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Theoretical framework and diagnostic criteria for the identification of palaeo-subglacial lakes

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Abstract

The Antarctic Ice Sheet is underlain by numerous subglacial lakes, which comprise a significant and active component of its hydrological network. These lakes are widespread and occur at a range of scales under a variety of conditions. At present much glaciological research is concerned with the role of modern subglacial lake systems in Antarctica. Another approach to the exploration of subglacial lakes involves identification of the geological record of subglacial lakes that once existed beneath former ice sheets. This is challenging, both conceptually, in identifying whether and where subglacial lakes may have formed, and also distinguishing the signature of former subglacial lakes in the geological record. In this work we provide a synthesis of subglacial lake types that have been identified or may theoretically exist beneath contemporary or palaeo-ice sheets. This includes a discussion of the formative mechanisms that could trigger onset of (or drain) subglacial lakes. These concepts provide a framework for discussing the probability that subglacial lakes exist(ed) beneath other (palaeo-)ice sheets. Indeed we conclude that the former mid-latitude ice sheets are likely to have hosted subglacial lakes, although the spatial distribution, frequency and type of lakes may have differed from today's ice sheets and between palaeo-ice sheets. Given this possibility, we propose diagnostic criteria for identifying palaeo-subglacial lakes in the geological record. These criteria are derived from contemporary observations, hydrological theory and process-analogues and provide an observational template for detailed field investigations.

Key words: palaeo-subglacial lake; subglacial lake; subglacial sedimentology; subglacial geomorphology; subglacial hydrology; ice sheet.

1. Introduction:

The dynamic behaviour of ice sheets is largely governed by basal conditions at the ice-bed interface. This relationship is exemplified by observations made beneath modern ice streams, where water and deformable wet sediments lubricate the bed facilitating rapid ice-flow (Engelhardt & Kamb, 1997; Kamb, 2001). However, the response of ice masses to changes in basal water pressure is non-linear because the hydrological system is dynamic and evolves to accommodate variations in meltwater flux (e.g. Hubbard et al. 1995). Beneath the Antarctic Ice Sheet the hydrological system is thought to comprise a complex network of subglacial lakes, distributed and discrete drainage pathways, water-saturated sediments and sub-ice aquifers (Siegert, 2005; Wingham et al. 2006; Peters et al. 2007; Carter et al. 2007; Fricker et al. 2007; Priscu et al. 2008; Smith et al. 2009; Langley et al. 2011; Uemura et al. 2011). As more than half of the grounded Antarctic Ice Sheet is thought to be warm-based, basal meltwater is thought to be widespread (e.g. Llubes et al. 2006; Pattyn, 2010) and significant heat and mass are redistributed under the ice sheet in the generation, drainage and freeze-on of meltwater (Bell et al. 2011; Tulaczyk & Hossainzadeh, 2011).

Over 380 subglacial lakes have been identified so far beneath the Antarctic Ice Sheet from a variety of geophysical techniques, with 9000-16000 km^3 of water predicted to be stored in these subglacial reservoirs (cf. Wright & Siegert, 2011). The largest is Vostok Subglacial Lake in East Antarctica, which reaches depths of >800 m and is comparable in size to Lake Ontario (Filina et al., 2008). Lakes are thus now recognised as major components of the subglacial environment, interacting with the surrounding hydrological network and capable of transmitting large volumes of water downstream (Gray et al. 2005; Wingham et al. 2006; Fricker et al. 2007; Stearns et al. 2008; Fricker & Scambos, 2009; Smith et al. 2009; Scambos et al. 2011). In at least one case, lake drainage has also been shown to affect ice-flow velocity (Stearns et al. 2008). Furthermore, the abundance of water-saturated sediments and lakes under Antarctica make it a prime habitat for life and this is supported by data indicating the presence of a significant subglacial microbial community (Priscu et al. 2008; Lanoil et al. 2009). For instance, Filina et al. (2008) estimated the sediment layer thickness within Vostok Subglacial Lake to be 300-400 m. These findings suggest that subglacial microbes may play an, as yet, poorly understood role in subglacial and global biogeochemical cycles (e.g. Wadham et al. 2008, 2010; Skidmore et al. 2010).

Subglacial lakes have provoked widespread interest from the media and the scientific community, with much glaciological research focused on investigating their role under the Antarctic Ice Sheet. Direct penetration of a subglacial lake has recently been achieved at Vostok Subglacial Lake by a Russian research team, and measurements have been taken from formerly subglacial lakes recently emerged from the ice margin (e.g. Hodgson et al. 2009a,b). There are also major international research programmes that plan to drill into two other subglacial lake environments (subglacial lakes Whillans and Ellsworth) over the next five years in search of unique forms of life and sedimentary archives of past climate change (e.g. Siegert et al. 2012). The size and scope of these programmes demonstrate both the wide appeal of subglacial lake exploration and the inherent difficulties associated with this endeavour, such as drilling through kilometre-thick ice and avoiding contamination of the subglacial water column (e.g. Siegert et al. 2007, 2012).

Another approach to the exploration of subglacial lakes involves identification of the geological record of subglacial lakes that once existed beneath former ice sheets. Palaeo-investigation of such lakes offers significant advantages because we have comprehensive information about the bed properties (e.g. its geology and topography), they are logistically much more accessible (by boat, car and on foot rather than kilometre long drillholes through ice) and we can examine and sample the sediments with ease. If we can find palaeo-subglacial lakes then we have the potential to make huge leaps in knowledge with regard to the topographic context and hydrological pathways of which the phenomena form a part of; specifically, we have the advantage of spatial and sedimentological information in relation to investigations of contemporary subglacial lakes but the disadvantage of not observing the short-term dynamics. Furthermore, with the drilling programs planning to sample the subglacial lake sediments, the search for palaeo-analogues will benefit both communities in interpreting these sediments (Bentley et al. 2011).

Despite the proliferation of studies on contemporary subglacial lakes, the palaeo-investigation of subglacial lakes is in its infancy. Isolated examples of putative subglacial lakes in the former northern hemisphere ice sheets have been identified (e.g. McCabe & Ó Cofaigh, 1994; Knight, 2002; Munro-Stasiuk, 2003; Christoffersen et al. 2008; Lesemann & Brennand, 2009; Sutinen et al. 2009). However, there is a general apathy towards research into subglacial lakes in the palaeo-community that appears to stem largely from problems identifying their geological signature; particularly in separating proglacial sedimentary environments from subglacial lacustrine ones. Other possible reasons include scepticism as to whether subglacial lakes could have formed beneath the British-Irish, Fennoscandian, Cordilleran and Laurentide ice sheets; and perhaps also caution against being drawn into the largely discredited mega-flood hypothesis for drumlin formation (e.g. Shaw et al. 1989; Evans et al. 2006).

And yet, this sphere of research can make a contribution to the study of subglacial lakes and their importance for ice sheet subglacial processes, and is in need of re-invigoration. Such a contribution involves: (i) the association of lakes with the surrounding glacial geomorphology, which can be used to assess the effect of subglacial lakes on meltwater drainage, ice flow and ice streams; (ii) details about the temporal evolution of subglacial lakes and how they relate to ice-sheet dynamics, sub-Milankovitch scale climatic events and possible palaeo-floods; (iii) subglacial lakes as a possible archive of long-term Quaternary climate change (preservation of subglacial and subaerial lake sediments over successive glacial cycles); and (iv) collaboration with glaciologists and glacial geologists sampling Antarctic subglacial lakes in interpreting palaeo-subglacial and contemporary lake sediments.

The embryonic state of research on palaeo-subglacial lakes necessitates a detailed theoretical assessment, including: (i) whether subglacial lakes could have existed beneath the great Quaternary ice sheets of the northern hemisphere; and (ii) how the likely processes and resulting sediment-landform assemblages might allow subglacial lakes to be identified in the sedimentary record, and how they are distinct from 'ordinary' (i.e. sub-marginal and glacier-fed) lakes that form under a different set of conditions. In this paper we attempt to answer both questions by looking at the physics behind subglacial lake formation and evolution, and by producing diagnostic criteria for identifying palaeo-subglacial lakes.

2. Theory:

2.1 Basal hydrology and thermal regime

The flow and storage of meltwater under ice masses is principally governed by the hydraulic potential (Φ), which is a function of the elevation potential and water pressure (Shreve, 1972):

$$\Phi = \rho_w gh + p_w \qquad [1]$$

where ρ_w is the density of water, g is the acceleration due to gravity, h is the bed elevation and p_w is the water pressure. Differences in hydraulic potential drive water down the hydraulic gradient, and it is assumed that water will 'pond' to form subglacial lakes where there are hydraulic minima (regions of relatively low hydraulic potential) (Clarke, 2005). Understanding the distribution of basal meltwater pressure (p_w) at the ice-bed interface is non-trivial and depends on the configuration of the drainage system, which is also dynamic (Cuffey & Patterson, 2010). However, limited borehole observations show that p_w may be close to the ice overburden pressure (e.g. Kamb et al. 2001) and therefore an assumption can be made that effective pressure is zero and that:

$$p_w = \rho_i gH \qquad [2]$$

where ρ_i is the density of ice and H is the ice thickness. Given this assumption, equation [1] can be re-written as follows:

$$\Phi = \rho_w gh + \rho_i gH \qquad [3]$$

Equation [3] predicts that the ice-surface gradient is ~10 times as important as the bedrock gradient in controlling the hydraulic potential. Indeed, where effective pressure is zero it is possible to predict the steady-state pattern of meltwater drainage and ponding (e.g. Evatt et al. 2006).

The thermal state at the ice-bed interface is a crucial constraint on subglacial hydrological conditions, including meltwater production (or freeze-on), meltwater flux, the development and evolution of the drainage system, and the efficiency of subglacial erosion and sediment transport. The physics governing the thermal regime of ice masses is well understood and is a function of the temperature field (at the ice surface and bed), climatic conditions (e.g. precipitation rate), and ice sheet geometry and velocity (Cuffey & Patterson, 2010). Of particular importance is the interaction between the ice and basal temperature field, with the resulting thermal-mechanical feedbacks crucial in controlling basal velocity (e.g. Payne, 1995). However, robust solutions to the thermal regime are still lacking owing to poor knowledge regarding key parameters such as geothermal heat flux (Pattyn, 2010).

2.2 Subglacial lake types:

Subglacial lakes are "discrete bodies of water that lie at the base of an ice sheet between and ice and substrate" (Siegert, 2000, p.29). Subglacial lakes are distinguished from 'wet spots' in the basal hydrological system as their surfaces are in hydrostatic equilibrium with the overlying ice. Wright & Siegert (2011) presented a revised classification of Antarctic subglacial lake types, summarised in Table 1. Their three categories comprise: (1) those subglacial lakes associated with thick ice near to ice divides in the interior of the continent; (2) those in the onset zones of ice streams near to the ice-sheet margin; and (3) those situated beneath the trunks of outlet glaciers and ice streams (Table 1). The first group is sub-divided according to basal topography into deep subglacial basins where the ice is thickest, those associated with large bedrock depressions, and lakes perched on the flanks of subglacial mountain ranges (Wright & Siegert, 2011). Subglacial lakes that form along ice divides, where the ice surface is flat or nearly flat, have small catchment areas and therefore are likely to be relatively inactive, with low recharge rates (Evatt et al. 2006). Subglacial lakes located in the onset zone of ice streams may modulate these fast-flow arteries, potentially acting both as seeds for enhanced flow (Bell et al. 2007) and triggering short-term flow accelerations (Stearns et al. 2008). The final category comprises very active subglacial lakes under the trunks of ice streams and outlet glaciers that are generally small in size, transient in nature and with sub-annual drainage-recharge cycles (Table 1, Smith et al. 2009). The warm-based corridors generated by rapidly flowing ice streams permit hydrological connections between these lakes and the ice-sheet margin (Dowdeswell & Siegert, 2002). Several 'fuzzy' and 'indistinct' lakes have also been identified in Antarctica, and these are interpreted as areas of saturated sediments (e.g. Carter et al. 2007).

In addition to the lake types described above we also expect subglacial lakes to form in regions where the geothermal heat flux is very high (lake-type 4, Table 1). This mechanism is fully described in section 2.3. Subglacial lakes may also theoretically originate in regions where meltwater drainage becomes blocked or constricted, such as behind a frozen ice tongue or where the ice sheet is rimmed by permafrost (lake-type 5, Table 1). Subglacial lake types (4) and (5) may form as blisters rising above the local topography.

2.3. Conditions and controls governing lake formation

The processes and controls influencing subglacial lake genesis and evolution are still poorly understood. However, given the variety of subglacial lakes and their distribution beneath the Antarctic Ice Sheet we expect formation to occur under a range of conditions.

Perhaps the most intriguing is the 'Captured Ice Shelf' hypothesis whereby subglacial lakes evolve from pre-glacial water bodies occupying glacial overdeepenings or deep (tectonic) basins that become trapped as an ice sheet (re)advances (Erlingsson, 1994a,b, 2006; Pattyn, 2004; Alley et al. 2006; and also see Hoffmann & Piotrowski, 2001; Lesemann & Brennand, 2009). The basic premise is that ice initially must advance across the body of water as a floating ice shelf, which then grounds on the opposite shore. To maintain an ice shelf the upper ice-surface temperatures must be below 0°C otherwise crevasses fill with meltwater causing calving and ultimately disintegration of the shelf (e.g. Scambos et al. 2000). Thus a lake must be deep enough both to allow water to collect beneath the ice and also to prevent the entire column of water freezing. Grounding creates a hydrostatic seal as a result of (a) freezing-on of the thin, grounded portion of the ice shelf and (b) the development of an 'ice rim' (i.e. local ice-air slope reversal) (Fig. 1A). Further advance will lead to thickening of ice above the lake, causing loss of the ice rim and over-pressurisation of the lake water, allowing basal thawing and removal of the hydrostatic seal. This may cause an outburst flood (Fig. 1B) (Erlingsson, 1994a,b; Alley et al. 2006), but the lake need not drain fully if the basin is sufficiently deep and/or the ice surface-slope remains low, perhaps as a result of the interaction of the lake with the ice-sheet (i.e.

producing a slippery-spot) (Fig. 2, Pattyn, 2004) or fast ice flow triggered during the outburst itself (Alley et al. 2006).

The formation and drainage of subglacial lakes *in situ* (as opposed to during ice advance) is a function of the evolving ice mass and the character of the underlying lithosphere (Priscu et al. 2008). The mechanisms controlling subglacial lake genesis, evolution and drainage are discussed below and shown schematically in Figs. 2-4, 6-7.

Firstly, high geothermal fluxes, such as sub-ice volcanism in Iceland (e.g. Björnsson, 2002) may trigger subglacial lake genesis (lake-type 4, Table 1). Subglacial lakes form because high rates of melting at the ice-bed interface draws down the ice surface creating a region of relatively low hydraulic potential (Fig. 2, Björnsson, 2002). The subglacial lake will expand as meltwater flows towards the hydraulic minima created by the ice-surface depression. Periodic drainage events occur as outburst floods when the hydraulic seal is broken by the gradual rise in basal water pressure. As the meltwater drains the basal water pressure will drop and this eventually will re-seal the lake, allowing the process to begin once more (Björnsson, 2002).

Secondly, bed topography influences where subglacial lakes are likely to form, although the importance of the ice surface slope in Equation (3) means not all lakes form in depressions and not all depressions hold lakes (Cuffey & Patterson, 2010). On glacial time-scales, tectonics, erosion (including glacial overdeepening) and sedimentation will all subtly affect the likely distribution of subglacial lakes (cf. Cook and Swift, in press).

An important context for potential lake formation is where focused glacial erosion has produced overdeepenings (Fig. 3), although Cook and Swift (in press) argue that the process of overdeepening itself is unlikely to form a lake. This is because an overdeepening that is approaching the ponding threshold will experience reduced basal water pressure variations and the evacuation of basal sediment by glacial and fluvial processes will diminish, such that further deepening will be unlikely. Further, although the higher basal water pressures associated with overdeepening might promote lake formation by reducing ice-bed friction and hence the ice-surface slope (Fig. 3), any incipient 'lake' is likely to be a focus for sediment deposition that might maintain contact between ice and the bed (a subtle increase in ice-surface slope would then cause overdeepening to be reinvigorated). Perhaps the most important limitation on lake formation in this context, however, is the limit to the depth of overdeepening that is provided by glaciohydraulic supercooling (e.g. Lawson et al. 1998), which in theory will prevent the evacuation of sediment by efficient subglacial channels when the ice-slope to bed-slope ratio approaches ~1:2 (Alley et al. 2003). Beyond this threshold, the limited potential for further erosion should prevent the overdeepening reaching the ponding threshold, and water will continue to exit the overdeepening via less efficient forms of subglacial drainage. In reality, steep hydraulic gradients promoted by seasonally- and diurnally-focussed surface melt production mean that erosion may continue well below the 'traditional' supercooling threshold (Creyts & Clarke, 2010), but cannot occur beyond the ponding threshold (see later discussion). Subglacial lakes formed in this manner are therefore likely to be small, transient features, possibly characterised by repeated cycles of overdeepening, lake formation, sediment accumulation and lake emptying (Fig. 3). Nevertheless, overdeepenings formed under certain conditions may subsequently be occupied by significant lakes when basal conditions or ice-surface slopes become favourable.

As the ice-surface slope evolves, the subglacial hydrological network will adjust to the change in hydraulic potential (see equation 3). For example, subglacial lake formation can be triggered as an ice sheet advances, provided the bed topography is also suitable, because flattening of the ice-surface produces localities where the hydraulic gradient is no longer able to evacuate meltwater to the ice margin (Fig. 4). Consequently, based purely on geometry, more subglacial lakes should form when ice sheets are fully extended and average surface slopes are lower. Conversely, as our theoretical ice sheet shrinks, the ice-surface slopes will become steeper, facilitating lake drainage (Jordan et al. 2010).

The very low basal friction of subglacial lakes means that they act as ‘slippy spots’ at the base of ice sheets. This results in local acceleration of the ice sheet across the lake and therefore flattening of the ice surface (Fig. 5). Significantly, this flattening will act as a positive feedback reinforcing the lake’s stability (Pattyn, 2008). It will also affect local ice-sheet dynamics, with the change in ice slope capable of modifying the ice-flow regime, ice-surface configuration, and catchment area (Fig. 5, Pattyn, 2003, 2004; Pattyn et al. 2004; Thoma et al. 2012).

Fourthly, a change in thermal regime at the ice-bed interface, from freezing to melting, is able to trigger subglacial lake formation (Fig. 6). This can occur by any of the following: (i) thickening of the ice mass; (ii) lowering the accumulation rate (thereby reducing vertical advection of cold ice to the ice bed); (iii) increasing the geothermal heat flux or ice-surface temperature; or (iv) strain heating from enhanced flow (by ice streams or extended flow). In contrast, a switch from warm- to cold-bedded conditions will, in many cases cause subglacial lakes to freeze-on to the overlying ice mass, eventually resulting in ‘fossil’ lakes (Price et al. 2002; Bell et al. 2011; Tulaczyk & Hossainzadeh, 2011).

In some instances, however, a switch of an ice stream from warm- to cold-bedded may, counter-intuitively, result in lake development (Fricker et al. 2007). As ice streams shut down due to basal freezing, sheet flow of water through basal sediments and at the ice/sediment interface becomes channelized into conduits (Bell et al. 2008), cut off from the broader ice stream bed. This channelized drainage may develop into a string of lakes joined by conduits. These unstable and transient lake systems are dependent upon changing ice and water fluxes, changes in the size of the ice stream drainage basin and changes in the thermal regime, thereby impacting upon the stability of the lake and its recharge capacity.

