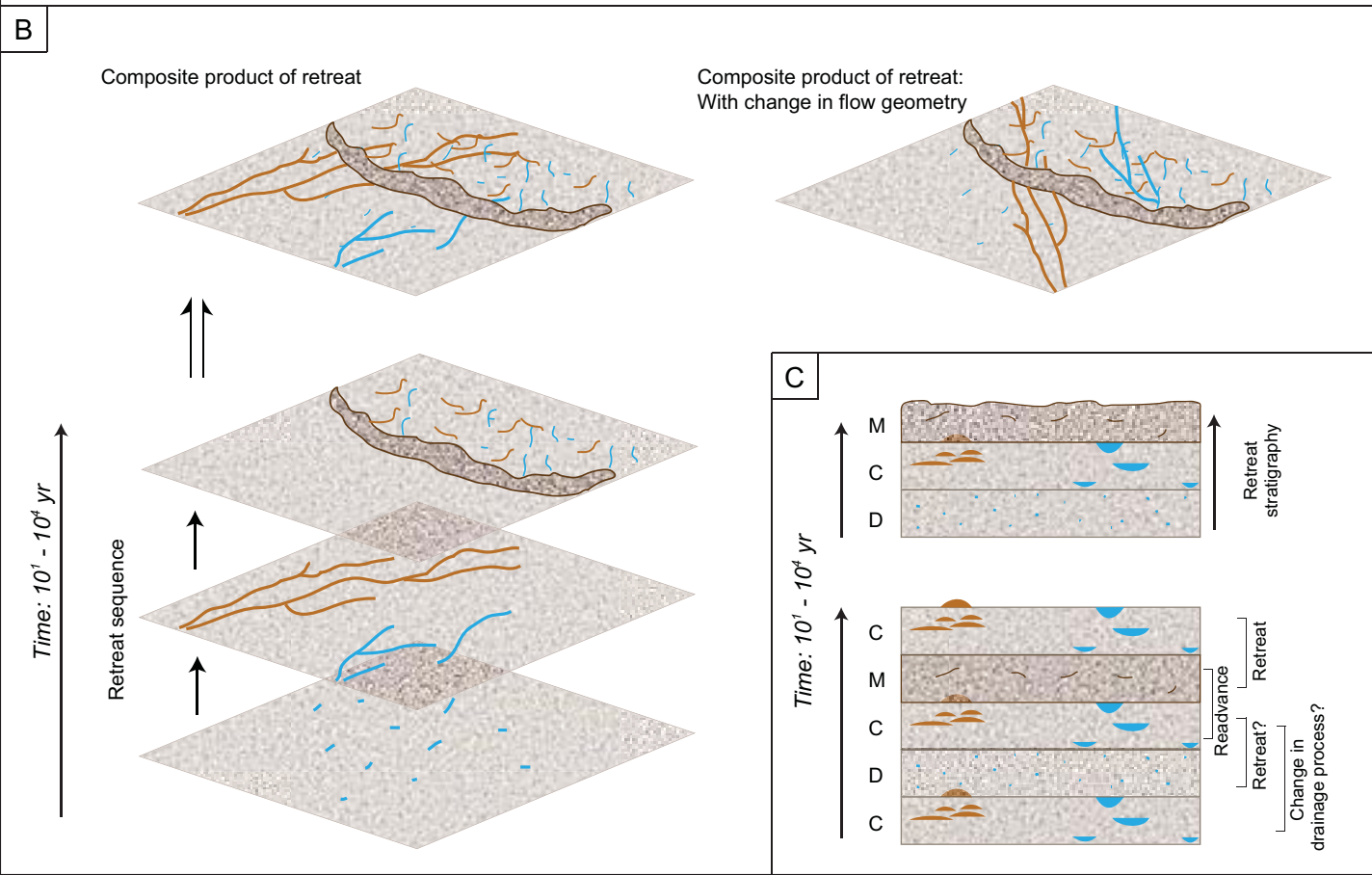
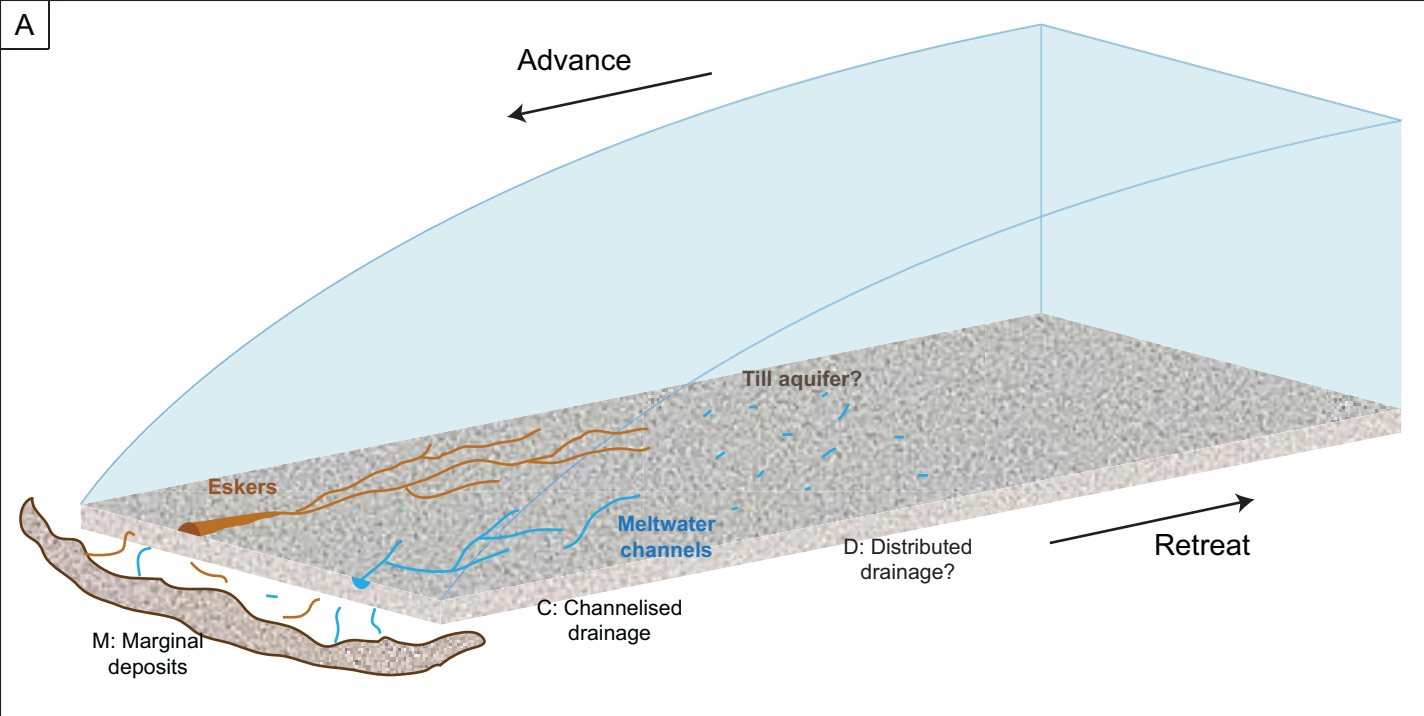
2584 First order hydrological regimes of an ice sheet through space and time. X axis represents ice  
2585 sheet span from centre to margin; Time (t) on vertical axis from maximum ice sheet extent to  
2586 deglaciation. A high pressure environment (H, red dots) is found throughout the ice sheet.  
2587 Röthlisberger-type channelised drainage (C) dissects the outer marginal zone, pockets of  
2588 efficient delivery may be found inward of this, and short-term high magnitude drainage events  
2589 (pale blue lines) are superimposed throughout the ice sheet history. The inward limit of the C  
2590 zone is uncertain: it reflects changing discharge, ice thickness, drainage event frequency  
2591 (amplitude of meltwater input variability). For further explanation refer to the main text.  
2592  
2593

