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Proglacial lakes: character, behaviour and geological importance

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Abstract

Proglacial lakes are ubiquitous within the Quaternary record and can provide exceptional breadth and depth of palaeoenvironmental information. Present deglaciation is increasing the number and size of proglacial lakes around the world. This study provides a synthesis of knowledge on proglacial lake character and behaviour and critically evaluates the importance of proglacial lakes from a geological perspective. We show how 'ice-marginal' or 'ice-contact' lakes and other distal proglacial lakes can be distinguished from each other by geomorphological, sedimentological, chemical and biological characteristics. The key controls on proglacial lake geomorphology and sedimentology are outlined and discussed. Proglacial lakes can exacerbate mountain glacier and ice sheet margin ablation via mechanical and thermal stresses, but very large lakes can moderate summer air temperatures and relatively retard summer ice ablation. Proglacial lakes interrupt meltwater flux and are very efficient sediment traps. Hydrological routing and consequent geomorphological activity can be radically modified by sudden drainage of proglacial lakes and resultant glacial lake outburst floods; exceptionally large proglacial lake drainages affected global ocean circulation and global climate during the Quaternary. Overall, analyses of proglacial lakes can provide a valuable insight into (i) patterns, character and behaviour of mountain glaciers, ice sheets and glaciations, and (ii) the impacts of past, present and future deglaciation.

Key words: glacial lake; ice-dammed; moraine-dammed; ice-marginal; ice-contact; deglaciation

1. Introduction and rationale

This study provides a synthesis of knowledge on proglacial lakes with an emphasis on local, regional and global effects on geological systems. Proglacial lakes are masses of water impounded at the edge of a glacier or at the margin of an ice sheet. The term ‘proglacial lake’ has been used to refer to ice-contact or ice-marginal lakes, which are physically attached to an ice margin, as well as lakes detached from, or immediately beyond, a contemporary ice margin. In this paper, the term proglacial lake therefore includes all lakes that are or have been directly influenced by (i) a glacier ice margin or (ii) subaerial glacial meltwater.

Proglacial lakes are very significant within the Quaternary record. The character and behaviour of proglacial lakes is intrinsically linked to climate though the surface energy balance and to wider geological systems through glacier dynamics, glacial meltwater and sediment fluxes (e.g. Larsen et al., 2011). Proglacial lakes can affect the stability of ice-sheet margins and mountain glaciers and can disengage glacier behaviour from climatic perturbations. Proglacial lakes interrupt the delivery of meltwater and sediment to proglacial zones and ultimately to oceans. Sedimentation within proglacial lakes is an exceptionally important geochronological archive recording short-term inter-seasonal patterns, inter-annual patterns, and long-term patterns of glacier-derived meltwater fluctuation. Thus proglacial lake records are used by proxy to reconstruct glacier mass balance (e.g. Phillips et al., 2006; Larsen et al., 2011), although that can be highly variable between catchments and through time (e.g. Carrivick and Chase, 2011). In practical terms, knowledge of proglacial lakes has application in assessments of sites vulnerable to glacier outburst floods, aquatic ecosystem monitoring and hydro-electric power generation, for example. Therefore, it is now critically important to (i) better understand the Quaternary record by using recent quantitative studies of modern analogues, and (ii) assess the potential geological importance of the observed evolution and development of proglacial lakes around the world. The major motivations for this study are that;

(i) Proglacial lakes are pervasive within worldwide geological records of Quaternary deglaciation (e.g. Teller, 2001; Jansson, 2003; Mangerud et al., 2004; Larsen et al., 2006; Livingstone et al., 2010; Fiore et al., 2011; Murton and Murton, 2012) and in most cases offer exceptional breadth and depth of palaeo-environmental information (e.g. Thomas and Briner, 2009).