Fifthly, subglacial ponding may also occur in response to an excess of meltwater supply over that which can be evacuated by drainage at and below the bed. A lake may thus form by local increases in meltwater production or decreases in hydraulic conductivity (Fig. 7). Melt rates are dependent on basal thermal regime, with drainage primarily occurring via groundwater flows, meltwater channels and fractures (Röthlisberger, 1972; Nye, 1976; Uemura et al. 2011), with freeze-on also able to redistribute mass (Bell et al. 2011). The hydraulic conductivity can therefore be reduced by generating a seal at the ice-bed interface or by imposing a less efficient form of subglacial drainage. Thermal seals are created where ice freezes to its bed and/or permafrost inhibits drainage through the sub-ice sediments (Fig. 7) (lake-type 5, Table 1). Cold-bedded ice margins may act as natural seals promoting subglacial meltwater storage up ice-flow (e.g. Piotrowski, 1994; Skidmore & Sharp, 1999; Cutler et al. 2002; Copland et al. 2003). Drainage of meltwater stored behind the ice margin may occur periodically as subglacial floods when the water pressure overcomes the resistance from the ice seal, similar to but on a larger-scale than floods observed at polythermal glacier margins (e.g. Skidmore & Sharp, 1999; Copland et al. 2003). Switching of drainage can occur when water in channels ascends a sufficiently steep adverse slope for glaciohydraulic supercooling to occur (e.g. Lawson et al. 1998), or when basal sliding velocities become too great for channelised drainage to be sustained (e.g. Bell et al. 2008). Sub-ice aquifers may also be important in this context, as ponding requires meltwater discharge into hydraulic minima to exceed groundwater outflow.

Finally, the character of the subglacial drainage system is also influenced by the underlying substrate. For example, as meltwater flows across a boundary between exposed crystalline bedrock and till covered sedimentary rock there will be a jump in basal water pressure (Clark & Walder, 1994). This will cause a damming effect resulting from the associated jump in the hydraulic head, whilst the change in basal stresses may also lead to enhanced subglacial erosion, creating depressions for the lakes (Evatt et al. 2006). Perhaps this is why the major lakes in Canada lie along the shield boundary (e.g., Great Bear, Great Slave, Athabasca and the Great Lakes).

3. Could subglacial lakes have formed beneath the Quaternary ice sheets of the northern hemisphere?

Given the physics behind subglacial lake genesis and evolution, we first of all consider whether there are theoretical grounds to expect subglacial lake formation beneath major palaeo-ice sheets. This is certainly a valid question given that subglacial lakes have been found in Antarctica but not in Greenland, despite observations showing the widespread presence of subglacial meltwater (Oswald & Gogineni, 2008). Do some ice sheets and bed topographies lend themselves towards lake formation whilst others do not? The following section looks specifically at the former North American and European ice sheets and briefly explores whether the controlling factors behind subglacial lake formation suggest that lakes should have occurred. We approach this be addressing a series of critical questions.

3.1 Were the ice-surface slopes too steep for subglacial lakes to form?

Ice-surface slope is a first order control on the hydraulic gradient and thus on subglacial lake genesis and stability (see section 2 above & Fig. 4). The vast majority of subglacial lakes in Antarctica are located along ice-divides in the ice-sheet interior and beneath ice streams, where the ice surfaces are relatively flat (Wright & Siegert, 2011). Yet, in contrast to Greenland and Antarctica, where the ice slope can be measured directly, the geometry of palaeo-ice sheets need to be inferred from numerical ice-sheet models constrained by geological data (e.g. Hagdorn, 2003; Peltier, 2004; Tarasov & Peltier, 2004; Hubbard et al. 2009). There is therefore an envelope of uncertainty concomitant with the derived palaeo-slopes. However, ice-divide locations and palaeo-ice streams are well documented in the geological record, and offer the most obvious settings for the former existence of subglacial lakes (e.g. Stokes & Clark, 1999, 2001; Winsborrow et al. 2004; De Angelis & Kleman, 2005, 2007; Ottesen et al. 2008; Ó Cofaigh et al. 2010; Livingstone et al. 2012).

Average ice-surface slopes are, in part, a function of ice sheet size, with smaller ice sheets typically characterised by steeper average slopes. This is useful given that the former margins of the major palaeo-ice sheets of North America and Europe are well constrained (see Fig. 8) (e.g. Boulton et al. 1985, 2001; Dyke & Prest, 1987; Kleman et al. 1997; Dyke et al. 2002; Svendsen et al. 2004; Chiverrell & Thomas, 2010; Clark et al. 2012). Significantly, the area of the Laurentide Ice Sheet during the Last Glacial Maximum (LGM) is comparable to the present day Antarctic Ice Sheet (both ~14 million km^2), whilst the European Ice Sheet was much smaller at the LGM (~6.2 million km^2). This implies that the European Ice Sheet was less able to trap meltwater at its bed.

3.2 Were beds rough enough to collect meltwater?

Bed topography is a key influence on the distribution of subglacial lakes, with meltwater tending to accumulate in basal depressions (Bell et al. 2006; Tabacco et al. 2006; Wright & Siegert, 2011). Northern Europe is defined by large variation in relief, with numerous topographic depressions, including basins, fjords, lochs, and mountain valleys that would make prime candidates for hosting subglacial lakes. In East Antarctica, many of the largest, stable lakes occupy elongate, rectilinear depressions, thereby indicating a tectonic control on their formation (Studinger et al. 2003; Bell et al. 2006; Tabacco et al. 2006). The morphology of these depressions is comparable to the lochs of Scotland, of which many are also tectonic features. Similarly the deep basins in east Antarctica offer useful analogies to Hudson Bay and the Baltic Depression, whilst the Norwegian fjords are comparable to the well-defined topographic troughs along the Antarctic Peninsula. In contrast to Europe, large tracts of North America are relatively flat, and therefore the best candidates for large former subglacial lakes are probably the depressions occupied by present-day lakes (Evatt et al. 2006; Christoffersen et al. 2008), which reach depths of over 600 m (e.g. Great Slave Lake).

In addition to pre-existing depressions in bed topography providing potential homes for subglacial lakes, another possibility is that glacial erosion has an inbuilt proclivity for manufacturing depressions that later become subglacial lakes. Bedrock landscapes dominated by glacial erosion frequently contain depressions cut below the typical valley long profiles that arise from fluvial erosion and producing overdeepenings of the order of 10-10^2 m in depth and 100-10^3 m in length (Linton, 1963). The reasons and mechanisms for producing such overdeepenings remain uncertain (see review; Cook and Swift, in press), but in Fig. 3 we highlight that once developed they are potential sites for lake formation. Depending on factors that drive changes in ice-surface slope, they may become subglacial lakes that slowly fill with sediment and remain as such, or enter a cycle of repeated lake formation, filling and sediment evacuation. Both circumstances are likely to have significant effects on ice dynamics because basal resistance to ice flow will be much less for ice traversing a surface that is water or soft sediments compared to bedrock, even if hydrological limits to overdeepening mean incipient lake and/or sediment layers remain almost insignificantly thin. We thus speculate that glaciers may overdeepen their beds as a means of increasing ice drainage efficiency.

The proclivity for focussed glacial erosion, however, also provides a means of reducing the number of potential homes for subglacial lakes in pre-existing tectonic depressions. This is because hydrological limits to subglacial erosion will focus erosion not into tectonic depressions but onto bed highs, with sedimentation occurring within bed lows. On glacial time scales, this should cause the gradient of bed slopes that are adverse to ice flow to evolve toward the threshold slope for water and sediment transport. As our review of lake formation processes has indicated, depressions characterised by such slopes clearly do not provide ideal situations for the formation of large or stable subglacial lakes. The importance of this effect will depend on the overall erosional potential of ice, being greatest for more dynamic ice masses where abundant meltwater reaches the bed, and where ice is thickest and fastest flowing, which is likely to occur in topographic lows where tectonic depressions are most likely. This factor might explain the dearth of evidence for lakes beneath the Greenland ice sheet.

3.3 How widespread was water production beneath the Quaternary ice sheets?

It is clear from the glacial bedform record that large portions of the mid-latitude palaeo-ice sheets were warm-based at some point during the last glacial stadial (Dyke & Prest, 1987; Boulton & Clark, 1990a,b; Kleman et al. 1997; Boulton et al. 2001; Greenwood & Clark, 2008; Hughes et al. 2010). However, in stark contrast to the interior of the Antarctic Ice Sheet, where ~55% of its bed is estimated to be at the pressure melting point (e.g. Pattyn, 2010), the central axes of the Laurentide, British-Irish and Fennoscandian ice sheets were thought to have been frozen to their beds during the LGM (Fig. 8, cf. Kleman et al. 1997; Kleman & Hättestrand, 1999; Kleman & Glasser, 2007). This is supported by ice altitude estimates for the former British-Irish (ice cover in Scotland 900 m to 1800 m a.s.l.) and Fennoscandian ice sheets (2100 to 3400 m a.s.l.), which indicates significantly thinner ice than in the interior of Antarctica (Boulton et al., 1977, 1985, 1991; Denton & Hughes, 1981; Lambeck, 1993, 1995; Boulton & Payne, 1994; Lambeck et al. 1998; Boulton et al. 2001; Ballantyne, 2010). Additionally, the core regions of Scandinavia and Britain are characterised by upland massifs, which contrasts with the deep basins (and hence thick ice) found in the interior of Antarctica. It is therefore not surprising that these central upland areas were protected by cold-based ice, and we can therefore speculate that these central zones would not have hosted the large subglacial lake systems that we observe under the Antarctic Ice Sheet. However, there is some theoretical and geological evidence to countenance warm-bedded conditions beneath the thick (>4 km) core of the Laurentide Ice Sheet in Hudson Bay (Sugden, 1977; Josenhans & Zevenhuizen, 1990). Model results suggest that 60-80% of the Laurentide Ice Sheet may have been frozen to its bed during the ice sheets' growth phase and at the LGM, with the melt fraction increasing significantly during deglaciation (up to 80% warm-based, Marshall & Clark, 2002).

The distribution of subglacial lakes in Antarctica testifies to strong variations in the basal thermal regime at the sub-ice sheet scale, with subglacial lakes formed by strain heating beneath the trunks of ice streams found in close proximity to the ice margin and under very thin ice (<800 m thick) (Fricker & Scambos, 2009; Smith et al. 2009; Scambos et al. 2011). Furthermore, thermal trimlines supported by surface exposure ages indicate that the generally cold-bedded upland massifs in the core regions of Britain, North America and Scandinavia would have been contemporaneous with warm-basal conditions in fjords and valleys (e.g. Staiger et al. 2005; Briner et al. 2006; Phillips et al. 2006; Ballantyne, 2010); it is these deep topographic depressions that were most likely to have stored meltwater. These observations are comparable to the Gamburtsev subglacial mountain range in Antarctica where wet basal conditions are pervasive throughout the valleys yet the peaks remain cold-based, and water that is forced up the steep valley sides is frozen onto the base of the ice sheet (Bell et al. 2006, 2011). Finally, the use of ribbed moraine to demarcate frozen-bed conditions is open to debate given alternative theories for their genesis, such as the bed instability mechanism, which invokes naturally arising instabilities in the coupled flow of ice and till (e.g. Dunlop et al. 2008). Hence there may be some overestimation of the extent of cold-beds.

One supplementary mechanism which has not been considered yet around the largely marine-based margins of the Antarctic Ice Sheet is the influence of permafrost. Certainly, permafrost is recognised to have been significant along the terrestrial fringes of the European and North American palaeo-ice sheets (e.g. Szewczyk & Nawrocki, 2011); and may have reduced the hydraulic conductivity in these marginal zones, thereby constricting drainage and promoting ponding up ice-flow (lake-type 5, Table 1) (see sections 2.2 & 2.3).

3.4 Were the palaeo-ice sheets stable for long enough for subglacial lakes to form?

In contrast to the East Antarctic Ice Sheet, which is thought to be ~34 million years old and has probably been stable for the past 14 million years (Sugden et al. 1993; Sugden, 1996; De Conto & Pollard, 2003), the mid-latitude palaeo-ice sheets are relatively transient features that evolve and disappear on glacial timescales. Geological and modelling evidence indicates that these palaeo-ice sheets were also dynamic systems characterised by significant re-organisations, fluctuations and abrupt shifts in ice-flow behaviour and direction (e.g. Boulton & Clark, 1990a,b; Kleman et al. 1997; Kleman & Glasser, 2007; Greenwood & Clark, 2008, 2009a,b; Hughes, 2008; Livingstone et al. 2008; Evans et al. 2009; Hubbard et al. 2009; Clark et al. 2012). Given these differences it is pertinent to consider whether there was sufficient time for subglacial lakes to develop beneath these dynamic ice masses. Note that subglacial lakes formed by the capture of pre-existing proglacial lakes (captured ice-shelf hypothesis) do not need to be filled and therefore, if lakes did form by this mechanism, then the question is of timescales is less relevant.

Observation of the voluminous subglacial lakes that congregate close to the ice divides in the interior of the East Antarctic Ice Sheet are characteristically stable, with low refill rates and residence times (i.e., the average time that water spends in the lake) on the order of 4500-125000 years (e.g. Bell et al., 2002; Thoma et al., 2008). This is not to say that subglacial lakes would not have formed in similar conditions (i.e. no/limited basal sliding and small catchment areas) beneath the mid-latitude palaeo-ice sheets, rather that the development of very large lake systems were less likely. However, more-active subglacial-lake systems beneath Antarctica (lake types 2 & 3, Table 1) offer possible analogues for the type of lake system we might expect beneath the former mid-latitude ice sheets. Monitoring of these active lakes reveals water systems evolving on sub-annual timescales, with water exchange occurring both between lakes and into the surrounding basal environment (Gray et al. 2005; Wingham et al. 2006; Fricker et al. 2007; Fricker & Scambos, 2009; Smith et al. 2009).

Meltwater was clearly abundant beneath the North American and European ice sheets, as evinced by intricate and extensive networks of subglacial meltwater channels, tunnel valleys and eskers (e.g. Brennand & Shaw, 1994; Clark & Walder, 1994; Kristensen et al. 2007; Hughes, 2008; Boulton et al. 2009; Clark et al. 2012; Kehew et al. 2012). Moreover, the large-scale distribution of some meltwater networks implies that stable hydrological regimes could operate at the base of these former ice sheets, in suture zones or controlled by the transmissivity of the bed (e.g. Boulton et al. 2009). The implication is groundwater flows alone are not able to account for the meltwater flux (see also discussion in Piotrowski, 2006), with the excess required to drain (or pond) along the ice-bed interface.

3.5 Could the generation and penetration of supraglacial meltwater to the bed have influenced subglacial lake evolution?

In contrast to Antarctica, surface melting was significant near the margins of the mid-latitude palaeo-ice sheets, especially during deglaciation when meltwater production would have been considerable. The influence of surface meltwater on basal processes is demonstrated by the seasonal acceleration of the Greenland Ice Sheet, which is attributed to the drainage of surface meltwater to the bed (Zwally et al. 2002; Joughin et al. 2008; Shepherd et al. 2009; Bartholomew et al. 2010). This process would have helped raise subglacial erosion rates of palaeo-ice sheets and may therefore have reduced the number of sites where lakes could have potentially formed (see section 3.2). Seasonally- and diurnally-high meltwater inputs (cf. Bartholomaus et al. 2008; Schoof, 2010) will produce extreme hydraulic gradients that allow hydraulically-efficient subglacial channel systems to remain stable at ice-bed slope ratios that are significantly in excess of the threshold ratio for glaciohydraulic supercooling (Creyts and Clarke, 2010). The implication is that the high meltwater inputs may 'flush' subglacial lakes resulting in diurnal/seasonal lake drainage, with water only stored in such depressions when surface melting subsides.

3.6 Synopsis

It is clear from the analysis presented above that there are compelling physical reasons to presume that the mid-latitude palaeo-ice sheets had subglacial lakes. For instance, subglacial lakes formed by strain heating in zones of enhanced flow are likely to have been commonplace beneath these palaeo-ice sheets (lake-type 3, Table 1), whilst subglacial lake blisters may also have formed behind frozen margins (lake-type 5, Table 1). However, in contrast to Antarctica large, stable lakes in the centre of the mid-latitude palaeo-ice sheets (lake-type 1a, Table 1) are less likely given the propensity for frozen-bed conditions, relatively dynamic ice sheets, and the low refill rates close to ice-divides. Despite this, lakes are still likely to form within valley systems where warm bedded conditions prevail, whilst the interior of the Laurentide Ice Sheet in Hudson Bay was potentially warm-based due to the large (>4 km) thickness of ice (e.g. Sugden, 1977; Josenhans & Zevenhuizen, 1990; Ross et al. 2011).

4. Diagnostic criteria for the identification of palaeo-subglacial lakes

Little work has been done toward developing a theoretical basis for predicting sediment and landform assemblages we expect a subglacial lake to produce (although see Bentley et al. 2011). Consequently we have no clear diagnostic criteria for identifying palaeo-subglacial lakes in the geological record. Any attempts to identify them in the geological record suffer from problems of circularity when the sedimentological criteria used to identify palaeo-subglacial lakes are derived from sediments *assumed* to be palaeo-subglacial lakes. However, because of the physical setting of subglacial lakes, it seems reasonable to suggest that they will leave behind a distinct imprint of their activity. The challenge is to distinguish palaeo-subglacial lakes from subaerial lake sediments that may exhibit subtly different signatures. The following section presents diagnostic criteria for identifying palaeo-subglacial lakes in the geological record. This is based on our knowledge of subglacial lake environments; conceptual models of sedimentation and sedimentary processes derived for contemporary subglacial lakes (Fig. 9, cf. Bentley et al. 2011); and analogous depositional processes and environments (e.g. sub-ice shelf melt-out; grounding-line fans/moraines, Fig. 10). These criteria are illustrated in Table 2 and discussed below.

4.1 Sediment facies

4.1.1. *Melt-out of debris-rich basal ice*

The rain-out of sediment from the lake ceiling, released by the melting of basal-rich ice, is comparable to ice-rafted debris (IRD) sourced from ice-shelves and icebergs (Fig. 10) (e.g. Gilbert, 1990; Anderson et al. 1991; Powell & Domack, 1995; Powell et al. 1996; Domack et al. 1998, 1999; Evans & Pudsey, 2002; Kilfeather et al. 2011). Sustained rain-out from the ice-ceiling and gravitational settling will produce drapes of massive- to weakly-stratified dropstone diamicton and dropstone mud with gradational boundaries (Figs. 10 & 11) (e.g. Evans & Pudsey, 2002). Winnowing of fine-grained sediments settling through the water column is also expected to occur given the presence of particles (<20 microns) in accreted ice above Vostok Subglacial Lake (Royston-Bishop et al. 2005). Dropstones falling through the water column will either produce a weak cluster fabric with a vertical preferred orientation, or a girdle fabric from clasts falling sideways when they hit the sediment (Benn & Evans, 2010). The rate and spatial extent of deposition is a function of the debris content of the basal ice, the area of the ice ceiling that is melting and the basal ice velocity.