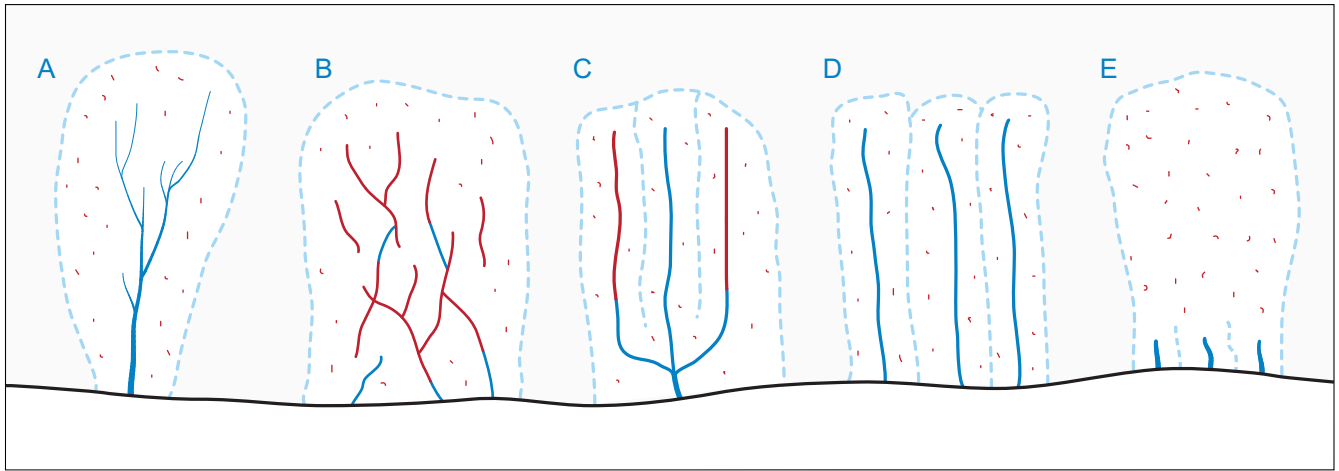


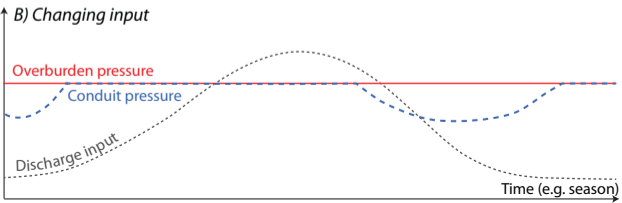
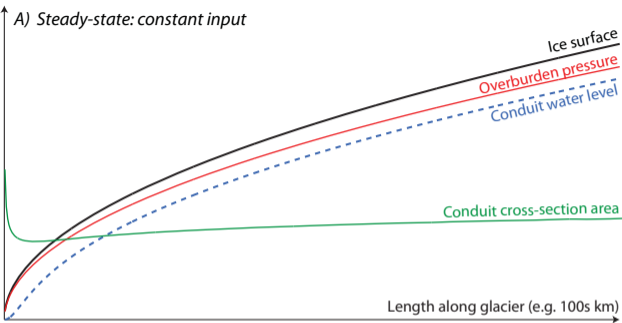




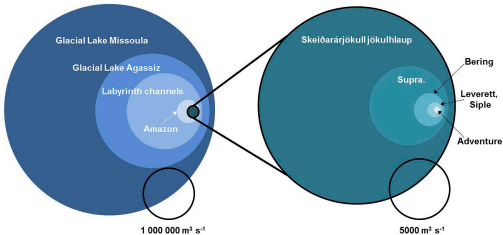












1 000 000 m<sup>3</sup> s<sup>-1</sup>

5000 m<sup>3</sup> s<sup>-1</sup>

Event	Reference	Discharge (m <sup>3</sup> s <sup>-1</sup> )
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