(ii) Present deglaciation is producing an increased number and size of proglacial lakes around the world; for example, as recently documented in the European Alps (e.g. Paul et al., 2007), the

Caucasus (e.g. Stokes et al., 2007), Iceland (e.g. Schomacker, 2010), South America (e.g. Loriaux and Cassasa, 2012; Thompson et al., 2011) and across the Himalaya (e.g. Gardelle et al., 2011) and specifically within the Mt Everest region (Tartari et al., 2008), in Bhutan (Komori, 2008) and Tibet (Chen et al., 2007; Wang et al., 2011). It is very important to understand the effects that these lakes could have on modern and future geological and environmental systems.

(iii) Glacial lake outburst floods (GLOFS), which are a type of jökulhlaup, are ubiquitous within the Quaternary record of deglaciation (e.g. Baker, 2007), and also have a modern geological imprint and are modern natural hazards.

The aims of this study are to provide a review of the character and behaviour of proglacial lakes, emphasising recent developments in knowledge and understanding and critically evaluating the importance of proglacial lakes from a geological perspective. Specifically, we focus on; (i) recent advances in understanding the formation and evolution of proglacial lakes; (ii) criteria for distinguishing proglacial lakes from other freshwater lakes, from marine glacier margins and from subglacial lakes in the Quaternary record; and (iii) identification of the linkages between proglacial lakes and glacier dynamics, meltwater and sediment fluxes and weather and climate, as determined from the Quaternary record and from modern measurements.

2. Proglacial lake character and behaviour

In this section, key attributes of proglacial lakes that shape their geological significance are discussed. Consideration is given to past and present locations and distributions of proglacial lakes before a discussion of controls on proglacial lake formation and evolution. Physical characteristics of proglacial lakes are subsequently described with emphasis on key geomorphological and sedimentary characteristics that are of use for (i) determining past glacial and hence palaeo-environmental patterns, and (ii) distinguishing proglacial lakes from other freshwater lakes, from marine glacier margins and from subglacial lakes in the Quaternary record. Whilst proglacial lakes occur worldwide, we note that the literature on proglacial lakes is dominated by studies within (temperate) high mountain and high latitude regions.

2a. Location and distribution of proglacial lakes

Proglacial lakes can be situated in front and at the sides of mountain glaciers, at the front and sides of ice sheet outlet glaciers and at the edge of nunataks and are most commonly dammed by ice, bedrock, moraine debris or landslide debris (Fig. 1). Less commonly, proglacial lakes can be dammed by other sediments; for example as a glacier retreats and thins behind the ice-contact slope of a glacifluvial fan or apron (Carrivick and Russell, 2013). Previous reviews have compiled data on natural dams (e.g. Costa and Schuster, 1988) and have discussed the formation and catastrophic drainage of ice-dammed lakes (e.g. Tweed and Russell, 1999), moraine-dammed lakes (e.g. Richardson & Reynolds, 2000) and landslide dams (e.g. Korup, 2002). Proglacial lakes exist in all currently glaciated regions of the world and the legacy of proglacial lakes is frequently evident in formerly glaciated areas. Table 1 provides examples of modern proglacial lakes from around the world. Table 2 provides examples of major palaeo-proglacial lakes, particularly those lakes associated with Late Pleistocene ice sheet deglaciation from which sudden drainage generated some of the largest floods on Earth and affected ocean currents and offshore sediment fluxes.

2b. Proglacial lake formation

Proglacial lakes can be impounded by ice, moraine, landslide debris or bedrock (Costa and Schuster, 1988). The formation, evolution and persistence of proglacial lakes are strongly linked to glacier dynamics and the nature of the surrounding environment. There are clear associations between both of these factors and changing climatic conditions, but factors independent of climate can influence proglacial lake behaviour. The type of dam has implications for lake character, lake evolution and lake drainage. Failure or overtopping of the impounding material frequently leads to glacier lake outburst floods (GLOFS) or jökulhlaups (e.g. Walder and Costa, 1996; Tweed and Russell, 1999; Richardson and Reynolds, 2000; Clague and Evans, 2000; Nayar, 2009), which can be powerful agents of landscape change through erosion and sediment deposition (e.g. Carling, 1996; Carling et al., 2002; Carrivick and Twigg, 2005, Carrivick et al., 2004a, b, 2007; Carrivick, 2007, Alho et al., 2007, Russell et al., 2006).