4.1.2 *Subaqueous debris-flow deposits*

Proximal to the up-ice-flow margin of the lake sedimentation is likely to be characterised by subaqueous debris-flows sourced from till either squeezed out at the subglacial 'grounding-line' or as the output from subglacial deforming layers, and from portal-derived sediment-laden meltwater (Fig. 10). Because the saline content of subglacial lakes is thought to be low the density contrast between freshwater and the denser sediment-laden meltwater means that underflows will predominate. This is analogous to depositional processes in glaciolacustrine settings (Ashley et al. 1985; Bentley et al. 2011). Deposition by suspension settling and bedload reworking will produce proximal-to-distal fining successions and a characteristic graded, vertical ('Bouma') sequence (Fig. 11) (Bouma, 1962), affected both by the Coriolis effect and the lake floor topography (Benn & Evans, 2010). The full Bouma sequence (turbidite), from bottom to top, comprises: massive or normally graded sand or gravel, often with a basal unconformity; planar-laminated sand; ripple cross-laminated sand and silt; interlaminated silt and/or clay; with a clay or silt drape (Fig. 11) (Benn & Evans, 2010). The sediments deposited when water is flushed into a subglacial lake may be similar to the rhythmic sequences of coarse-grained turbidites deposited in proglacial fluvial systems during outburst floods (O'Connor, 1993; Smith, 1993; Russell & Knudsen, 1999). Sediments distal to the up-ice flow margin will trend into normally-graded sand and silt couplets deposited by gravitational settling of the far-travelled, fine-grained component (Fig. 11) (Bentley et al. 2011). Cohesionless debris-flows may originate on subaqueous slopes or by gravitational transport delivered from subglacial stream flow (Benn & Evans, 2010). These typically form dipping sheets or lobate masses characterised by erosive (scoured) bases, and may exhibit inverse grading due to dispersive pressures (Postma, 1986; Nemec et al., 1999) (Fig. 11). Cohesive debris-flows from the remobilization of unstable grounding-line sediments and input of the mobile till layer will occur proximal to the up-ice flow margin, forming sheet-like or lobate beds. Typical facies include massive, matrix-supported diamicton or muddy sands and gravels, which may pass vertically into normally-graded and stratified gravels, sands and silts (Figs. 10 & 11). Structures produced by shearing at the base of debris-flows include attenuated strings of underlying sediment, shear planes and normal or inverse grading (Benn & Evans, 2010). If the debris-flows are sourced from the deforming till layer the lake sediments will also grade into it.

Because subglacial lakes are hydraulic sinks and therefore water and sediment may theoretically enter from anywhere the palaeocurrent directions, as measured from the associated sediment facies, may display multiple orientations. This is in contrast to proglacial lake palaeocurrent orientations, which are likely to record sediment input from one overall direction.

4.2 <u>Conceptual model of subglacial lake sediments and landforms</u>

4.2.1 *Sediment architecture and facies associations*

The geometry of subglacial lakes is strongly influenced by the hydraulic gradient (Equation 3). Thus, lake sediments may occur in 'odd' settings, where topography alone is not able to constrain the lake's formation (Table 2). Furthermore, the sloping ice-lake interface means that subglacial lake sediments may occur over a large vertical extent, giving a lop-sided geometry where the influx points, associated landforms and proximal sediments are lower than the more distal sediments (Table 2). This offers some of the most conclusive evidence for a subglacial rather than proglacial lake genesis, although isostatic rebound may tend to reduce this difference as rebound will be greater up ice where the ice was thicker. Sedimentation rates in subglacial lakes are likely to vary considerably, much like pro-glacial lake environments. Those lakes located along ice-divides in the centre of the ice sheet (lake type 1a, Table 1) are likely to have very low (almost negligible) sedimentation rates, possibly dominated by the small-input of dust melting out from the lake ceiling. In contrast, those lakes located beneath ice streams (lake-type 3, Table 1) may exhibit very high sediment fluxes more akin to estimated rates at ice stream grounding-lines (see Livingstone et al. 2012 for a review).

In contrast to proglacial lakes (see Smith & Ashley, 1985), one would not expect pronounced seasonal lake-surface warming and the formation of the epilimnion and hypolimnion. Hence, there would be no seasonal overturning effect as the water-surface cools during the autumn. Thus at any one time there will be less suspended sediment in the subglacial water column (due to the lack of overturned suspended sediment from the deeper parts of the lake) and therefore sedimentation of the fine-grained fraction may be higher than in proglacial lake environments (Table 2). However, the presence of sustained lake circulation may keep the finest material in suspension (Bentley et al. 2011)

Facies successions are likely to be characterised by stacked (stratified) sequences of interbedded and interfingering glacigenic sediments owing to the complex interactions between melt-out and subaqueous debris-flow processes (Table 2 & Fig. 14). For example, the periodic drainage of sediment-laden meltwater into subglacial lakes (cf. Smith et al. 2009) will result in cyclic successions of turbidites separated by melt-out facies and waterlain diamicton (Fig. 11). Conversely, flushing events through efflux points will erode sediments at the down-ice flow margin (Bentley et al. 2011). Full drainage of the lake, either on a permanent or cyclical basis, will cause the ice to re-connect with the bed, leading to erosion and deformation of the underlying lake sediments and deposition of a capping subglacial traction till (Table 2). Where the ice re-grounds at the down-ice flow margin the lake sediments may become truncated and strongly deformed in the direction of ice flow. Significantly, where a glacier overrides an infilled proglacial lake the deformation signature is likely to be pervasive across the sediments, whereas subglacial lake sediments may only display a strong deformation and erosion signature at the down-ice flow end (Table 2).

As an ice sheet is lowered onto subglacial lake sediments the confining effect of the ice-lid is expected to inhibit porewater escape. We therefore predict that subglacial lake sediments will not be overconsolidated (unless compacted during subsequent glacial cycles or at a late-stage of ice retreat when water has drained away) due to the maintenance of high porewater pressures. Conversely, if a glacier overrides its own proglacial lake deposits porewater can be more easily squeezed out in front of the advancing ice margin and the deposits are more likely to consolidate. Thus, the consolidation state may be a useful proxy for identifying palaeo-subglacial lake sediments.

The likelihood that subglacial lake deposits could survive subglacial erosion and be preserved after deglaciation is an open question. It is especially relevant given that subglacial lakes tend to form under conditions that are favourable for widespread subglacial erosion and/or deformation. Examples of a few small lakes emerging from under the ice in Antarctica (Hodgson et al. 2009a,b) demonstrates that lakes may be maintained as sediment sinks through the transition to ice-free conditions. Tectonic and overdeepened basins may be particularly favourable for the preservation of sedimentary sequences, as they would act as foci for sediment deposition. Conversely subglacial lake sediments deposited beneath ice streams are more prone to erosion or deformation given the large sediment fluxes and erosion rates observed (e.g. Smith et al. 2012), and also because the lakes themselves tend to be smaller in sizes and more active and ephemeral (e.g. Smith et al. 2009). This is likely to be especially prevalent towards the onset zone of ice streams where erosion is dominant, rather than towards the grounding-line, where lake sediments may instead quickly become buried by the advection of glacigenic material delivered from upstream.

4.2.2 *Landform-sediment associations*

A number of landform-sediment associations typically found in glaciomarine or glaciolacustrine environments, including grounding-line fans, morainal banks (Fig. 10) and grounding-zone wedges, may also form in subglacial lakes; depending on the sediment flux and stability of both the lake level and influx points (Table 2) (Bentley et al. 2011). Significantly though, the lake-ice interface will prevent vertical aggradation of subaerial landforms such as deltaic topsets, with deposition instead forced to prograde or fan laterally (McCabe & Ó Cofaigh, 1994; Bentley et al. 2011). However, deltaic bottomsets and foresets are deposited subaqueously and therefore may readily be found in subglacial lakes.

Grounding-zone wedges are formed by the subglacial transport and deposition of deformation till at the grounding-line in regions of fast ice-flow (e.g. Alley et al. 1989; Larter & Vanneste, 1995; Shipp et al. 1999; Anandakrishnan et al. 2007). It is therefore plausible that grounding-zone wedges also form beneath ice streams where abundant volumes of deforming till are deposited into subglacial lakes via gravity flow processes (Bentley et al. 2011).

Grounding-line fans, dominated by subaqueous mass flow deposits should build up at efflux points where subglacial meltwater is discharged into subglacial lakes (Fig. 10) (e.g. Cheel & Rust, 1982; McCabe & Ó Cofaigh, 1994; Bennett et al. 2002). Persistent underflows may lead to the down-fan development of a channels and levee system, while mass flow diamictons and slump facies are also commonly observed in analogous glaciolacustrine settings (Fig. 10) (Cheel & Rust, 1982). Moreover, the connection with the subglacial hydrological system means grounding-line fans probably form at the down-flow end of esker networks (e.g. Banerjee & McDonald, 1975; Gorrell & Shaw, 1991). Indeed, the esker beads identified by Gorrell & Shaw (1991) are interpreted to have formed in cavities beneath the ice and are therefore effectively subglacial fans.

Morainal banks are elongate masses of sediment deposited at the grounding-line (Benn & Evans, 2010). In a subglacial setting their formation may occur through the melt-out of sediment close to the lake margin, output of subglacial deforming till and sediment laden meltwater discharged through conduits or by sheet flow (Fig. 10) (Powell & Domack, 1995; Powell, 2003).

The deposition of landforms at the subglacial lake grounding-line will define the lake “shoreline”. The largest sediment-landform assemblages are likely to form at the up-ice flow margin, particular in the case of grounding-zone wedges which are flow dependent (Bentley et al., 2011). However, given that subglacial lakes are hydraulic sinks at the ice-bed interface we would expect sediment delivery from a radial network of conduits surrounding the lake (which may be reflected in the meltwater record; see section 4.2.3). This may be represented in the sedimentary record as an array of foreset units found around the “shoreline” with dip direction towards the centre of the lake, or by concentric rings of morainal banks and/or grounding-line fans (Table 2). These pseudo-shorelines may therefore pick out the sloping ice-lake interface, whilst multiple shoreline units may reflect fluctuating lake levels (Table 2). Sediment delivery into proglacial lakes conversely is normally from one (ice-contact) margin only.

4.2.3 *Landform-sediment associations associated with subglacial lake drainage/recharge*

The realisation that subglacial lakes can interact with the surrounding hydrological system, including other subglacial lakes (e.g. Wingham et al. 2006; Smith et al. 2009), implies that the arrangement of eskers, tunnel valleys and meltwater channels will tell us something about the whereabouts of palaeo-subglacial lakes, and also the behaviour of the subglacial hydrological system itself. Thus, we might expect palaeo-subglacial lakes along ice stream corridors (or perhaps suture zones) to be connected at both ends via a network of subglacial meltwater channels/tunnel valleys or eskers, which may stretch to the ice margin (Table 2), with little evidence of such phenomena in the location of the lake itself. The identification of palaeo-subglacial lakes might therefore help explain the often bewildering array of meltwater features of the former Quaternary ice sheets (e.g. Clark et al. 2012).

Subglacial lake drainage events that reach the grounding-line will be debouching large volumes of meltwater and sediment in front of the ice mass and this is likely to leave a geological fingerprint. Moreover, as large portions of the Quaternary ice sheets were marine terminating, evidence of large lake drainage events to the grounding line are expected to be predominantly found in the marine record. Certainly rapidly deposited glacimarine sediments associated with a short period(s) of very high sedimentation rates (e.g. Lekens et al. 2005) may be an expression of large drainage events, possibly from upstream subglacial lakes.

4.2.4 *Subglacial lake sediment-landform transitions*

Subglacial lake genesis and termination should be defined by distinctive sediment-landform transitions, which allow us to distinguish their signature from proglacial lake sediments and landforms (Fig. 12). Lakes formed subglacially are likely to be bounded by subglacial traction tills (Fig. 12a), and in contrast to proglacial lakes (Fig. 12 e-f), there will be no gradual transition from subglacial till into proglacial outwash and debris flows and then into proglacial lake sediments caused by the advance and subsequent damming of the lake by ice. Instead, the initiation of a subglacial lake will trigger a switch from subglacial till deposition/erosion to subglacial lake deposition (Fig. 12a). If the lake periodically drains, subglacial lake sediments are likely to be interbedded with subglacial traction till facies (Fig. 12b), assuming that the lake sediments are not completely eroded during ice-bed recoupling. The contact between the lacustrine deposits and the till above (if there is any) may be either erosional or gradational. Erosional contacts can occur in both subglacial and proglacial lake environments. However, a graded transition is good evidence for subglacial sedimentation, with the upper till deposited as a subaqueous drop-till. Theoretically subaqueous drop-till can also be generated under an ice shelf in a proglacial lake, but this will only occur when the water column is sufficiently deep for a floating margin.

The Captured Ice Shelf hypothesis (e.g. Fig. 1) lends itself to the long term preservation of sediments across multiple glacial cycles and these stable glacial-interglacial aquatic environments are therefore potential archives of long term Quaternary climate change. This will manifest as either transitions from marine sediments overlain by glaciomarine and then sub-ice shelf sediments and finally subglacial lake sediments at sites below sea level (Fig. 12c) or from preglacial lake sediments overlain by proglacial lake and then sub-ice shelf sediments and finally subglacial lake sediments (Fig. 12d) (all assuming that the ice sheet did not ground).

4.3 <u>Biogeochemical tracers</u>

Geochemically speaking, the upper facies of subglacial lake sediments are likely to consist largely of comminuted bedrock, with minor impurities derived from aerosol from meteoric ice melt. At greater depth there might lie a range of sediments from both ice-free conditions and previous glaciations, the upper units of which will almost certainly have been eroded by ice as the current ice column accumulated. The proportion of sediment due to basal erosion processes is likely to be very high in all environments except those close to the ice divide, where basal velocities are low. Even here, there will be some upstream production of comminuted bedrock. Therefore, whilst aerosol inputs might have specific geochemical characteristics, they will be hard to discern against the prevailing glacial matrix.

Glacial sediment was used by earlier geochemists as a proxy for the composition of the average crust (e.g. Keller & Reesman, 1963). It was believed that pervasive erosion of the bed and mixing of sediment *en route* to the proglacial zone integrated rock compositions across a large area. It therefore follows from the above that the upper sediment in subglacial lakes are likely to be representative of the bedrock composition in the upstream hydrological and glaciological catchments. As a consequence, geochemical information on the eroded bedrock as an integrated whole may be anticipated, in which there is little bias resulting from the mobilisation of selective grain types and sizes. One geochemical consequence that may be of use in palaeoenvironmental reconstruction is that the upper sediments, since they are unlikely to be flushed and leached of reactive elements, are likely to be relatively high in Ca, Mg, Na and K, species which decline in relative proportion in soils during progressive weathering (e.g. Taylor & Blum, 1995).

Glacial sediment is characteristically strained and often has adhering microparticles (Tranter, 1982). The strained surfaces and microparticles have excess Gibbs Free Energies relative to the grain interiors, and therefore can be expected to be relatively reactive over periods of years. The grain surfaces and microparticles will dissolve, and aluminosilicate dissolution will result in secondary weathering products such as the simple clays, as has been shown in feldspar dissolution experiments conducted in the 1970’s and 80's (see Petrovic et al., 1976). A diagnostic Ge:Si ratio might be imparted onto these clays, since the clays form in an environment which is not flushed, and so capture the lower Ge:Si ratio of the parent rock. This contrasts with other flushed soil environments where Si is mobilised in preference to Ge, and so the Ge:Si ratio of the clays is higher (Chillruud et al., 1994; Jones et al., 2002).

The organic matter found in subglacial lakes is likely to be dominated by that derived from rock in the hydrological and glaciological catchments, and so is likely to have biosignatures that are characteristically ancient, and lack the higher plant biomarkers of more modern soil and vegetation. Microbial processes are likely to operate in the lakes, particularly at the sediment surface, and they might utilise reactive elements of the rock organic matter and bioavailable components melting out of the meteoric ice. Hence, microbial biomarkers may be present and diagenetic reactions will occur in the sediment.

There is little difference between subglacial lacustrine sediment and deep sea lithogenic sediment as a microbial habitat, except in terms of salinity and the source of organic matter. Both are relatively cold, near zero degrees, dark and at high pressure, equivalent to kilometres of water. They are also likely to be poor in organic matter, of the order of 0.4%, the crustal average rock composition, and the organic matter is likely to be recalcitrant overall. Some sub-ice sheet waters are likely to be much more dilute than sea water, for example those in Vostok Subglacial Lake (Siegert et al., 2003), but some are as concentrated as seawater, for example those beneath the Kamb Ice Stream (Skidmore et al., 2010). However, just as lithogenic deep sea sediments are host to microbes catalysing diagenesis, so too are subglacial lacustrine sediments. Microbes utilise compounds such as organic matter and sulphide minerals to obtain energy for metabolic processes. Comminution of bedrock liberates primary sulphides in the silicate mineral mass, exposing them to microbial processes. Sulphides can be microbially oxidised by a number of oxidising agents, including O_2, SO_4^{2-}, NO_3^-, Fe(III) and Mn(IV) (see Hodson et al., 2008; Tranter et al., 2005). Hence, microbial processes in subglacial lacustrine environments are likely to preferentially weather comminuted primary sulphides, so depleting them. Deeper in the sediment, where anoxic conditions are likely to occur, secondary sulphides may form, and so a characteristic of subglacial lacustrine sediments might be the lack of primary sulphides and the presence of secondary sulphides formed following sulphate reduction (Wadham et al. 2004).

One final effect which might occur in the subglacial lacustrine sediment is that freezing of pore waters might occur during the termination of glacial periods, when the ice thickness decreases and the pressure melting point increases. Then, secondary sulphate minerals such as gypsum and mirabilite might form. The preservation potential of sulphates is very low, because of their solubility, but their presence, in combination with the other features described above might be helpful in paleoenvironmental reconstructions.

Although a large proportion of the bacteria found in subglacial lakes are likely to be psychrophilic, and thus adapted to a low temperature habitat, it is unlikely that they will differ from those found in other sedimentary environments near glaciers, such as proglacial lakes and till (Hodson et al., 2008). Therefore, the concept of using endemic microorganisms as diagnostic criteria for identifying palaeo-subglacial lake sediments is unrealistic. However, a distinct feature will be the lack of chlorophyll and pigments associated with eukaryotic and bacterial photosynthesis. Therefore, it might be possible to distinguish benthic sediments that have been deposited during the transition of a subglacial lake into a proglacial lake through the appearance of such pigments at times coincident with the deglaciation of the site. However, the remains of a number of other organisms associated with aquatic environments (e.g. diatoms and chironomids) and the surrounding catchment (e.g. pollen, spores and plant macrofossils) could not survive subglacially, but might be present as reworked material, or material in transit through the glacier/ice sheet after burial. Therefore, the biology of a palaeo-subglacial lake is likely to be dominated by bacteria with a narrow range of characteristics necessary to survive the extreme conditions.

4.4 Dating palaeo-subglacial lake sediments

Dating of the lake sediments may provide further evidence for the existence of palaeo-subglacial lakes (Table 2). However, many conventional methods are not applicable in subglacial environments, which are notoriously difficult to date. For instance, both radiocarbon and Optically Stimulated Luminescence (OSL) dating will be compromised by the unknown provenance and transport history of the material being dated (Bentley et al. 2011). However, negative evidence such as anomalous dates or the absence of discrete tephra layers might themselves provide a clue that the lake sediment originated subglacially. In the case of OSL dating, it has been suggested that subglacial grinding and crushing can reset the sediment (Swift et al. 2011; Bateman et al. 2012) and that subglacially-reset material may exhibit diagnostic luminescence signatures as a result of the specific bleaching mechanism (Swift et al., 2011). Thus, providing material can be proven to have been mechanically-reset, anomalous OSL dates in well-preserved lake sediments may in fact record the time since the sediment was deposited in a palaeo-subglacial setting, and could therefore potentially be used to constrain lake age.

The magnetic properties of sediments are independent of the environment they were deposited and may therefore be used to establish a chronology for subglacial lake sediments (cf. Bentley et al. 2011; Table 2). The orientation of fine-grained magnetic minerals that dropped through the water column can be measured in the sedimentary sequence and compared with dated reference curves, which include magnetic reversals (e.g. Brunhes-Matayuma ~780ka) and magnetic excursions (e.g. Hodgson et al. 2009a). In the case of the last glaciation four magnetic excursions - Blake Event (~122 ka BP), post-Blake Event (~95 ka BP), Norwegian-Greenland Event (~75 ka BP) and Laschamp Event (~40 ka BP) - help constrain this chronology.