Formation of ice-dammed lakes is a consequence of local topography and favourable hydraulic pressure gradients (Tweed and Russell, 1999; Roberts, 2005) and is usually associated with ice margin advance or thickening, which permits the impounding of water. In most settings, ice-dammed lake formation is a gradual, quasi-periodic to episodic process, linked to glacier mass balance and, ultimately, climate forcing (Evans and Clague, 1994). However, sudden glacial advance during surging can block drainage channels and create ice-dammed lakes independently of climate (e.g. Clarke and Mathews, 1981; Mayo, 1989). The rate of lake formation may be linked to glacier surface gradient and velocity (Quincey et al., 2007) and the behaviour of ice-dammed lakes, along with their drainage triggers, is intimately connected to and influenced by, the nature of the damming ice.

Formation of moraine-dammed lakes is quasi-periodic and is usually associated with periods of glacier retreat or down-wasting following previous ice advance (Clague and Evans, 2000; Korup and Tweed, 2007). However, the development of large terminal moraines can result from a variety of processes some of which are independent of climate; for example changes in supraglacial debris fluxes controlled by sediment input from valley sides (Benn and Owen, 2002). Moraine-dammed lakes expand into topographic lows formerly occupied by ice or major meltwater channels and lakes can grow rapidly, fed by both precipitation and glacial meltwater (Yao et al., 2010; Wang et al. 2011).Moraine-dammed lakes often develop on debris-covered glaciers as debris-charged glacier snouts can become separated from more rapidly ablating ice up-glacier and eventually stagnate; in these circumstances, ice-cored and non-ice cored moraine then serves as an effective barrier to meltwater runoff leading to proglacial lake development (e.g. Ageta et al., 2000; Stokes et al., 2007). Moraine-dammed lakes also grow in depressions formed by large masses of buried ice (e.g. Howarth and Price, 1969; Kääb and Haeberli, 2001). In many settings, moraine-dammed lakes gradually evolve to become separated from any ice mass and can persist beyond regional deglaciation (Krivonogov et al., 2005; Pasquini et al., 2008).

Formation of landslide-dammed lakes in glacial environments is stochastic and is usually associated with slope de-buttressing on glacier retreat (Korup, 2002; Korup and Tweed, 2007). Increases in slope angle and relief coupled with the loss of internal friction and cohesion of slope materials prepare slopes for failure. River-damming landslides can be triggered by rainstorms and snowmelt leading to elevated pore-water pressures, by seismic vibration or by undercutting (Korup, 2002). Repeated landslides at particular sites can form stacked dams (e.g. Chigira et al., 2003). Landslide-dammed lakes are often ephemeral as firstly, the damming material frequently lacks cohesion and secondly, they generally form in steep mountain terrain with high denudation rates and consequently lakes decrease in size due to sediment infill (Hicks et al., 1990; Korup et al., 2006). Bedrock-dammed lakes require a glacial over-deepening (Cook and Swift, 2012) in which to form and are therefore common in mountainous regions, where reigels form effective dams, on ice cap fringes and in ice sheet outlet troughs. In some locations, proglacial lakes are impounded by a combination of dam types, or evolve through a sequence of different lake/dam configurations (see section 2c).

2c. Proglacial lake evolution

Once formed, proglacial lakes can be persistent elements of a landscape, but they can also be transitory and dynamic. The type of dam and the environmental setting exert key controls on proglacial lake growth, contraction, filling, emptying and persistence, all of which are important for understanding the landform and sedimentary record, for the reconstruction of past environmental conditions and for the prediction of future environmental change.

Evolution of ice-contact lakes is strongly influenced by their proximity to ice and the dynamic nature of their relationship with it. Ice-dammed lakes fill and empty, their drainage frequently resulting in jökulhlaups (Walder and Costa, 1996; Tweed and Russell, 1999; Russell et al., 2011). Evans and Clague (1994) first described the 'jökulhlaup cycle' and further discussions of lake formation, drainage, re-filling and re-emptying in response to glacier fluctuations and associated ice-dammed lake drainage trigger mechanisms have been provided by Tweed and Russell (1999). More recently, rates of lake level rise and lake water temperature as a control on a jökulhlaup cycle have been modelled by Ng and Liu (2009). The magnitude and frequency of jökulhlaups from ice-dammed lake drainage is connected to different phases of the jökulhlaup cycle. Thinner ice dams consequent on glacier retreat impound progressively decreasing levels of lake water, causing more frequent low magnitude floods (e.g. Roberts et al., 2005). Conversely, thickening ice dams created by glacier advance have the capacity to impound larger bodies of water, generating less frequent jökulhlaups of greater magnitude (e.g. Clague and Evans, 1997). Cycles of filling and emptying can persist on annual (e.g. Mathews and Clague, 1993; Walder et al., 2006), one to five years (e.g. Sturm and Benson, 1985), decadal (e.g. Waitt, 1985) and centennial (e.g. Teller and Leverington, 2004) timescales and have implications for rates of erosion, sediment transfer and deposition. Proglacial lake cycles are discussed more fully in section 2e.

Moraine-dammed lakes (e.g. Fig. 2A) evolve as a consequence of their environmental setting, particularly their proximity to ice. Identification and monitoring of moraine-dammed lakes in mountainous terrain has become increasingly important chiefly due to the occurrence of GLOFs as moraine-dammed lakes expand on glacial retreat (Vuichard and Zimmerman, 1987; Watanabe and Rothacher, 1996; Ageta et al., 2000; Richardson and Reynolds, 2000; Sakai et al., 2000; Aniya and Naruse, 2001; Kattelman, 2003; Shresta and Shresta, 2004; Harrison et al., 2006; Quincey et al., 2007; Bajracharya et al., 2007; Komori, 2008). There has been widespread documentation of the evolution of moraine-dammed lakes associated with debris-covered glaciers (e.g. Bolch et al., 2008; Wang et al., 2011); the enlargement and deepening of moraine-dammed lakes on and adjacent to debris-charged glacier snouts is linked to the melting and subsidence of dead ice beneath and adjacent to them (e.g. Stokes et al., 2007; Kääb and Haeberli, 2001). These environments are often highly dynamic and the changes within them can create complex and interdependent processes and impacts; for example, an outburst flood from a moraine-dammed lake at Chamdo in the Boschula mountain range of Tibet triggered a large landslide that led to the formation of a landslide-dammed lake (Wang et al., 2011), which in turn could generate further floods.

Proglacial lakes can expand, coalesce, decrease in size (e.g. Fig. 2B), disappear, or become detached from an ice margin (Carrivick, 2011; Fig. 3). The evolution and persistence of proglacial lakes exhibit strong links to glacier behaviour and therefore to climate change. When a glacier is advancing, it is coupled with the sandur and meltwater can freely disperse onto it. On glacier recession, an over-deepened glacier basin, either in bedrock (Fig. 2C) or within sediments, is frequently revealed, creating a depression in the proglacial zone which forms a gap between the ice front and proglacial rivers (e.g. Marren, 2002). If a glacier continues to retreat, ice eventually decouples from the sandur (e.g. Fig. 2D), resulting in the ponding of meltwater and the trapping and storage of sediment within proglacial lakes (Syverson, 1998; Schomacker and Kjaer, 2008); for example, Liermann et al. (2012) report that 85 % of sediment from the Bødalen catchment in western Norway is retained by a proglacial lake. Meltwater exits a proglacial lake system via a few outlets onto a sandur as glacier retreat continues (e.g. Gomez et al., 2002; Fig. 2E). Glacier retreat into over-deepened glacier basins often accelerates due to enhanced calving, which in turn causes lake expansion and further calving and retreat. Such positive feedbacks between proglacial lakes and ice dynamics are complicated and have been suggested to cause (i) rapid disintegration of ice sheet margins, or (ii) rapid advance or surges, by encouraging low effective pressures at an ice margin (e.g. Fyfe, 1990; Stokes and Clark, 2003). Proglacial lakes usually grow rapidly following the transition to the calving phase, as this increases meltwater input over that of glacier ice melting alone (Kirkbride, 1993; Röhl, 2006; Fujita et al, 2009; Sakai et al., 2009; Schomacker, 2010). It is possible to infer meltwater generation volumes and rates from proglacial lake growth (e.g. Teller, 1995; Clark et al., 1999; Marshall and Clarke, 1999), which, alongside their hazard potential, strengthens the case for monitoring of proglacial lakes.