4.5 Exclusion criteria

Given the scarcity of straight-forward diagnostic criteria for identifying subglacial lakes in the geological record we may also make use of exclusion criteria (i.e. what one would expect to find in a proglacial lake record and *not* in a subglacial lake) as supplementary negative evidence, or to eliminate proglacial lake sediments.

The isolation of subglacial lakes from atmospheric processes precludes wave action or wind-generated current activity. This may be reflected in the lack of upper shoreface longshore troughs and foreshore sediment facies (for example, interfingering large-scale cross-bedding with density segregation of minerals) (Harms, 1979; Prothero & Schwab, 2004). Because the annual melt cycles caused by atmospheric temperature fluctuations are small due to the dampening effect of the thick ice above subglacial lakes it follows that annual laminations in the sediment are unlikely, although non-annual rhythmites are likely (Bentley et al. 2011). Thus, the identification of varved sediments indicates a proglacial rather than subglacial lake origin (Smith, 1978; Smith & Ashley, 1985). Finally, in a subglacial lake environment there should be no evidence of periodic drying out, which would otherwise be indicated by desiccation cracks, mud roll-ups and specific types of trace fossils (Reading, 1998).

5. Concluding Remarks

Subglacial lakes are active components of the still poorly understood subglacial hydrological system. To date, the study of these phenomena has primarily focused on contemporary examples situated beneath the Antarctic Ice Sheet. This has left the palaeo-community playing catch-up, despite the potentially powerful spatial and sedimentological information that these investigations could provide, and is perhaps symptomatic of the shaky foundations upon which palaeo-subglacial lake identification and investigation currently rests.

This paper offers an initial attempt to tackle some of the key uncertainties concerning palaeo-subglacial lake research, including theoretical concepts relating to subglacial lake genesis and their spatial (intra and inter ice-sheet) proclivity, and also the geological evidence that they leave behind following drainage or ice sheet retreat. Significantly it is recognised that subglacial lakes may form under a range of different conditions and processes, including by ice-shelf capture, geothermal heating, within (pre-existing or glacially overdeepened) bed depressions, by changes in the basal thermal regime, ice-surface flattening, or from local increases in meltwater flux or a reduction in subglacial hydraulic conductivity (Figs. 1, 2-4, 6, 7).

Given the mechanisms and conditions governing subglacial lake genesis, we hypothesise that these phenomena were commonplace beneath the former Quaternary ice sheets of the northern hemisphere. In particular, we anticipate that palaeo-ice streams were foci for palaeo-subglacial lakes given the low ice-surface slopes, and warm-bedded conditions generated by strain heating. These systems, as in Antarctica, are likely to have been smaller than some of the larger ice-divide lakes, but nonetheless are thought to have played an active role in the subglacial hydrological system. However, the basal thermal regime of these mid-latitude palaeo-ice sheets differed from the contemporary Antarctic Ice Sheet bed, with the central sectors of the North American and European ice sheets determined to have been cold-bedded and therefore incapable of supporting subglacial lakes. Therefore, the genesis of very large, stable lakes may not have been a feature of the former mid-latitude ice sheets.

A significant conclusion resulting from this discussion is that subglacial lakes should also exist beneath the Greenland Ice Sheet as a result of it being warm based and having numerous ice streams. However, the extent of glacial modification of the glacier bed, particularly of pre-existing tectonic depressions, will mean such lakes may be small, shallow and transitory in comparison to many of those found beneath the Antarctic Ice Sheet. We therefore predict that the coming years will see subglacial lakes identified beneath the Greenland Ice Sheet, and speculate that the current paucity of evidence may result from differences in the spatial distribution, scale and depth of subglacial lakes compared to Antarctica. For instance, geophysical surveying of heavily crevassed ice stream margins remains difficult, whilst the scale at which subglacial lakes are identified is dependent on the resolution of the instruments.

Finally, palaeo-subglacial lake research has great potential for advancing our understanding of subglacial lakes and their association with bed properties and the flow of water at the bed of the ice sheet. This relies on being able to confidently identify palaeo-subglacial lake locations in the geological record, and we therefore see the development of diagnostic criteria, outlined in this paper (Table 2), as a crucial step in introducing more objectivity to their investigation. Although a first attempt, these criteria should provide a useful template for future detailed field investigations. Significantly though, there is no one individual criteria that can provide incontrovertible proof of a former palaeo-subglacial lakes existence. Their identification (over proglacial lakes) in the geological record therefore relies upon the recognition of multiple strands of evidence outlined in Table 2 that when combined build a strong case for the palaeo-subglacial lakes existence.

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References

Alley, R.B., Lawson, D.E., Larson, G.J., Evenson, E.B., Baker, G.S., 2003. Stabilizing feedbacks in glacier-bed erosion. Nature 424, 758-760.

Alley, R.B., Dupont, T.K., Parizek, B.R., Anandakrishnan, S., Lawson, D.E., Larson, G.J., Evenson, E.B., 2006. Outburst flooding and the initiation of ice-stream surges in response to climatic cooling: a hypothesis. Geomorphology 75: 76-89.

Anandakrishnan, S., Catania, G.A., Alley, R.B., Horgan, H.J., 2007. Discovery of till deposition at the grounding line of Whillans Ice Stream. Science 315: 1835-1838.

Anderson, J.B., Kennedy, D.S., Smith, M.J., Domack, E.W., 1991. Sedimentary facies associated with Antarctica's floating ice masses. In: Anderson, J., Ashley, G.M., (Eds.) Glacial marine sedimentation: Palaeoclimatic significance: Boulder, Colorado, Geological Society of America Special Paper 261.

Ballantyne, C.K., 2010. Extent and deglacial chronology of the last British-Irish Ice Sheet: implications of exposure dating using cosmogenic isotopes. Journal of Quaternary Science 25: 515-534.

Banerjee, I., McDonald, B.C., 1975. Nature of esker sedimentation. In: Glaciofluvial and Glaciolacustrine Sedimentation, Jopling, A.V., McDonald, B.C., (Eds.). Special Publications Soc. Econ. Paleont. Miner., Tulsa 23: 132-154.

Bartholomaus, T.C., Anderson, R.S., Anderson, S.P., 2008. Response of glacier basal motion to transient water storage. Nature Geoscience 1: 33-37.

Bartholomew, I., Nienow, P., Mair, D., Hubbard, A., King, M.A., Sole, A., 2010. Seasonal evolution of subglacial drainage and acceleration in a Greenland outlet glacier. Nature Geoscience 3: doi10.1038/ngeo863.

Bateman, M.D. Swift, D.A., Piotrowski, J.A., Sanderson, D.C.W. 2012. Investigating the effects of glacial shearing of sediment on luminescence. Quaternary Geochronology 10: 230-236.

Bell, R.E. 2008. The role of subglacial water in ice-sheet mass balance. Nature Geoscience 1, 297-304.

Bell, R.E., Studinger, M., Tikku, A.A., Clarke, G.K.C., Gutner, M.M., Meertens, C. 2002. Origin and fate of Lake Vostok water frozen to the base of the East Antarctic ice sheet. Nature 416, 307-310.

Bell, R.E., Studinger, M., Fahnestock, M.A. & Shuman, C.A. 2006. Tectonically controlled subglacial lakes on the flanks of the Gambutsev Subglacial Mountains, East Antarctica. Geophysical Research Letters 33: L02504.

Bell, R.E., Studinger, M., Shuman, C.A., Fahnestock, M.A., Joughin, I., 2007. Large subglacial lakes in East Antarctica at the onset of fast-flowing ice streams. Nature 445: 904-407.

Bell, R.E., Ferraccioli, F., Creyts, T.T., Braaten, D., Corr, H., Das, I., Damaske, D., Frearson, N., Jordan, T., Rose, K., Studinger, M., Wolovick, M., 2011. Widespread persistent thickening of the East Antarctic Ice Sheet by freezing from the base. Science 331: 1592-1595.

Benn, D.I. 1996. Subglacial and subaqueous processes near a glacier grounding line: sedimentological evidence from a former ice-dammed lake, Achnasheen, Scotland. Boreas 25: 25-36.

Benn, D.I., Evans, D.J.A., 2010. Glaciers and Glaciation, Second Edition. Hodder Education.

Bennett, M.R., Huddart, D., Thomas, G.S.P., 2002. Facies architecture within a regional glaciolacustrine basin: Copper River, Alaska. Quaternary Science Reviews 21: 2237-2279.

Bentley, M.J., Christoffersen, P., Hodgson, D.A., Smith, A.M., Tulaczyk, S., Le Brocq, A.M., 2011. Subglacial lake sedimentary processes: potential archives of ice sheet evolution, past environmental change and the presence of life. In: Subglacial Antarctic Aquatic Environments. (Eds.). Siegert, M J., Kennicutt, C., Bindschadler, B. AGU Monograph.

Björnsson, H., 2002. Subglacial lakes and jökulhlaups in Iceland. Global and Planetary Change 35: 255-271.

Boulton, G.S., Jones, A.S., Clayton, K.M., Kenning, M.J., 1977. A British ice sheet model and patterns of glacial erosion and deposition, F.W. Shotton, Editor, British Quaternary Studies: Recent Advances, Oxford University Press, Oxford, pp. 231-246.

Boulton, G.S. Smith, G.D., Jones, A.S., Newsome, J., 1985. Glacial geology and glaciology of the last mid-latitude ice sheets. Journal of the Geological Society 142: 447-474.

Boulton, G.S., Clark, C.D., 1990a. A highly mobile Laurentide ice sheet revealed by satellite images of glacial lineations. Nature 346: 813-817.

Boulton, G.S., Clark, C.D., 1990b. The Laurentide ice sheet through the last glacial cycle: the topology of drift lineations as a key to the dynamic behaviour of former ice sheets. Transactions of the Royal Society of Edinburgh: Earth Sciences 81: 327-347.

Boulton, G.S., Peacock, J.D., Sutherland, D.G., 1991. Quaternary. Craig, G.Y, Editor, Geology of Scotland, (third ed.), The Geological Society, London, pp. 503-543.

Boulton, G.S., Payne, A., 1994. Mid-latitude ice sheets through the last glacial cycle: glaciological and geological reconstructions. In: Duplessy, J.-C., Spyridakis, M-T., (Eds.), Long-term climatic variations. NATO ASI Series I, vol. 22, pp. 177-212.

Boulton, G.S. Dongelmans, P., Punkari, M., Broadgate, M., 2001. Palaeoglaciology of an ice sheet through a glacial cycle: the European ice sheet through the Weichselian. Quaternary Science Reviews 20: 591-625.

Boulton, G.S., Hagdorn, M., Maillot, P.B., Zatsepin, S., 2009. Drainage beneath ice sheets: groundwater-channel coupling, and the origin of esker systems from former ice sheets. Quaternary Science Reviews 28: 621-638.

Bouma, A.H., 1962. Sedimentology of some flysch deposits. Elsevier, Amsterdam.

Brennand, T.A., Shaw, J., 1994. Tunnel channels and associated landforms, south-central Ontario: their implications for ice-sheet hydrology. Canadian Journal of Earth Sciences 31: 505-522.

Briner, J.P., Miller, G.H., Davis, P.T., Fink, R.C., 2006. Cosmogenic radionuclides from fjord landscapes supporting differential erosion by overriding ice sheets. Geological Society of America Bulletin 118: 406-420.

Carter, S.P., Blankenship, D.D., Peters, M.E., Young, D.A., Holt, J.W., Morse, D.L., 2007. Radar-based subglacial lake classification in Antarctica. Geochemistry Geophysics Geosystems 8: Q03016.

Chillrud S.N., Pedrozo F.L., Temporetti P.F., Planas H.F., 1994. Chemical weathering of phosphate and germanium in glacial meltwater streams: effects of subglacial pyrite oxidation. Limnol. Oceanogr. 39: 1130–1140.

Cheel, R.J., Rust, B.R., 1982. Coarse grained facies of glacio-marine deposits near Ottowa, Canada. In: Davidson-Arnott, R., Nickling, W., Fahey, B.D., (Eds.). Research in Glacial, Glacio-Fluvial, and Glacio-Lacustrine Systems, Geobooks, Norwich, pp. 279-295.

Chiverrell, R.C., Thomas, G.S.P., 2010. Extent and timing of the Last Glacial Maximum (LGM) in Britain and Ireland: a review. Journal of Quaternary Science 25: 535-549.

Christofferson, P., Tulaczyk, S., 2003. Response of subglacial sediments to basal freeze-on 1. Theory and comparison to observations beneath the West Antarctic Ice Sheet. Journal of Geophysical Research 108: doi:10.1029/2002JB001935.

Christoffersen, P., Tulaczyk, S., Wattrus, N.J., Peterson, J., Quintana-Krupinski, N., Clark, C.D., Sjunneskog, C., 2008. Large subglacial lake beneath the Laurentide Ice Sheet inferred from sedimentary sequences. Geology 36: 563-566.

Clark, C.D., Hughes, A.L.C., Greenwood, S.L., Jordan, C., Sejrup, H.P., 2012. Pattern and timing of retreat of the last British-Irish Ice Sheet. Quaternary Science Reviews 44: 112-146.

Clark, P.U., Walder, J.S., 1994. Subglacial drainage, eskers, and deforming beds beneath the Laurentide and Eurasian ice sheets. Geological Society of America Bulletin 106: 304-314.

Clarke, G.K.C.. 2005. Subglacial Processes. Annual Reviews of Earth and Planetary Science 33, 247–76.

Cook, S.J., Swift, D.A., in press. Glacial overdeepening: its origin and importance in glacial systems and landscapes. Earth Science Reviews.

Copland, L., Sharp, M.J., Nienow, P., 2003. Links between short-term velocity variations and the subglacial hydrology of a predominantly cold polythermal glacier. Journal of Glaciology 49: 337-348.

Creyts, T.T., Clarke, G.K.C., 2010. Hydraulics of subglacial supercooling: Theory and simulations for clear water flows. Journal of Geophysical Research-Earth Surface 115.

Cuffey, K.M., Patterson, W.S.B., 2010. The Physics of Glaciers, Fourth Edition. Elsevier.

Cutler, P.M., Colgan, P.M., Mickelson, D.M., 2002. Sedimentologic evidence for outburst floods from the Laurentide Ice Sheet margin in Wisconsin, USA: implications for tunnel-channel formation. Quaternary International 90: 23-40.

De Angelis, H., Kleman, J., 2005. Palaeo-ice streams in the northern Keewatin sector of the Laurentide ice sheet. Annals of Glaciology 42: 135-144.

De Angelis, H., Kleman, J., 2007. Palaeo-ice streams in the Foxe/Baffin sector of the Laurentide Ice Sheet. Quaternary Science Reviews 26: 1313-1331.

De Conto, R.M., Pollard, D., 2003. A coupled climate-ice sheet modelling approach to the Early Cenozoic history of the Antarctic ice sheet. Palaeogeography, Palaeoclimatology, Palaeoecology 198: 39-52.

Denton, G.H., Hughes, T.J., 1981. The Last Great Ice Sheets, pp. 484, Wiley-Interscience, New York.

Domack, E., O'Brien, P., Harris, P., Taylor, F., Quilty, G., De Santis, L., Raker, B., 1998. Late Quaternary sediment facies in Prydz Bay, East Antarctica and their relationship to glacial advance onto the continental shelf. Antarctic Science 10: 236-246.

Domack, E.W., Jacobson, E.A., Shipp, S., Anderson, J.B., 1999. Late Pleistocene – Holocene retreat of the West Antarctic Ice-Sheet system in the Ross Sea: Part 2 – sedimentologic and stratigraphic signature. GSA Bulletin 111: 1517-1536.

Dowdeswell, J.A., Siegert, M.J., 2002. The physiography of modern Antarctic subglacial lakes. Global Planetary Change 35: 221-236.

Dunlop, P., Clark, C.D., Hindmarsh, R.C.A., 2008. Bed ribbing instability explanation: testing a numerical model of ribbed moraine formation arising from coupled flow of ice and subglacial sediment. Journal of Geophysical Research 113: F03005.

Dyke, A.S., Prest, V.K., 1987. Late Wisconsinan and Holocene history of the Laurentide Ice Sheet. Géographie Physique et Quaternaire 41: 237-263.

Dyke, A.S., Andrews, J.T., Clark, P.U., England, J.H., Miller, G.H., Shaw, J., Veillette, J.J., 2002. The Laurentide and Innuitian ice sheets during the Last Glacial Maximum. Quaternary Science Reviews 21: 9-31.

Engelhardt, H.F., Kamb, B., 1997. Basal hydraulic system of a West Antarctic ice stream: constraints from borehole observations. Journal of Glaciology 43: 207-230.

Erlingsson, . U., 1994a. The 'captured ice shelf' hypothesis and its applicability to the Weichselian glaciation. Geografiska Annaler 76A: 1-12.

Erlingsson, U., 1994b. A computer model along a flow-line of an ice dome – 'captured ice shelf'. Geografiska Annaler 76A: 13-24.

Erlingsson, U,. 2006. Lake Vostok behaves like a 'captured lake' and may be near to creating an Antarctic jökulhlaup. Geografiska Annaler 88A: 1-7.

Evans, J., Pudsey, C.J., 2002. Sedimentation associated with Antarctic Peninsula ice shelves: implications for palaeoenvironmental reconstructions of glacimarine sediments. Journal of the Geological Society 159: 233-237.

Evans, D.J.A., Rea, B.R. Hiemstra, J.F., Ó Cofaigh, C. 2006. A critical assessment of subglacial mega-floods: a case study of glacial sediments and landforms in south-central Alberta, Canada. Quaternary Science Reviews 25: 1638-1667.

Evans, D.J.A., Livingstone, S.J., Vieli, A., Ó Cofaigh, C., 2009. The palaeoglaciology of the central sector of the British and Irish Ice Sheet: reconciling glacial geomorphology and preliminary ice sheet modelling. Quaternary Science Reviews 28: 739-757.

Evatt, G.W., Fowler, A.C., Clark, C.D., Hulton, N.R.J., 2006. Subglacial floods beneath ice sheets. Philosophical Transactions of the Royal Society of London, Series A 374: 1769-1794.

Filina, I.Y., Blankenship, D.D., Thoma, M., Lukin, V.V., Masalov, V.N., Sen, M.K. 2008. Earth and Planetary Science Letters 276, 106-114.

Fricker, H.A., Scambos, T., Bindschadler, R., Padman, L., 2007. An active subglacial water system in West Antarctica mapped from space. Science 315: 1544-1548.

Fricker, H.A., Scambos, T., 2009. Connected subglacial lake activity on lower Mercer and Whillans Ice Streams, West Antarctica, 2003-2008. Journal of Glaciology 55: 303-315.

Gilbert, R., 1990. Rafting in glacimarine environments. Geological Society, London, Special Publication 53: 105-120.

Gorrell, G., Shaw, J., 1991. Deposition in an esker, bead and fan complex, Lanark, Ontario, Canada. Sedimentary Geology 72: 285-314.

Gray, L., Joughin, S., Tulaczyk, S., Spikes, V.B., Bindschadler, R., Jezek, K., 2005. Evidence for subglacial water transport in the West Antarctic ice sheet through three-dimensional satellite radar interfermometry. Geophysical Research Letters 32: L03501.

Greenwood, S.L., Clark, C.D., 2008. Subglacial bedforms of the last Irish Ice Sheet. Journal of Maps v2008: 332-357.

Greenwood, S.L., Clark, C.D., 2009a. Reconstructing the last Irish Ice Sheet 1: changing flow geometries and ice flow dynamics deciphered from the glacial landform record. Quaternary Science Reviews 28: 3085-3100.