The worldwide retreat of many ice sheet margins and glacier termini is currently giving rise to the formation of new proglacial lakes and the enlargement of existing proglacial lakes (e.g. references in Table 3; Fig. 3). Recent formation and expansion of proglacial lakes is particularly well-documented in the Himalaya and other high mountain regions of the world (Table 3), but to-date has not been fully recognised at ice sheet margins (Carrivick, 2011; Fig. 3)

Glacier fluctuations exert a strong influence on the development and evolution of proglacial lakes. Cycles of draining and refilling of both ice-dammed and moraine-dammed lakes are profound controls on their sedimentary impact. As glaciers retreat, ice-dammed lakes may drain completely and fail to reform as a consequence of thinning ice dams, but moraine-dammed lakes begin to develop and expand, particularly at the frontal margins of glaciers and ice sheets. It is possible for ice-dammed lakes to evolve into moraine-dammed lakes, as they gradually separate from glacial ice. Some such lakes may survive over centennial or millennial timescales, depending on the coherence of the moraine dam; some modern proglacial lakes are the remnants of ancient lakes (e.g. Krivonogov et al., 2005).

2d. Multiple lake development

There are marked differences in the evolution, behaviour and geological significance of individual proglacial lakes as compared to the development of a suite of lakes along an ice-margin and their eventual coalescence. The Quaternary landform and sedimentary record provides numerous examples of the development and drainage of complex proglacial lake systems associated with the advance and retreat of large ice masses, especially during the late Pleistocene. Several investigations have revealed evidence of multiple lake growth and coalescence, the development of lake drainage routeways and their re-use and subsequent abandonment as these lakes evolved and drained. For example, the complex history and impacts of proglacial lake evolution and drainage associated with the deglaciation of the Cordilleran ice sheet; e.g. Glacial Lake Missoula and Glacial Lake Columbia; Fig. 4) and the Laurentide ice sheet; e.g. Glacial Lake Agassiz; Fig. 4) have been comprehensively researched (e.g. Baker and Bunker, 1985; Teller, 1987; O'Connor and Baker, 1992; Benito and O'Connor, 1993; Kehew and Teller, 1994; Teller et al., 2002; Clarke et al., 2003; Jansson, 2003). The Quaternary ice sheet margins of northern Eurasia and central Asia dammed huge proglacial lakes (Fig.4); Baker (2007) highlights the glacially diverted drainage system of central Asia in which lakes were connected by a series of spillways. Larsen et al. (2006) provide a synthesis of the glacier and lake history of north-west Russia, which links glacier dynamics to the evolution and behaviour of ice-dammed lakes and reveals a complex lake history with ice-dammed lakes forming at different locations along ice margins at different times. Rudoy and Baker (1993), Baker et al. (1993) Carling et al., (2002), Herget (2005) and Komatsu et al., (2009) present geomorphological and sedimentological evidence for cataclysmic floods in the Altai Mountains of southern Siberia, due to the emptying of multiple palaeolakes impounded by ice, moraine and lava flows. Current glacier and ice sheet recession is also giving rise to complex patterns of proglacial lake development (e.g. Fig. 3) and behaviour as lakes grow, coalesce and drain (e.g. Ageta et al, 2000; Stokes at al, 2007; Carrivick, 2011). This complex behaviour of proglacial lakes in space and time has a range of impacts on the landform and sedimentary record, which will be discussed in sections 2f v, vii.

2e. Proglacial lake cycles

Observation and measurement of proglacial lake cycles is very rare. In general, research into proglacial lake evolution takes one of two approaches. The first approach is to model the behaviour of modern lakes using historical documentation of events and climatic conditions as a basis (e.g. Ng et al., 2007). This approach has severe limitations, principally because the historical recorded observations upon which these models are often based are limited to the last fifty years or less and are often incomplete. The second approach uses the Quaternary record and is usually based on geomorphological indicators and sedimentary evidence (e.g. Fiore et al., 2011), but also employs geo-chronological data, such as the analysis of tree-rings (e.g. Capps et al., 2011) and pollen or other organics within strata immediately above or below glacial units (e.g. Astakhov, 2008) to produce models of lake evolution (e.g. Etienne et al., 2006). However, this approach is hampered by a lack of modern analogue studies.

2f. Physical characteristics of proglacial lakes

Observations and measurements of geomorphology and sedimentology permit process-based interpretations and reconstructions of past environments and processes. Therefore understanding of the physical characteristics of proglacial lakes is crucial for targeted sampling for both (i) direct reconstruction of proglacial lake processes; and by proxy glacial activity, and (ii) recognising proglacial lakes in the Quaternary record and thereby indirect reconstruction of past environmental conditions. Furthermore, proglacial lake sediments and simultaneously deposited organic material provide some of the best uninterrupted geochronological archives; alongside tree rings, speleothems and ice cores. Indeed, the stratigraphic record of proglacial lakes is one of the most continuous continental records of climate change through the Holocene to the present day (e.g. Hicks et al., 1990; Leonard and Reasoner, 1999; Charlet et al., 2008; Rayburn et al., 2011).

'Ice-marginal' or 'ice-contact' lakes and distal proglacial lakes are distinct from each other in terms of both geomorphology and sedimentology. Ice-marginal lakes have a unique lacustrine character that reflects both a close proximity to an ice margin and also the rapidly changing nature of this situation. Teller (1987) and more recently Rubensdotter and Rosquist (2009) have summarised the main factors controlling the geomorphological and sedimentary characteristics of ice-contact lakes as; location of a glacier margin, elevation and topography of surrounding landscape, location and elevation of the lake overflow channel, and volume and nature of sediment supply. Each of these factors is discussed below for ice-contact and for distal proglacial lakes by considering the key controls and effects of water depth, bathymetry, hydrography / limnology and specifically thermal regime and suspended sediment density stratification, chemical and biological characteristics and sedimentation. It is important to recognise the importance of spatio-temporal scale (e.g. lake area, lake depth, catchment size, time period) when considering proglacial processes.

i) Water depth

Water depth in ice-marginal lakes is prone to sudden changes due to (i) opening and closing of outlets by glacier margin position changes and by changes in glacier thickness, (ii) periodic or episodic outburst floods or 'jökulhlaups', and iii) recharge events. Fluctuating water level is one key characteristic of ice-marginal lakes that discriminates them from subglacial lakes (Livingstone et al., 2012). Fluctuating proglacial lake water levels has impacts on (i) glacier character and dynamics (Table 4), and (ii) landform development (e.g. Sturm, 1986; Winsemann et al., 2011), and thus has important implications for environmental reconstruction. Water depth rather than climate appears to have controlled moraine systems deposited along the Wisconsian ice margin (Hillaire-Marcel et al., 1981; Vincent, 1989). Water depth has also been inferred to control subglacial drainage configuration and hence the style and pattern of sedimentation along the ice/lake interface (e.g. Fyfe, 1990; Winsemann et al., 2011). Specifically, Fyfe (1990) reports that large deltas were the product of conduit-focused sedimentation, whilst lower, narrower coalescing fans of finer material formed from a distributed drainage system. Furthermore, Fyfe (1990) suggested that subglacial conduit systems are unstable where marginal water depths are greatest, thereby favouring development of a distributed drainage system. Perhaps it is important to note that water depth does not control sediment thickness (Gilbert, 2003) but that it is critical for preserving fine-laminated sediments (something like mm-scale) versus more massive and mixed sediments. In this latter respect, water depth must be carefully considered in the context of numerous other controls when discussing sedimentation; water depth controls oxygen and thermal stratification that in turn affect the mixing potential of the water and sediments. Other factors to consider as a control on sediment preservation include alkalinity, dissolved oxygen, diagenesis and fetch.