Greenwood, S.L., Clark, C.D., 2009b. Reconstructing the last Irish Ice Sheet 2: a geomorphologically-driven model of ice sheet growth, retreat and dynamics. Quaternary Science Reviews 28: 3101-3123.

Hagdorn, M.K.M., 2003. Reconstruction of the past and forecast of the future European and British Ice Sheets and associated sea-level change. Unpublished Ph.D. thesis, Edinburgh.

Harms, J.C., 1979. Primary sedimentary structures. Annual Review of Earth and Planetary Sciences 7: 227-248.

Hodgson, D.A., Roberts, S.J., Bentley, M.J., Smith, J.A., Johnson, J.S., Verleyen, E., Vyverman, W., Hodson, A.J., Leng, M.J., Cziferszky, A., Fox, A.J., Sanderson, D.C.W., 2009a. Exploring former subglacial Hodgson Lake, Antarctica Paper I: site description, geomorphology and limnology. Quaternary Science Reviews 28: 2295-2309.

Hodgson, D.A., Roberts, S.J., Bentley, M.J., Carmichael, E.L., Smith, J.A., Verleyen, E., Vyverman, W., Geissler, P., Leng, M.J., Sanderson, D.C.W., 2009b. Exploring former subglacial Hodgson Lake, Antarctica Paper II: palaeolimnology. Quaternary Science Reviews 28: 2310-2325.

Hodson, A.J., Anesio, A.M., Tranter, M., Fountain, A.G., Osborn, A.M., Priscu, J., Laybourn-Parry, J., Sattler, B., 2008. Glacial Ecosystems, Ecological Monographs 78, 41-67.

Hoffmann, K., Piotrowski, J.A., 2001. Till mélange at Amsdorf, central Germany: sediment erosion, transport and deposition in a complex, soft-bedded subglacial system. Sedimentary Geology 140, 215-234.

Hubbard, B., Sharp, M.J., Willis, I.C., Neilsen, M.K., Smart, C.C., 1995. Borehole water-level variations and the structure of the subglacial hydrological system of Haut Glacier d'Arolla, Valais, Switzerland. Journal of Glaciology 41: 572-583.

Hubbard, A. Bradwell, T., Golledge, N., Hall, A., Patton, H., Sugden, D., Cooper, R., Stoker, M., 2009. Dynamic cycles, ice streams and their impact on the extent, chronology and deglaciation of the British-Irish Ice Sheet. Quaternary Science Reviews 28: 758-776.

Hughes, A.L.C., 2008. The last British Ice Sheet: a reconstruction based on glacial landforms. University of Sheffield, unpublished Ph.D.

Hughes, A.L.C., Clark, C.D., Jordan, C.J., 2010. Subglacial bedforms of the last British Ice Sheet. Journal of Maps v2010: 543-563.

Jones, I., Munhoven, G., Tranter, M., Huybrechts, P., Sharp, M. 2002. Modelled glacial and non-glacial HCO_3^-, Si and Ge fluxes since the LGM: little potential for impact on atmospheric CO_2 concentrations and a potential proxy of continental chemical erosion, the marine Ge:Si ratio, Global and Planetary Change, 33: 139-153.

Jordan, T.A., Ferraccioli, F., Corr, H., Graham, A., Armadillo, E., Bozzo, E. 2010. Hypothesis for mega-outburst flooding from a palaeo-subglacial lake beneath the East Antarctic Ice Sheet. Terra Nova 22, 283-289.

Josenhans, H.W., Zevenhuizen, J., 1990. Dynamics of the Laurentide Ice Sheet in Hudson Bay, Canada. Marine Geology 92: 1-26.

Joughin, I., Das, S.B., King, M.A., Smith, B.E., Howat, I.A., Moon, T., 2008. Seasonal speedup along the western flank of the Greenland Ice Sheet. Science 320: 781-783.

Kamb, B., 2001. Basal zone of the West Antarctic ice streams and its role in lubrication of their rapid motion. In: Alley, R.B., Bindschadler, R.A. (Eds.). The West Antarctic Ice Sheet: behaviour and environment. Washington DC, American Geophysical Union, 157-199. (Antarctic Research Series 77.).

Kehew, A.E., Piotrowski, J.A., Jørgensen, F., 2012. Tunnel valleys: Concepts and controversies — A review. Earth-Science Reviews 113: 33-58.

Keller, W.D., Reesman, A.L., 1963. Glacial milks and their laboratory-simulated counterparts. Geological Society of America. Bulletin, 74(I): 61-76.

Kilfeaether, A.A., Ó Cofaigh, C., Lloyd, J.M., Dowdeswell, J.A., Xu, S., Moreton, S.G., 2011. Ice-stream retreat and ice-shelf history in Marguerite Trough, Antarctic Peninsula: sedimentological and foraminiferal signatures. GSA Bulletin 123: 997-1015.

Kilfeather, A.A. Ó Cofaigh, C., Lloyd, J.M., Dowdeswell, J.A., Xu, S., Moreton S.G., 2011. Ice stream retreat and ice shelf history in Marguerite Bay, Antarctic Peninsula: sedimentological and formainiferal signatures. Geological Society of America Bulletin 123, 997-1015.

Kleman, J., Hättestrand, C., Borgstrom, I, Stroeven, A.P., 1997. Fennoscandian palaeoglaciology reconstructed using a glacial geological inversion model. Journal of Glaciology 43: 283-299.

Kleman, J., Hättestrand, C., 1999. Frozen-bed Fennoscandian and Laurentide ice sheets during the Last Glacial Maximum. Nature 402: 63-66.

Kleman, J., Glasser, N.F., 2007. The subglacial thermal organisation (STO) of ice sheets. Quaternary Science Reviews 26: 585-597.

Knight, J., 2002. Subglacial patterns, subglacial meltwater events, and Late Devensian ice sheet dynamics in north-central Ireland. Global and Planetary Change 35: 237-253.

Kristensen, T.B., Huuse, M., Piotrowski, J.A., Clausen, O.R., 2007. A morphometric analysis of tunnel valleys in the eastern North Sea based on 3D seismic data. Journal of Quaternary Science 22: 801-815.

Lambeck, K., 1993. Glacial Rebound of the British Isles – I. Preliminary model results. Geophysical Journal International 115: 941-959.

Lambeck, K., 1995. Late Devensian and Holocene shorelines of the British Isles and North Sea from models of glacio-hydro-isostatic rebound. Journal of the Geological Society of London 152: 437-448.

Lambeck, K., Smither, C., Johnston, P., 1998. Sea-level change, glacial rebound and mantle viscosity for northern Europe. Geophysical Journal International 134: 102-144.

Langley, K., Kohler, J., Matsuoka, K., Sinisalo, A., Scambos, T., Neumann, T., Muto, A., Winther, J.-G., Albert, M., 2011. Recovery Lakes, East Antarctica: radar assessment of sub-glacial water extent. Geophysical Research Letters 38: L05501.

Lanoil, B., Skidmore, M., Priscu, J.C., Han, S., Foo, W., Vogel, S.W., Tulaczyk, S., Engelhardt, H., 2009. Bacteria beneath the West Antarctic Ice Sheet. Environmental Microbiology 11: 609-615.

Larter, R.D., Vanneste, L.E., 1995. Relict subglacial deltas on the Antarctic Peninsula outer shelf. Geology 23: 33-36.

Lawson, D.E., Strasser, J.C., Evenson, E.B., Alley, R.B., Larson, G.J., Arcone, S.A., 1998. Glaciohydraulic supercooling: a freeze-on mechanism to create stratified, debris-rich basal ice: I. Field evidence. Journal of Glaciology 44, 547-562.

Lekens, W.A.H., Sejrup, H.P., Haflidason, H., Petersen, G.Ø., Hjelsteun, B., Knorr, G. 2005. Laminated sediments preceding Heinrich event 1 in the Northern North Sea and Southern Norwegian Sea: Origin, processes and regional linkage. Marine Geology 216: 27-50.

Lesemann, J.-E., Brennand, T.A., 2009. Regional reconstruction of subglacial hydrology and glaciodynamic behaviour along the southern margin of the Cordilleran Ice Sheet in British Columbia, Canada and northern Washington State, USA. Quaternary Science Reviews 28, 2420-2444.

Linton, D.L., 1963. The forms of glacial erosion. Transactions and Papers (Institute of British Geographers) 33, 1-28.

Livingstone, S.J., Ó Cofaigh, C., Evans, D.J.A., 2009. Glacial geomorphology of the central sector of the last British-Irish Ice Sheet. Journal of Maps, v2008: 358-377.

Livingstone, S.J., Ó Cofaigh, C., Stokes, C.R., Hillenbrand, C.-D., Vieli, A., Jamieson, S.S.J., 2012. Antarctic palaeo-ice streams. Earth Science Reviews 111, 90-128

Llubes, M., Lanseau, C., Rémy, F., 2006. Relations between basal conditions, subglacial hydrological networks and geothermal flux in Antarctica. Earth and Planetary Science Letters 241: 655-662.

Marshall, S.J., Clark, P.U., 2002. Basal temperature evolution of North American ice sheets and implication for the 100-kyr cycle. Geophysical Research Letters 29, 2214, doi:10.1029/2002GL015192.

McCabe, A.M., Ó Cofaigh, C. 1994. Sedimentation in a subglacial lake, Enniskerry, eastern Ireland. Sedimentary Geology 91: 57-95.

Munro-Stasiuk, M.J., 2003. Subglacial Lake McGregor, south-central Alberta, Canada. Sedimentary Geology 160: 325-350.

Nemec, W., Lønne, I., Blikra, L., 1999. The Kregnes moraine in Gauldalen, west-central Norway: anatomy of a Younger Dryas proglacial delta in a palaeofjord basin. Boreas 28, 454-476.

Nye, J.F., 1976. Water flow in glaciers: jökulhlaups, tunnels and veins. Journal of Glaciology 17: 181-207.

Ó Cofaigh, C., Evans, D.J.A., Smith, I.R. 2010. Large-scale reorganization and sedimentation of terrestrial ice streams during late Wisconsinan Laurentide Ice Sheet deglaciation. GSA Bulletin 122: 743-756.

O'Connor, J., 1993. Hydrology, hydraulics and geomorphology of the Bonneville Flood. GSA Special Paper 274, pp. 83.

Oswald, G.K.A., Gogineni, S.P., 2008. Recovery of subglacial water extent from Greenland radar survey data. Journal of Glaciology 54: 94-106.

Ottesen, D., Stokes, C.R., Rise, L., Olsen, L., 2008. Ice-Sheet dynamics and ice streaming along the coastal parts of northern Norway. Quaternary Science Reviews 27: 922-940.

Pattyn, F., 2003. A new three-dimensional higher-order thermomechanical ice sheet model: Basic sensitivity, ice stream development, and ice flow across subglacial lakes. Journal of Geophysical Research 108: B8, 2382, doi:10.1029/2002JB002329.

Pattyn, F., 2004. Comment on the comment by M.J. Siegert on “A numerical model for an alternative origin of Lake Vostok and its exobiological implications for Mars” by N.S. Duxbury et al., Journal of Geophysical Research 109: E11004.

Pattyn, F., De Smedt, B., Souchez, R. 2004. Influence of subglacial Vostok lake on the regional ice dynamics of the Antarctic ice sheet: a model study. Journal of Glaciology 50: 583-589.

Pattyn, F., 2008. Investigating the stability of subglacial lakes with a full Stokes ice-sheet model. Journal of Glaciology 54: 353-361.

Pattyn, F., 2010. Antarctic subglacial conditions inferred from a hybrid ice sheet/ice stream model. Earth and Planetary Science Letters 295: 451-561.

Payne, A.J., 1995. Limit cycles in the basal thermal regime of ice sheets. Journal of Geophysical Research 100: 4249-4263.

Peltier, W.R., 2004. Global isostasy and the surface of the ice age Earth: the ICE-5G(VM2) model and GRACE. Annual Review of Earth and Planetary Science 32: 111-149.

Peters, L.E., Anandakrishnan, S., Alley, R.B., Smith, A.M., 2007. Extensive storage of basal meltwater in the onset region of a major West Antarctic ice stream. Geology 35: 251-254.

Phillips, W.M., Hall, A.M., Mottram, R., Fifield, K., Sugden, D.E., 2006. Cosmogenic exposure ages of tors and erratics on the Cairngorm plateau, Scotland: timescales for the development of a classic landscape of selective linear erosion. Geomorphology 73: 222-245.

Piotrowski, J.A., 1994. Tunnel-valley formation in northwestern Germany – geology, mechanisms of formation and subglacial bed conditions for the Bornhöved tunnel valley. Sedimentary Geology 89(1/2), 107-141.

Piotrowski, J.A., 2006. Groundwater under ice sheets and glaciers. In: Knight, P.G., (Ed.): Glacier Science and Environmental Change, 50-59. Blackwell Publishing, Oxford.

Postma, G., 1986. Classification for sediment-gravity flow deposits based on flow conditions during sedimentation. Geology 14, 291-294.

Powell, R.D., Domack, E.W. 1995. Modern glaciomarine environments. In Menzies, J., (Ed.), Glacial Environments, Vol 1: Modern Glacial Environments: Processes, Dynamics and Sediments. Butterworth-Heinemann, Oxford, pp. 445-486.

Powell, R.D., Dawber, M., McInnes, J.N., Pyne, A.R., 1996. Observations of the grounding-line area at a floating glacier terminus. Annals of Glaciology 22: 217-223.

Powell, R.D. 2003., Subaquatic landsystems: fjords. In Evans, D.J.A., (Ed.), Glacial landsystems. Arnold, London, 313-347.

Price, P.B., Nagornov, O.V., Bay, R., Chirkin, D., He, Y., Miocinovic, P., Richards, A., Woschnagg, K., Kod, B., Zagorodinov, V., 2002. Temperature profile for glacial ice at the South Pole: implications for life in a nearby subglacial lake. PNAS 99: 7844-7847.

Priscu, J.C., Tulaczyk, S., Studinger, M., Kennicutt II, M.C., Christner, B.C., Foreman, C.M., 2008. Antarctic Subglacial Water: Origin, Evolution and Microbial Ecology. In W. Vincent and J. Laybourn-Parry (Eds.), Polar Limnology. Oxford University Press, pp. 119-135.

Prothero, D.R., Schwab, F., 2004. Sedimentary Geology. 2nd Edition, W.H. Freeman & Co., 557 pp.

Reading, H.G., 1998. Sedimentary Environments, Processes, Facies and Stratigraphy. 3rd Edition, Blackwell, 688 pp.

Ross, M., Lajeunesse, P., Kosar, K.G.A., 2011. The subglacial record of northern Hudson Bay: insights into the Hudson Strait Ice Stream catchment. Boreas 40, 73-91.

Röthlisberger, H., 1972. Water pressure in intra- and subglacial channels. Journal of Glaciology 11: 177-203.

Royston-Bishop, G., Priscu, J.C., Tranter, M., Christner, B., Siegert, M.J., Lee, V., 2005. Incorporation of particulates into accreted ice above subglacial Vostok lake, Antarctica. Annals of Glaciology 40: 145-150.

Russell, A.J., Knudsen, Ó., 1999. An ice-contact rhythmite (turbidite) succession deposited during the November 1996 catastrophic outburst flood (jökulhlaup), Skeiðarárjökull, Iceland. Sedimentary Geology 127: 1-10.

Scambos, T.A., Berthier, E., Shuman, C.A., 2011. The triggering of subglacial lake drainage during rapid glacier drawdown: Crane Glacier, Antarctic Peninsula. Annals of Glaciology 52: 74-82.

Schoof, C., 2010. Ice-sheet acceleration driven by melt supply variability. Nature 468: 803-806.

Shaw, J., Kvill, D., Rains, B., 1989. Drumlins and catastrophic subglacial floods. Sedimentary Geology 62: 177-202.

Shipp, S.S., Anderson, J.B., Domack, E.W., 1999. Late Pleistocene-Holocene retreat of the West Antarctic Ice-Sheet system in the Ross Sea: Part 1 – Geophysical results. GSA Bulletin 111: 1486-1516.

Smith, G.A., 1993. Missoula flood dynamics and magnitudes inferred from sedimentology of slack-water deposits on the Columbia Plateau, Washington. Geological Society of America Bulletin 105: 77-100.

Skidmore, M., Tranter, M., Tulaczyk, S., Lanoil. B., 2010. Hydrochemistry of ice sheet beds – Evaporitic or microbial effects? Hydrological Processes 24: 517-523.

Shepherd, A., Hubbard, A., Nienow, P., King, M., McMillan, M., Joughin, I., 2009. Greenland ice sheet motion coupled with daily melting in late summer. Geophysical Research Letters 36: L01501.

Shreve, R.L. 1972. Movement of water in glaciers. Journal of Glaciology 11, 205-214.

Siegert, M.J., 2000. Antarctic subglacial lakes. Earth-Science Reviews 50, 29-50.

Siegert, M.J., Tranter, M., Ellis-Evans, J.C. Priscu, J.C., Lyons, W.B., 2003. The hydrochemistry of Lake Vostok and the potential for life in Antarctic subglacial lakes. *Hydrological Processes* **17**: 795–814.

Siegert, M.J., 2005. Lakes beneath the ice sheet: the occurrence, analysis, and future exploration of Lake Vostok and other Antarctic subglacial lakes. Annual Reviews of Earth and Planetary Science 33: 215-245.

Siegert, M.J., Behar, A., Bentley, M., Blake, D., Bowden, S., Christoffersen, P., Cockell, C., Corr, H., Cullen, D.C., Edwards, H., Ellery, A., Ellis-Evans, C., Griffiths, G., Hindmarsh, R., Hodgson, D.A., King, E., Lamb, H., Lane, L., Makinson, K., Mowlem, M., Parnell, J., Pearce, D.A., Priscu, J., Rivera, A., Sephton, M.A., Sims, M.R., Smith, A.M., Tranter, M., Wadham, J.L., Wilson, G., Woodward, J. 2007. Exploration of Ellsworth Subglacial Lake: a concept paper on the development, organisation and execution of an experiment to explore, measure and sample the environment of a West Antarctic subglacial lake. Reviews in Environmental Science and Biotechnology 6: 161-179.

Siegert, M.J., Clarke, R.J., Mowlem, M., Ross, N., Hill, C.S., Tait, A., Hodgson, D., Parnell, J., Tranter, M., Pearce, D., Bentley, M.J., Cockell, C., Tsaloglou, M.-N., Smith, A., Woodward, J., Brito, M.P., Waugh, E. 2012. Clean access, measurement, and sampling of Ellsworth Subglacial Lake: A method for exploring deep Antarctic subglacial lake environments. Review of Geophysics 50: RG1003, 40 pp.

Skidmore, M.L., Sharp, M.J., 1999. Drainage system behaviour of a High-Arctic polythermal glacier. Annals of Glaciology 28: 209-215.

Skidmore, M.L., Tranter, M., Tulaczyk, S., Lanoil, B., 2010. Hydrochemistry of ice stream beds – evaporitic or microbial effects? Hydrological Processes 24: 517-523.

Smith, N.D., 1978. Sedimentation processes and patterns in a glacier fed with a low sediment input. Canadian Journal of Earth Science 15, 741-756.

Smith, N.D., Ashley, G.M. 1985. Proglacial lacustrine environment. In Ashley, G.M., Shaw, J., Smith, N.D. (Eds.): Glacial Sedimentary Environments, 135-212. Society of Palaeontologists and Mineralogists, Tulsa.

Smith, B.E., Fricker, H.A., Joughin, I.R., Tulaczyk, S., 2009. An inventory of active subglacial lakes in Antarctica detected by ICESat (2003-2008). Journal of Glaciology 55: 573-595.

Smith, A.M., Bentley, C.R., Bingham, R.G., Jordan, T.A. 2012. Rapid subglacial erosion beneath Pine Island Glacier, West Antarctica. Geophysical Research Letters 39, L12501.

Staiger, J.K.W., Gosse, J.C., Fastook, J., Gray, J.T., Stockli, D.F., Stockli, L., Finkel, R., 2005. Quaternary relief generation by polythermalglacier ice. Earth Surface Processes and Landforms 30: 1145-1159.

Stearns, L.A., Smith, B.E., Hamilton, G.S., 2008. Increased flow speed on a large East Antarctic outlet glacier caused by subglacial floods. Nature Geoscience 1: 827-831.

Stokes, C.R., Clark, C.D., 1999. Geomorphological criteria for identifying Pleistocene ice streams. Annals of Glaciology 28: 67-74.

Stokes, C.R., Clark, C.D., 2001. Palaeo-ice streams. Quaternary Science Reviews 20: 1437-1457.

Studinger, M., Karner, G.D., Bell, R.E., Levin, V., Raymond, C.A., Tikku, A.A., 2003. Geophysical models for the tectonic framework of the Lake Vostok region, East Antarctica. Earth and Planetary Science Letters 216: 663-677.

Sugden, D.E., 1977. Reconstruction of the morphology, dynamics and thermal characteristics of the Laurentide Ice Sheet at its maximum. Arctic and Alpine Research, 9: 21-47.

Sutinen, R., Piekkari, M., Middleton, M., 2009. Glacial geomorphology in Utsjoki, Finnish Lapland proposes Younger Dryas fault-instability. Global and Planetary Change 69: 16-28.

Sugden, D.E., Marchant, D.R., Denton, G.H., 1993. The case for a stable East Antarctic Ice Sheet: the background. Geografiska Annaler: Series A, Physical Geography 75: 151-154.

Sugden, D.E., 1996. The East Antarctic Ice Sheet: unstable ice or unstable ideas? Transactions of the Institute of British Geographers 21: 443-454.

Svendsen, J.I., Alexanderson, H., Astakhov, V.I., Demidov, I., Dowdeswell, J.A., Funder, S., Gataullin, V., Henriksen, M., Hjort, C., Houmark-Nielsen, M., Hubberten, H.W., Ingólfsson, Ó, Jakobsson, M., Kjaer, K.H., Larsen, E., Lokrantz, H., Lunjja, J.P., Lysa, A., Mangerud, J., Matiouchkov, A., Murray, A., Möller, P., Niessen, F., Nikolskaya, O., Polyak, L., Saarnisto, M., Siegert, C., Siegert, M.J., Spielhagen, R.F., Stein, R., 2004. Late Quaternary ice sheet history of northern Eurasia. Quaternary Science Reviews 23: 1229-1271.

Swift, D.A., Sanderson, D.C.W., Nienow, P.W., Bingham, R.G., Cochrane, I.C., 2011. Anomolous luminescence of subglacial sediment at Haut Glacier d'Arolla, Switzerland – a consequence of resetting at the glacier bed? Boreas 40: 446-458.

Szewczyk, J., Nawrocki, J., 2011. Deep-seated relict permafrost in northeastern Poland. Boreas 40: 385-388.

Tabacco, I.E., Cianfarra, P., Forieri, A., Salvini, F., Zirizotti, A., 2006. Physiography and tectonic setting of the subglacial lake district between Vostok and Belgica subglacial highlands (Antarctica). Geophysical Journal International 165: 1029-1040.

Tarasov, L., Peltier, W.R., 2004. A geophysically constrained ensemble analysis of the deglacial history of the North American ice sheet complex. Quaternary Science Reviews 23: 359-388.

Taylor, A., Blum, J.D., 1995. Relation between soil age and silicate weathering rates determined from the chemical evolution of a glacial chronosequence, Geology, 23: 979 – 982.

Thoma, M., Grosfeld, K., Mayer, C. 2008. Modelling accreted ice in the subglacial Lake Vostok, Antarctica. Geophysical Research Letters 35, L11504.

Thoma, M., Grosfeld, K., Mayer, C., Pattyn, F. 2012. Ice-flow sensitivity to boundary processes: a coupled model study in the Vostok Subglacial Lake area, Antarctica. Annals of Glaciology 53, 173-180.

Tranter, M., Skidmore, M. L., Wadham, J.L., 2005. Hydrological controls on microbial communities in subglacial environments. *Hydrological Processes* **19**: 996–998.

Tulaczyk, S., Hossainzadeh, S., 2011. Antarctica's deep frozen "lakes". Science 331: 1524-1525.

Uemura, T., Tanigucji, M., Shibuya, K., 2011. Submarine groundwater discharge in Lützow-Holm Bay, Antarctica. Geophysical Research Letters 38, L08402.

Wadham, J. L., Bottrell, S., Tranter, M., Raiswell, R., 2004. Stable isotope evidence for microbial sulphate reduction at the bed of a polythermal high Arctic glacier, Earth Planet. Sci. Lett., 219: 341–355.

Wadham, J.L., Tranter, M., Tulaczyk, S., Sharp, M., 2008. Subglacial methanogenesis: a potential climatic amplifier? Global Biogeochemical Cycles 22: GB2021.

Wadham, J.L., Tranter, M., Skidmore, M., Hodson, A.J., Priscu, J., Lyons, W.B., Sharp, M., Wynn, P., Jackson, M. 2010. Biogeochemical weathering under ice: size matters. Global Biogeochemical Cycles 24: GB3025.

Wingham, D.J., Siegert, M.J., Shepherd, A., Muir, A.S., 2006. Rapid discharge connects Antarctic subglacial lakes. Nature 440: 1033-1036.

Winsborrow, C.M., Clark, C.D., Stokes, C.R., 2004. Ice streams of the Laurentide Ice Sheet. Géographie physique et Quaternaire 58: 269-280.

Wright, A., Siegert, M.J., 2011. The identification and physiographical setting of Antarctic subglacial lakes: an update based on recent advances. In: Subglacial Antarctic Aquatic Environments. (Eds.). Siegert, M J., Kennicutt, C., Bindschadler, B. AGU Monograph.

Zwally, H.J., Abdalati, W., Herring, T., Larson, K., Saba, J., Steffen, K., 2002. Surface melt-induced acceleration of Greenland Ice-Sheet flow. Science 297: 218-222.

Tables:

Table 1: Classification of subglacial lake types (modified from Wright & Siegert, 2011).

Table 2: Diagnostic criteria for the identification of palaeo-subglacial lakes in the geological record.

Figures:

Fig. 1: Cartoon illustrating the evolution of a Captured Ice Shelf (adapted from Erlingsson, 1994a). A: advance of a glacier into a proglacial lake and the development of an ice shelf; and B: where the glacier grounds on the downflow side of the lake an ice rim will form creating an ice-air reversal, which traps the water under the ice. As the ice continues to thicken (grey to black lines) the ice rim may be removed and drainage can occur (solid lines indicate thickening over a large subglacial lake capable of modifying the overlying ice and dotted line indicates thickening over a small subglacial lake).

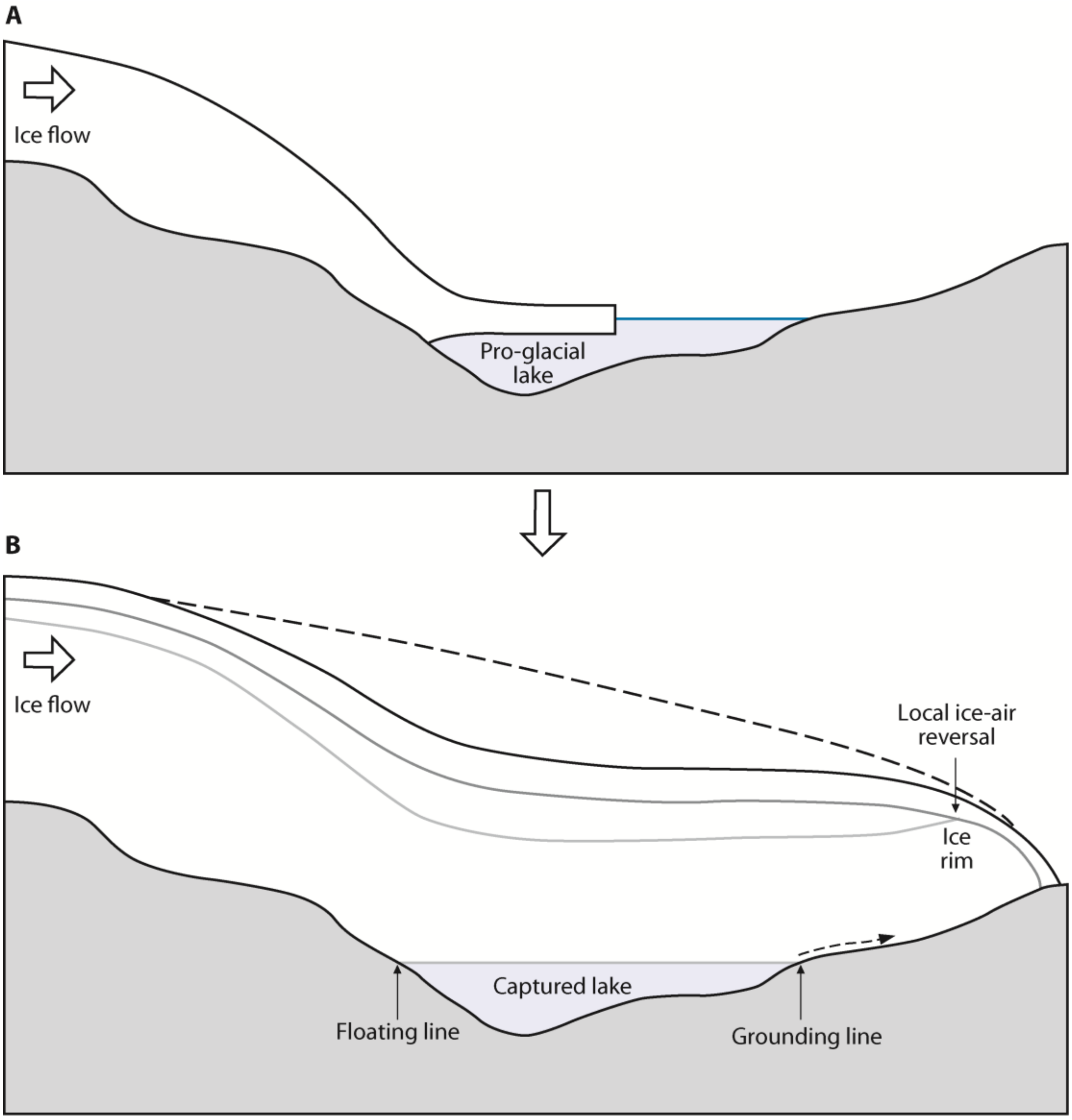


Fig. 2: Cartoon (not drawn to scale) illustrating geothermal control on subglacial lake formation and evolution (adapted from Björnsson, 2002).

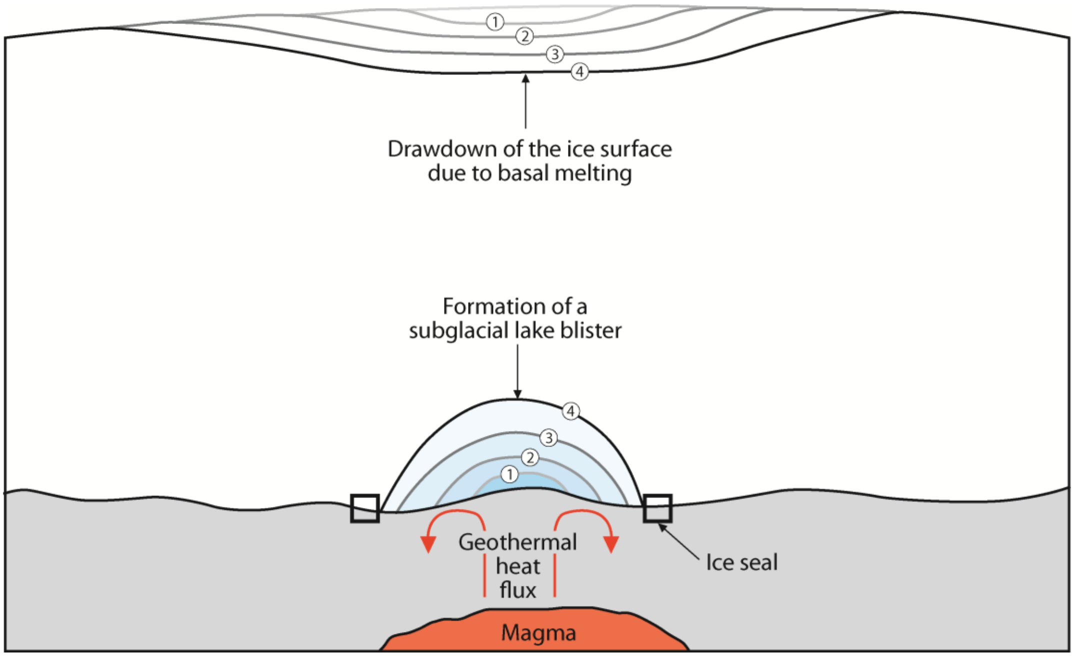


Fig. 3: Cartoon (not drawn to scale) illustrating the effect of ice sheet erosion on subglacial lake genesis. A: Glacial overdeepening; B: the overdeepened basin eventually becomes deep enough that subglacial meltwater cannot escape and begins to pond. As a subglacial lake develops sediment will start to accumulate; and C: sedimentation will cause a shallowing of the basin and this may eventually result in the subglacial lake being lost and erosion re-occurring. Thus it is possible that subglacial lakes will repeatedly form and drain over long timescales due to cycles of sedimentation and erosion.

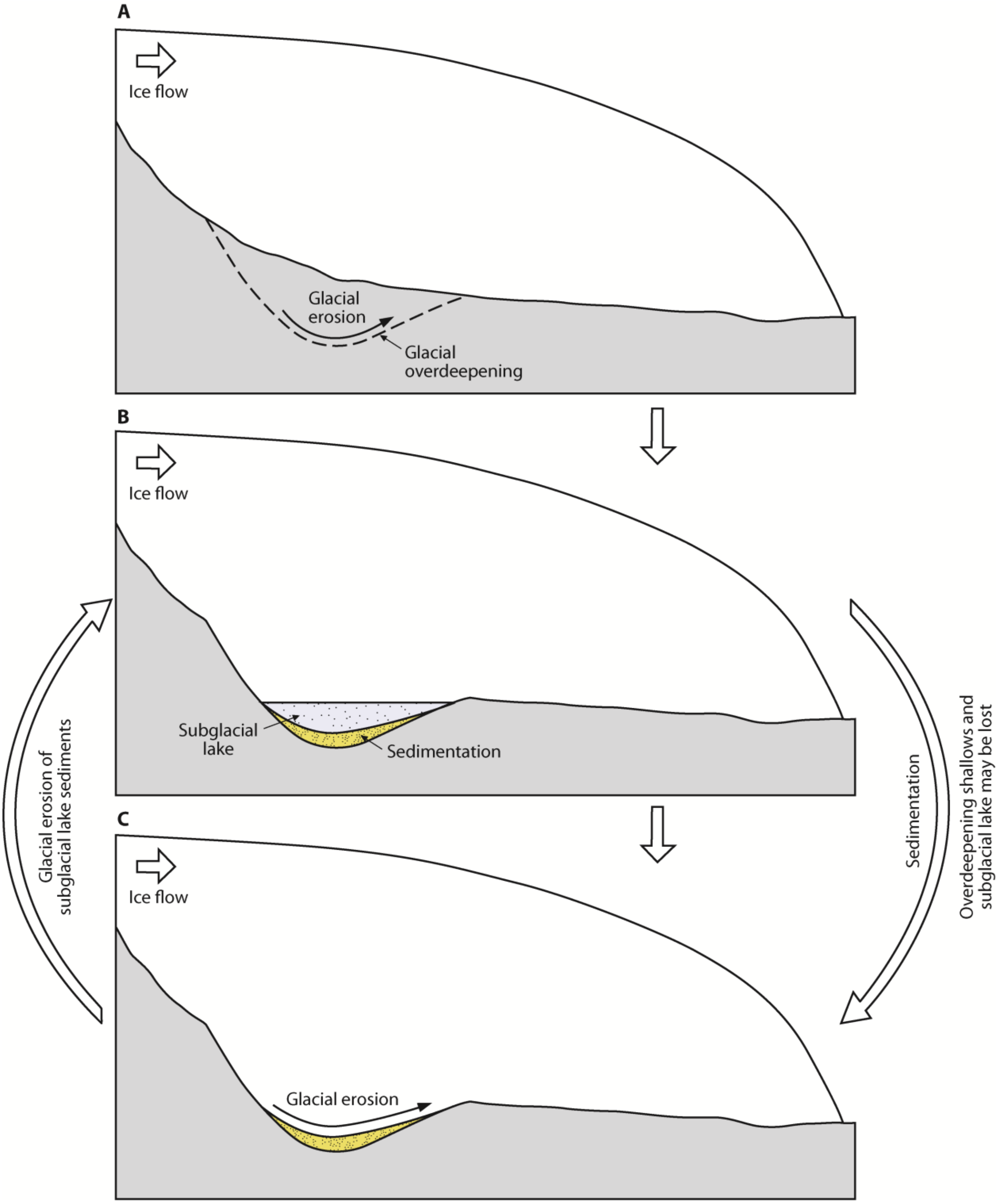


Fig. 4: Cartoon (not drawn to scale) illustrating how changes in the ice-sheet surface slope can influence subglacial lake genesis and drainage. As the ice mass advances (grey to black lines) or retreats (black to grey lines) the ice surface slope will also vary. Thus, as the ice surface slopes flattens a subglacial lake forms in the depression, if meltwater is present. Conversely, a steepening of the ice surface may cause the water to drain.

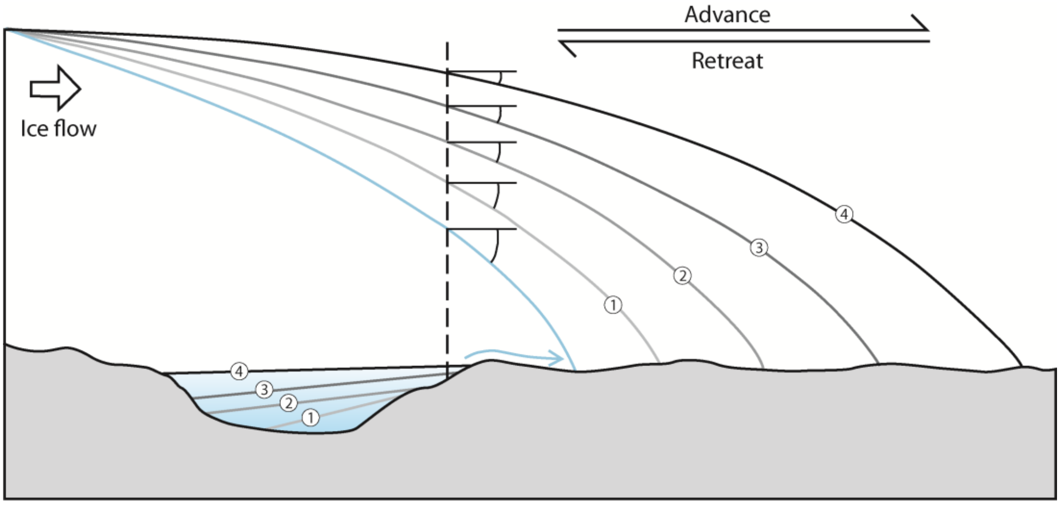


Fig. 5: The theoretical effect of a large subglacial lake on the topography and surface velocity of the ice sheet (modified from Cuffey & Patterson, 2010). The x-axis is distance (km).

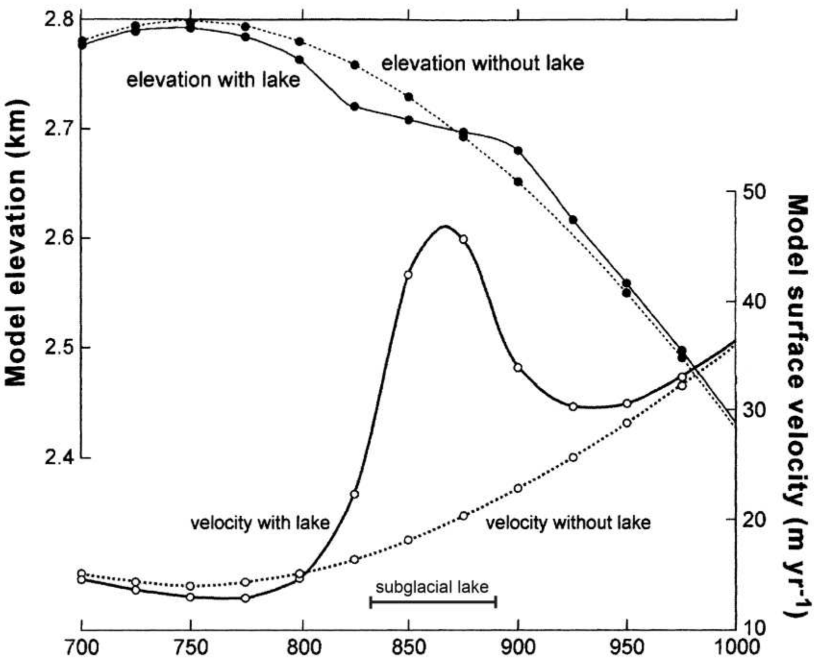


Fig. 6: Cartoon (not drawn to scale) illustrating how changes to the basal thermal regime influence subglacial lake genesis. A: cold-bedded glacier that switches to a polythermal regime in B (due to any of the factors highlighted in the diagram). B: warm-bedded conditions permit subglacial meltwater production and ponding in depressions; and C: if the thermal regime switches back towards cold-bedded conditions subglacial lakes in this zone may freeze and become fossil lakes.

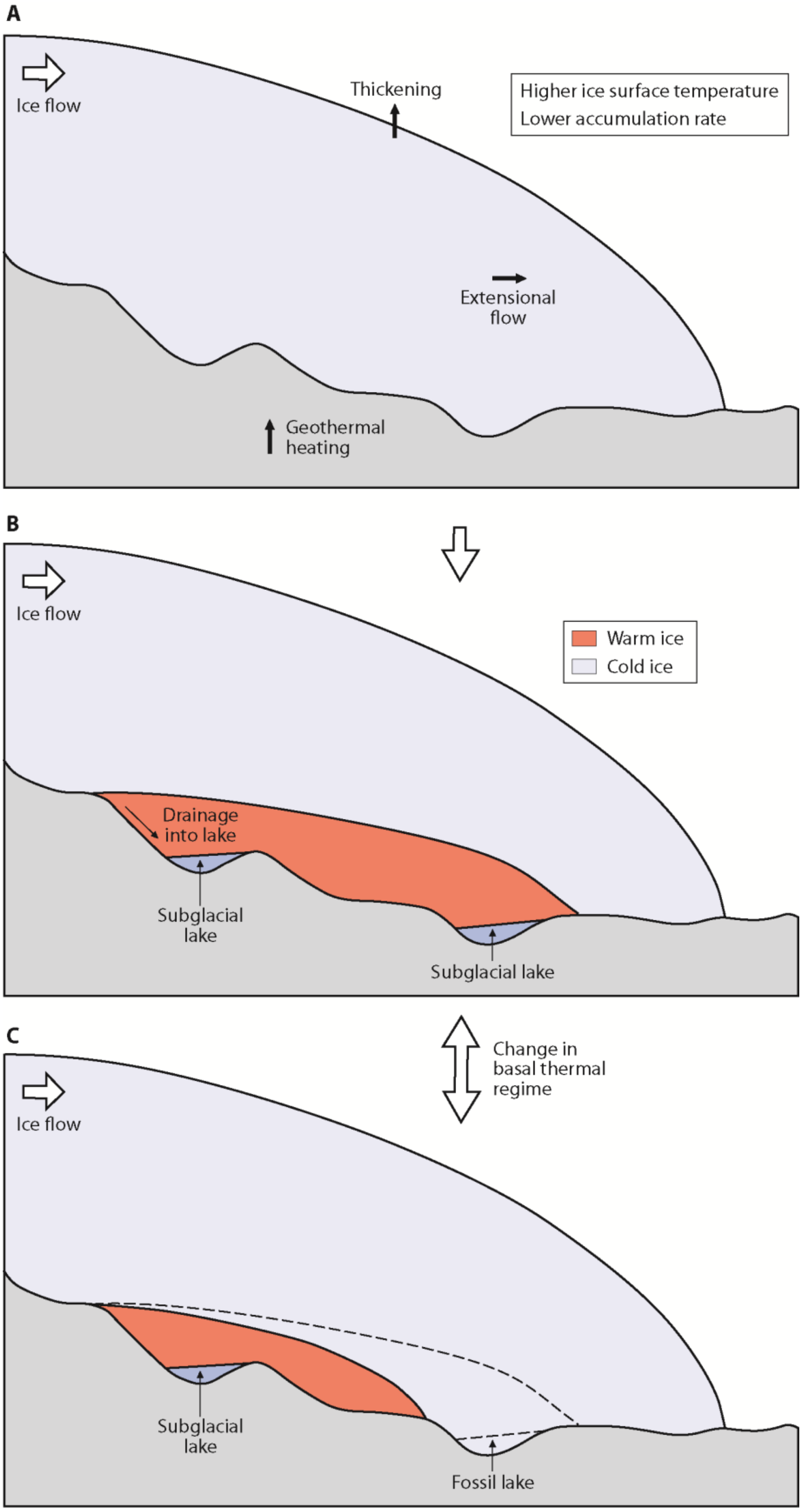


Fig. 7A: The importance of hydraulic conductivity in subglacial lake genesis. Line (1): the bed is able to drain all of the available meltwater generated at the bed (meltwater flux (blue) sits below hydraulic conductivity (black); line (2): if hydraulic conductivity is lowered then regions develop where more water is generated than can be drained (melwater flux exceeds hydraulic conductivity), and therefore excess water will begin to pond (although note this excess water may also drain through channels at the ice-bed interface); and line (3) – the hydraulic conductivity is depressed locally (perhaps due to the influence of permafrost at the ice sheet margin) and even more water can potentially pond at the surface. Other spatially non-uniform influences on hydraulic conductivity (e.g. lithology) could introduce similar effects. B: Cartoon (not drawn to scale) illustrating one example where hydraulic conductivity is lowered and a subglacial lake forms. Prior to advance, the ice mass is unaffected by permafrost and meltwater is free to drain (grey line and arrow, number 1). When the ice mass advances onto permafrost (black line), meltwater drainage through the sediment and at the ice-water interface becomes inhibited (black arrow, number 2). As meltwater production exceeds drainage water will begin to pond (number 2) at the cold-bedded margin.

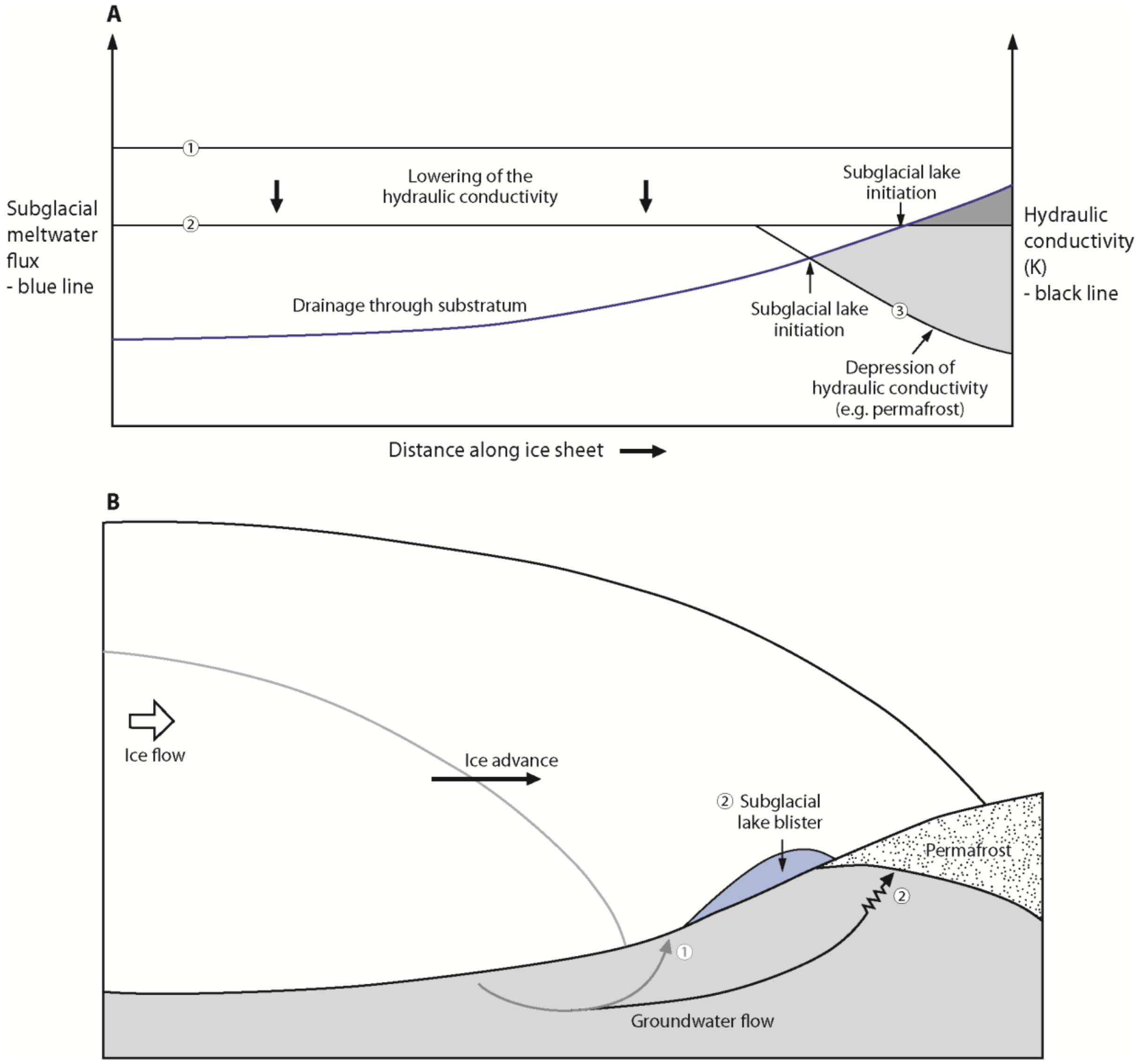


Fig. 8: Basal conditions beneath the Laurentide (a,c) and Fennoscandian (b,d) Ice Sheets. Panels (a) and (b) show mapped landscape or lithology types and (c) and (d) show frozen bed extent. (from Kleman & Hättestrand, 1999). This figure has been reprinted with permission from Nature Publishing Group.

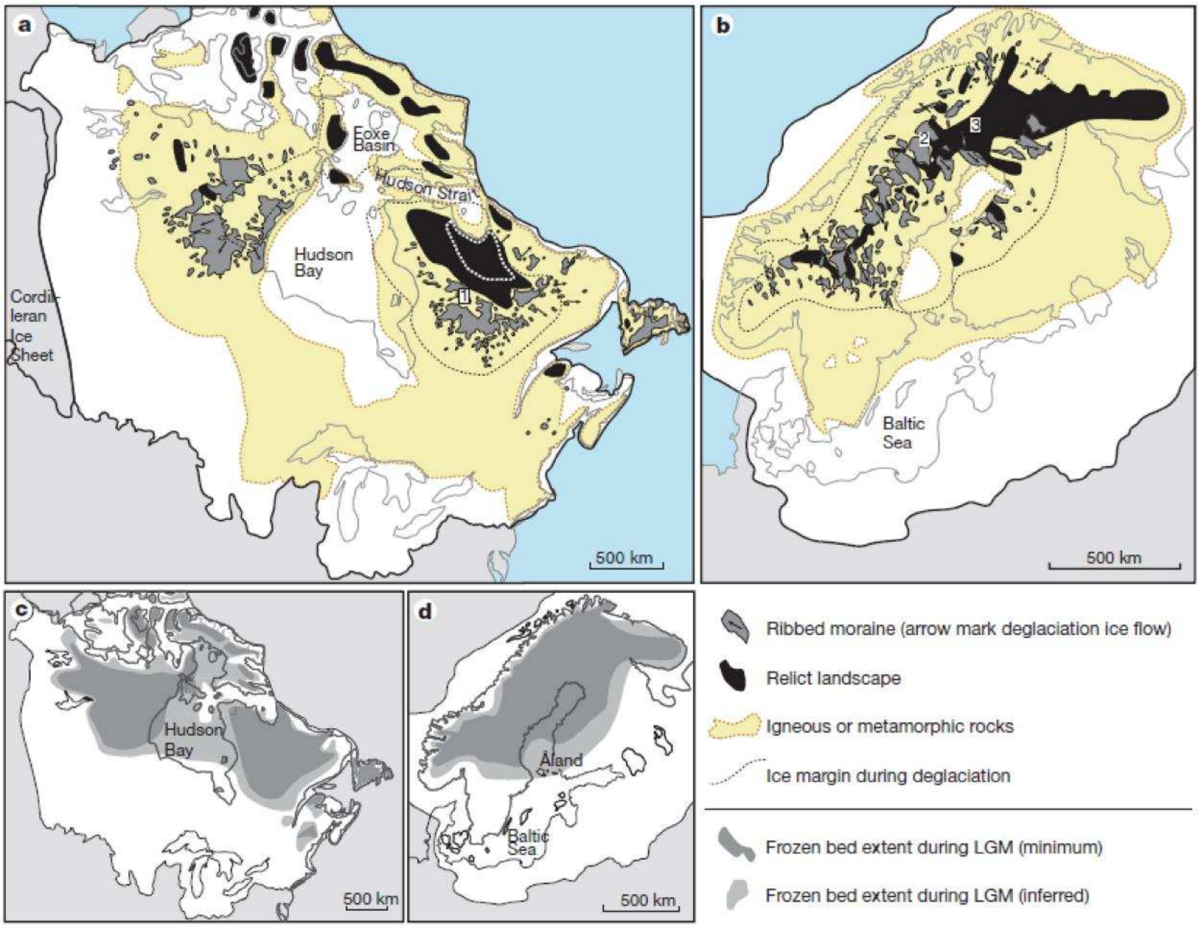


Fig. 9: Conceptual model of subglacial processes and sediment facies in Vostok Subglacial Lake, Antarctica (from Bentley et al. (2011).

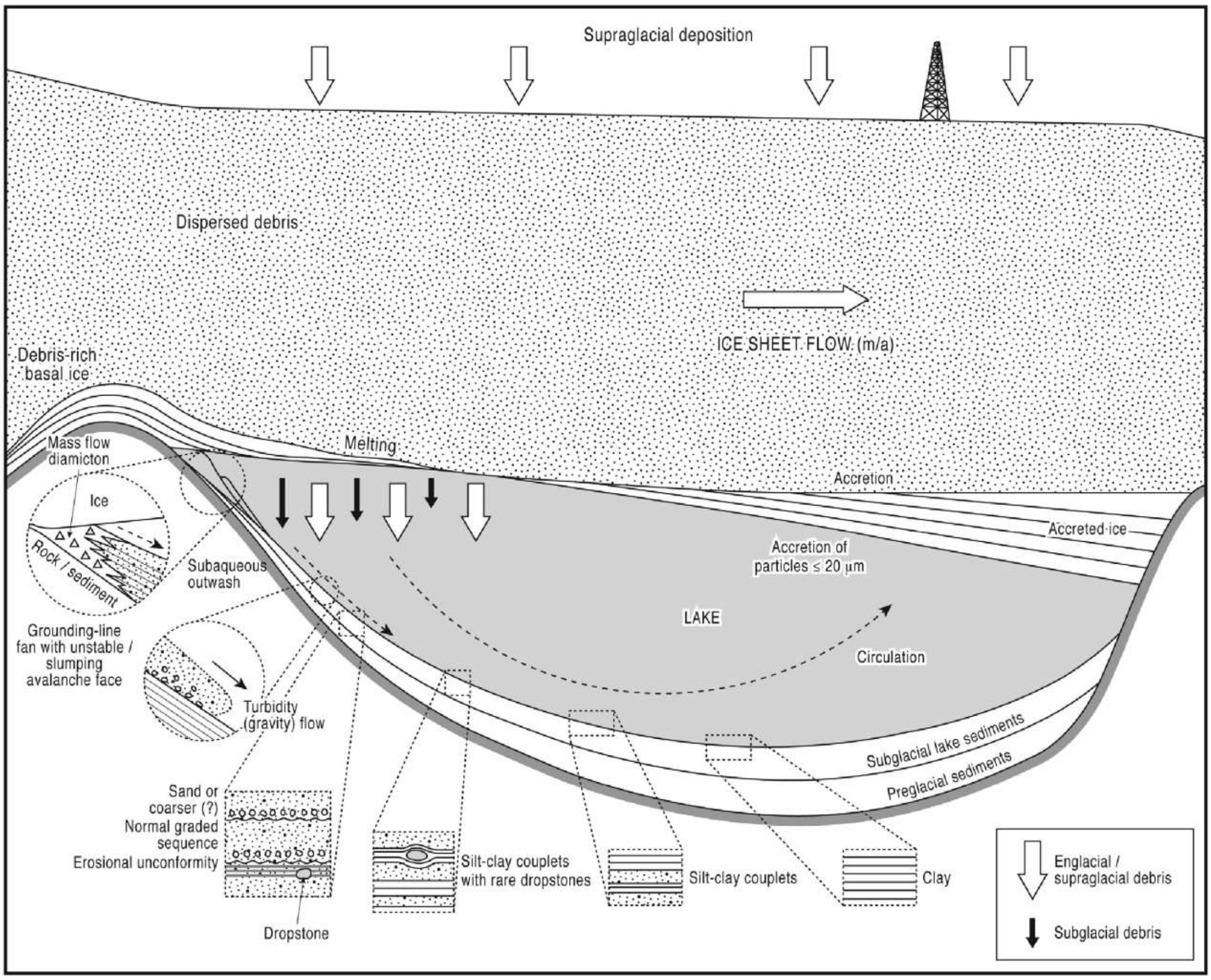


Fig. 10: Subglacial lake analogue of grounding-line processes and associated sedimentary facies and landforms (from Benn, 1996: depositional model for the Strath Bran cross-valley moraine). 1. Deforming till; 2 & 3. diamictic sedimentary-gravity flows; 4. subaqueous outwash fan; 5. sand and gravel foresets; 6. distal facies associated with sand-silt couples (turbidites), laminated muds and diamictons (rain-out deposits), dropstones and dumped clast clusters; 7. dropstones and iceberg dump moraines. These facies that are associated with grounding-line fan and morainal bank formation at the grounding-line of an ice mass may similarly occur at the grounding-line of a subglacial lake. Instead of melt-out from icebergs though, it would be dominated by melt out from the ice ceiling. This figure has been reprinted with permission from John Wiley and Sons.

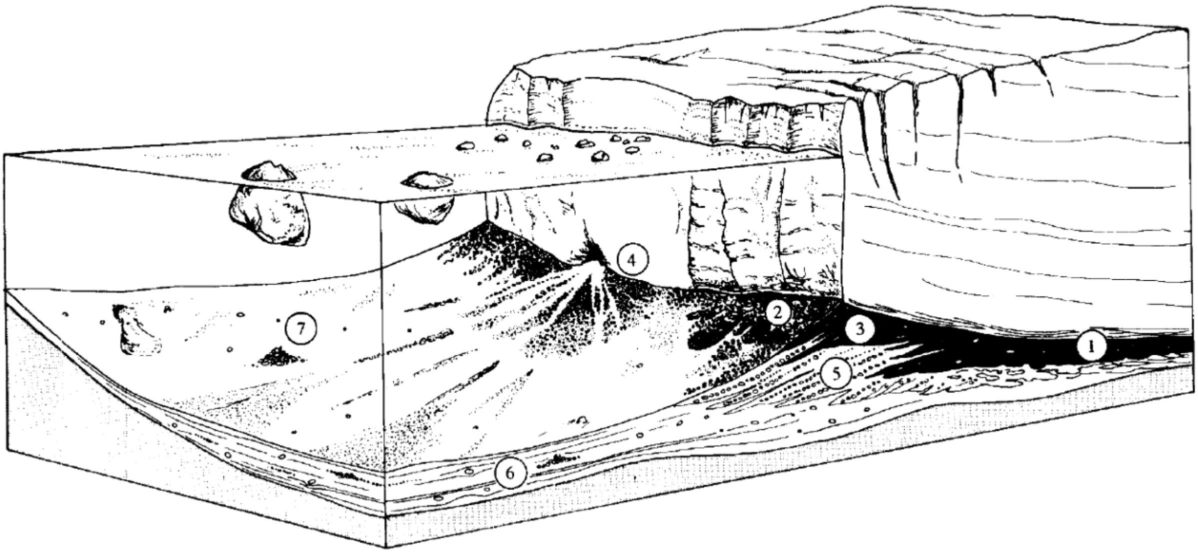
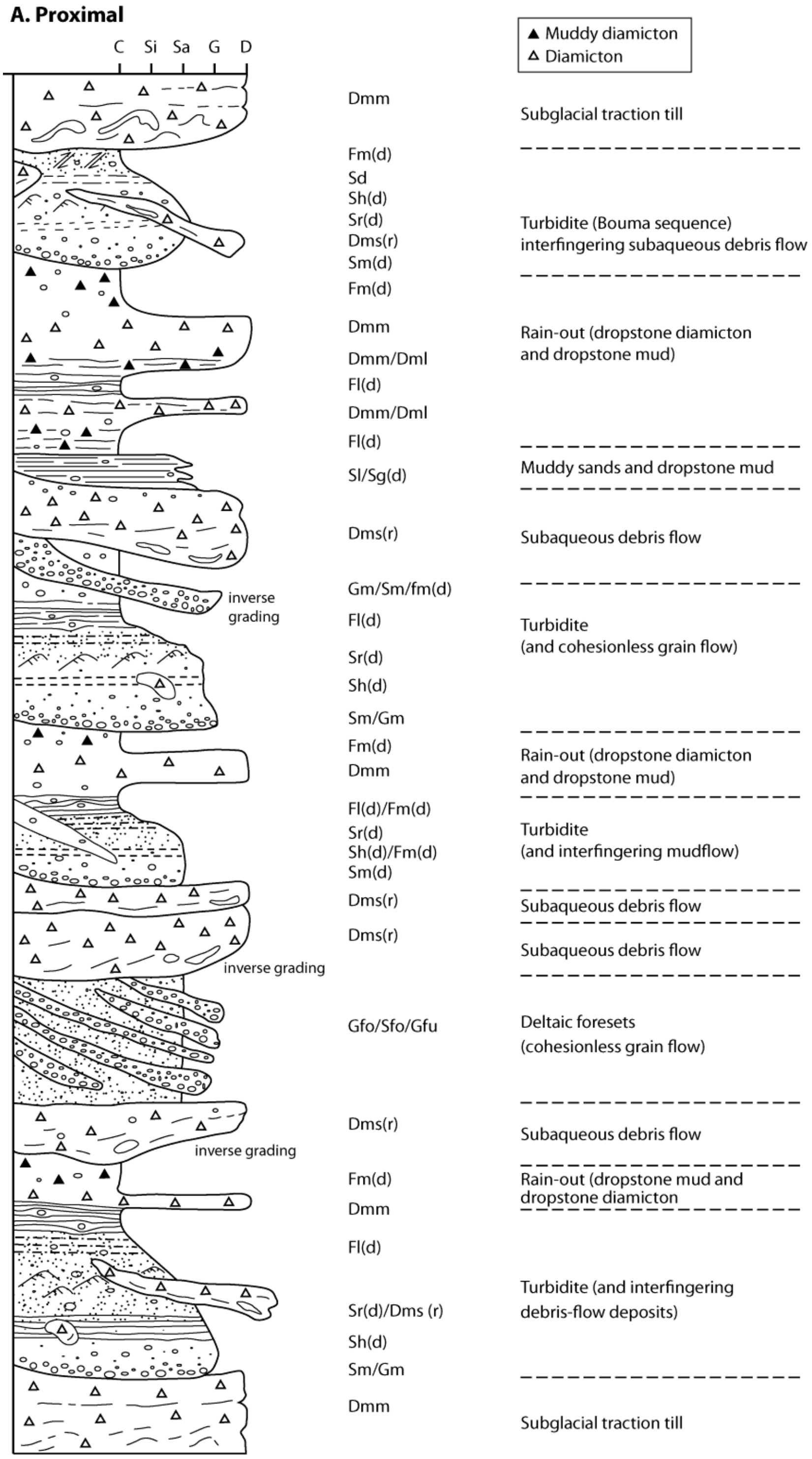
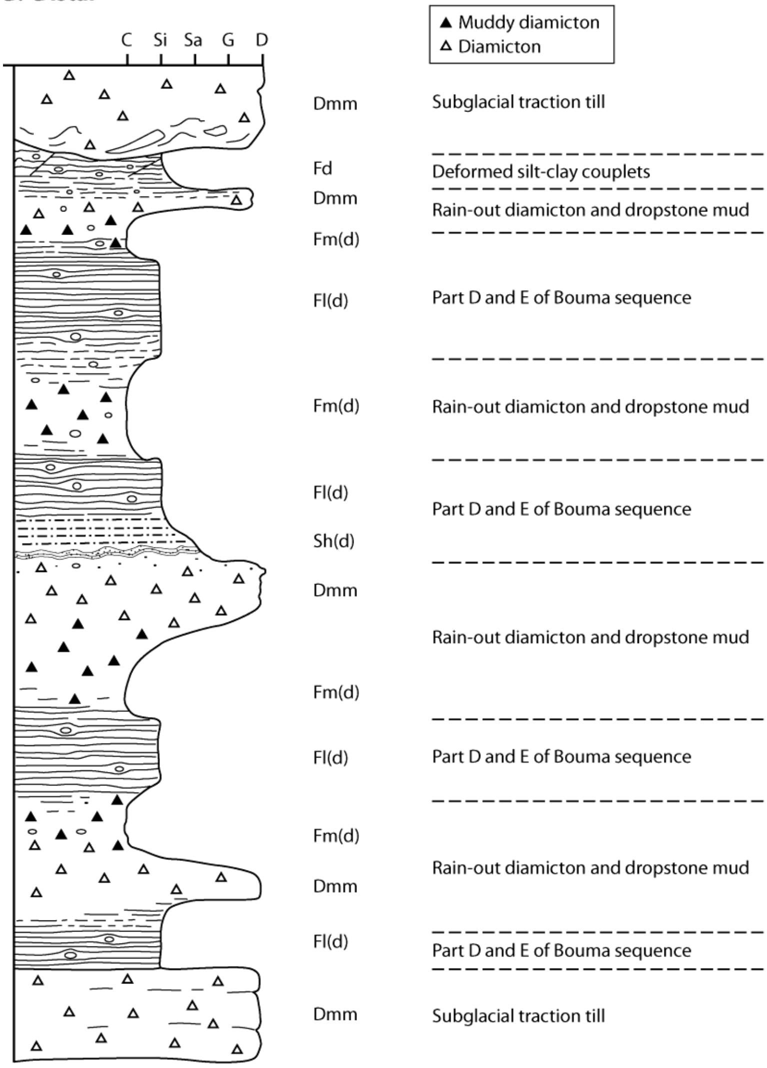


Fig. 11: Hypothetical sedimentary logs illustrating the typical facies and facies associations we would expect to find in subglacial lakes: (a) proximal to the subglacial lake margin and influx portals; and (b) distal to the subglacial lake margin.



B. Distal



▲ Muddy diamicton
△ Diamicton
C Si Sa G D
Dmm
Subglacial traction till
Fd
Deformed silt-clay couplets
Dmm
Rain-out diamicton and dropstone mud
Fm(d)
Fl(d)
Part D and E of Bouma sequence
Fm(d)
Rain-out diamicton and dropstone mud
Fl(d)
Part D and E of Bouma sequence
Sh(d)
Dmm
Rain-out diamicton and dropstone mud
Fm(d)
Fl(d)
Part D and E of Bouma sequence
Fm(d)
Rain-out diamicton and dropstone mud
Dmm
Fl(d)
Part D and E of Bouma sequence
Dmm
Subglacial traction till

Fig. 12: Hypothesised subglacial lake transitions: (a) subglacially initiated subglacial lake; (b) subglacially initiated subglacial lake with periodic full drainage leading to re-grounding and this multiple units of subglacial traction till; (c) capture and evolution of a preglacial to a subglacial lake; (d) evolution of a subglacial lake from a marine setting (note (c) & (d) assume that the ice did not ground). Figs. 13e-f illustrates typical proglacial lake transitions as a comparison (note that in proglacial lakes the sediments will display coarsening and fining sequences due to the advance and retreat of the ice sheet; and in Fig. 3f another ice-marginal unit may also be deposited following lake drainage).

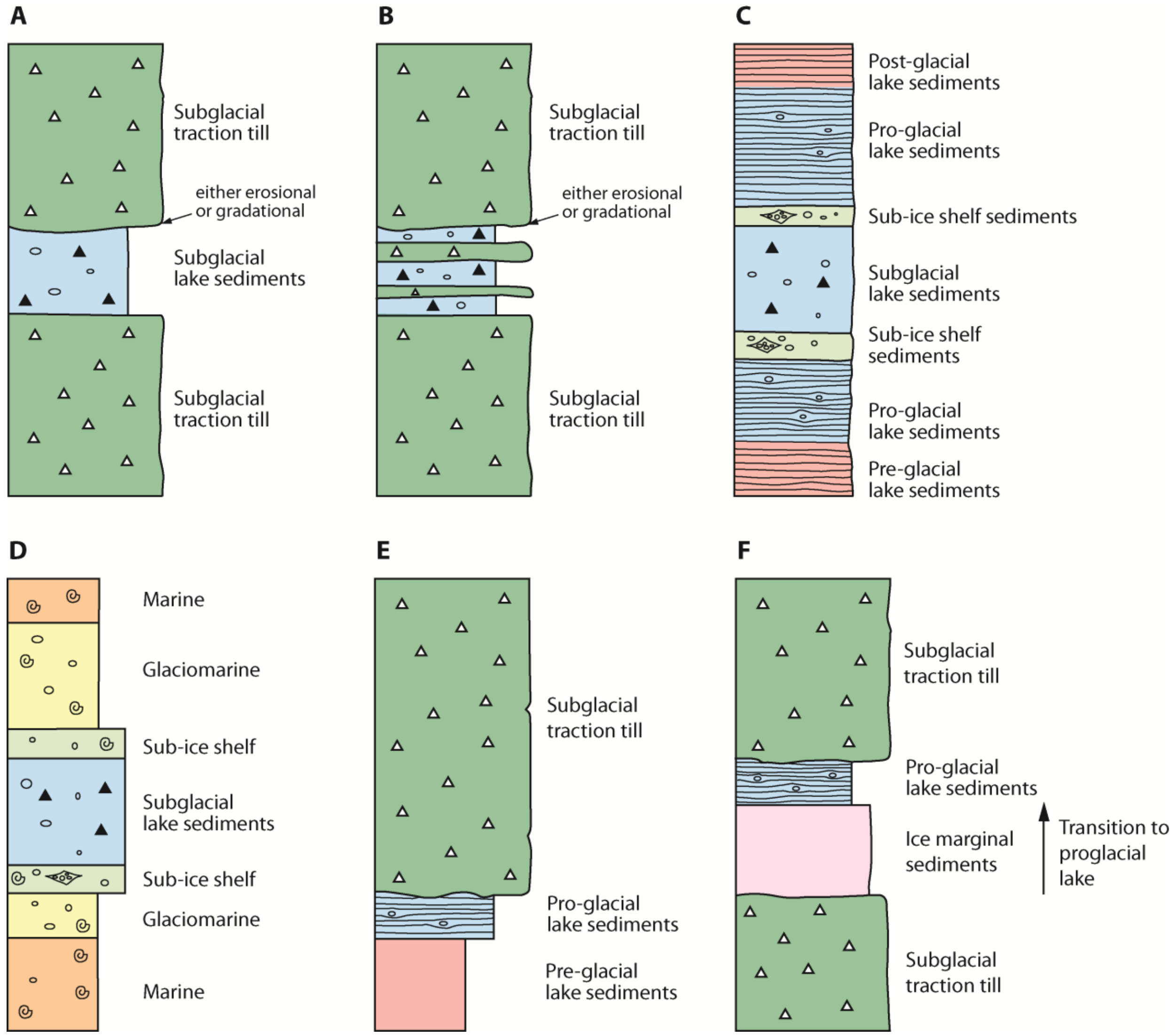


TABLE 1: CLASSIFICATION OF SUBGLACIAL LAKE TYPES (MODIFIED FROM WRIGHT & SIEGERT, 2011)

LAKE TYPE	DESCRIPTION
1. Lakes in the ice sheet interior (along ice-divides)	(a) Deep subglacial basins where ice is thickest and conditions are favourable for basal melting.
	(b) Significant topographic depressions (tectonically controlled).
	(c) Flanks of subglacial mountain ranges – small lakes perched on steep local topography.
2. Lakes associated with the onset of enhanced ice flow	
3. Lakes beneath the trunks of ice streams	Active lake systems, generally small in size and transient in nature.
4. Lakes association with high geothermal heating	Lake blisters rising above the local topography
5. Lakes trapped behind frozen margins	Lake blister rising above the local topography; theoretical lake type

TABLE 2: DIAGNOSTIC CRITERIA FOR THE IDENTIFICATION OF PALAEO-SUBGLACIAL LAKES IN THE GEOLOGICAL RECORD

* The lake-ice interface will prevent the vertical aggradation of subaerial landforms such as deltaic topsets, with deposition instead forced to prograde or fan laterally.

CATEGORY	CRITERIA	ASSOCIATED SUBGLACIAL LAKE PROCESSES OR CONDITIONS	EXAMPLES/ ANALOGUES
1. Sediment facies, associations and architecture	(i) Drapes of massive to weakly stratified dropstone diamicton & dropstone mud. (ii) Turbidites (including Bouma sequences). (iii) Cohesionless debris flows and mudflows (iv) Stacked stratified sequences of conformably interbedded fine-grained sorted sediments & waterlain diamicton. (v) Truncation & deformation of lake sediments & deposition of a capping subglacial traction till at the down-ice flow margin. (vi) Large vertical spatial extent of sediment facies. (vii) Multiple palaeo-current directions.	Melt-out of basal debris-rich ice from the lake ceiling. Subaqueous debris-flows. Subaqueous debris-flows. Complex interactions between melt-out & subaqueous debris-flow processes. Re-grounding of ice at the down-ice flow margin and erosion of sediments by flushing events. Sloping ice-lake interface. Subglacial lakes are hydraulic sinks and therefore sediment will be input from multiple directions.	See Figs. 9, 10 & 11
2. Landform-sediment associations	(i) Grounding-zone wedges* (ii) Grounding-line fan* (iii) Morainal bank* (iv) Deltaic bottomsets and foresets* (v) Foreset units found around the "shoreline" with dip direction towards the centre of the lake, or by concentric rings of morainal banks and/or grounding-line fans (vi) Evidence for lakes in 'odd' locations where topography alone is not able to account for their genesis. (vii) Sediment-landform associations comprising influx points lower than efflux points & distal sediments. (viii) Non-parallel sediment-landform associations. (ix) Low sedimentation rates in subglacial lakes under ice divides & higher component of fine-grained sediment. (x) Subglacial lake sediments not overconsolidated.	Subglacial transport & deposition of till. Delivery of sediment-laden meltwater. Melt-out of sediment close to the subglacial lake margin, efflux of deforming till & sediment-laden meltwater. Cohesionless debris flows, debris-falls & underflows. Subglacial lakes are hydraulic sinks and therefore there will be multiple directions of sediment input. Sloping ice-lake interface Sloping ice-lake interface. Sloping ice-lake interface. Low recharge rates near ice-divides & no seasonal overturning. Confining effect of the ice lid when lowered onto subglacial lake sediments.	See Fig. 10 See Fig. 10
3. Subglacial lake drainage	(i) Association with meltwater channels, tunnel valleys & eskers (i.e. connection at both ends of the subglacial lake. (ii) Connected subglacial lakes via drainage pathways.	Active subglacial lake drainage & recharge. Drainage between subglacial lakes.	See Wingham et al. (2006); Smith et al. (2009)

	(iii) Repeat sequences of high and low energy sediment facies. (iv) Inter-bedded subglacial traction till facies.	Periodic drainage events. Full drainage and recoupling of the ice with the bed.	
4. Subglacial lake transitions	(i) Transition from glacimarine to sub-ice shelf & then into subglacial lake sediments. (ii) Transition from preglacial lake to proglacial lake, sub-ice shelf & then subglacial lake sediments. (iii) Transition from subaerial facies into a subglacial traction till, then subglacial lake sediments, capped by another subglacial traction till. (iv) Graded transition between lake sediments and upper till.	Captured Ice Shelf. Captured Ice Shelf Subglacially triggered subglacial lake. Subaqueous drop-till capping the sequence.	Erlingsson et al. (1994a,b); Figs. 1, 12c,d Figs. 5, 12a
5. Biogeochemical tracers	(i) Relatively high concentrations of Ca, Mg, Na and K, species. (ii) Low Ge:Si ratio. (iii) Ancient microbial biomarkers (iv) Lack of primary sulphides and presence of secondary sulphides. (v) Presence of secondary sulphate minerals such as gypsum and mirabilite. (vi) Micropalaeontology dominated by pychrophilic bacteria. In contrast, organisms associated with aquatic environments (e.g. diatoms & chironomids) and the surrounding catchment (e.g. pollen, spores and plant micro-fossils) would only be present as reworked material or material in transit after burial.	Upper sediments unlikely to be flushed and leached of sediments. Environment is not flushed. Organic matter primarily derived from rock. Microbial weathering and anoxic conditions. Freezing of pore waters during glacial terminations or due to reversal of conditions necessary for lake formation. Isolated, extreme environment.	See Tranter et al. (2005)
6. Dating palaeo-subglacial lake sediments	(i) Anomalous OSL dates that do not conform to ice-marginal retreat histories. (ii) Magneto-stratigraphic dating of palaeo-subglacial lake sediments.	Mechanical re-setting of the luminescence signal by subglacial shearing.	See Swift et al. (2011) See Bentley et al. (2011)
7. Exclusion criteria	(i) Lack of upper shoreface longshore troughs and foreshore sediment facies. (ii) No varves. (iii) No dessication cracks, mud roll-up or trace fossils indicative of drying out.	No wave action or wind generated current activity. No seasonal variability in melt cycles. No periodic drying out.	