ii) Bathymetry

Bathymetry strongly determines sediment dispersal within proglacial lakes. Turbid density/gravity currents or 'underflows', which are particularly prominent in ice-contact lakes, are directed by into low points on a lake floor (Gustavson, 1975; Smith et al., 1982; Ashley, 1995). Common bathymetric elements include submerged bedrock hills and depositional landforms such as moraines, subaqueous channels and enclosed basins that are created by erosional over-deepening or uneven sedimentation. Knowledge of lake bathymetry is also important for geological investigations because submerged topographic barriers or 'sills' can act as barriers to the transfer of water and sediment and to the transport of icebergs. Ice-rafted debris (IRD) can become grounded in water that is shallower than the critical depth required for flotation.

The vast majority of proglacial lake basins have flat floors produced by voluminous sedimentation. Terrain that was previously glacially subdued and smoothed can become draped and obscured by thin beds and laminae of silts and clays. Pleistocene proglacial lake basins, such as those in the Hudson Bay area and on the Canadian Shield (Dredge and Cowan, 1989), can be recognised today as very extensive areas of peatlands that have accumulated due to relatively impermeable sediments and poor drainage. These palaeo-lake areas are usually further distinguished by encircling wave-cut cliffs or by coarser sediment that was deposited in shallower water, such as beaches and lags of wave-washed sediment (Teller, 2003).

On a large scale, bathymetry of proglacial lakes is dependent on regional topography. Very large Pleistocene proglacial lakes across North America developed at least partly because the continental land surface slope trended northwards towards the Arctic Ocean (Teller, 1987). This slope, which was inverse to the direction of ice motion, was accentuated by isostatic depression and thus a considerable accommodation space for meltwater was created in the landscape. To quantify this effect, in North America and also in Scandinavia the isostatic depression from south to north as measured by (now deformed) lake shorelines or 'beaches' is ~ 200 m (e.g. Lemmen et al., 1994). Today, many proglacial lakes are developing along the western margins of the Greenland ice sheet not just due to enhanced meltwater generation, but also because accommodation space for meltwater to accumulate in the landscape is created as the ice sheet retreats on an adverse bed slope.

iii) Hydrography / Limnology

Very little hydrographic / limnological data exist for modern proglacial and particularly for ice-contact lakes; work by Francus et al (2008) and Schiefer and Gilbert (2008) are notable exceptions. In general, proglacial lakes experience pronounced, frequent and regular fluctuations in thermal (e.g. Richards et al., 2012) and sediment load inputs and generally develop some sort of density stratification. Density stratification makes proglacial lakes, especially ice-contact lakes, distinct from other 'river-fed' lakes (e.g. Gustavson, 1975; Weirich, 1985, 1986) and distinct from subglacial lakes (Livingstone et al., 2012). Density stratification is important geologically because it controls circulation (currents) of water and thus determines sediment distribution within the lake and the overall architecture and detailed structure of sedimentary deposits (Fig. 5). Both density stratification and circulation patterns are controlled by specific physical forcing within the glacial system (e.g.

Josberger et al., 2010). A summary of particular differences in physicochemical properties between ice-contact and other (distal) proglacial lakes is provided in Table 5. Note that studies listed in Table 5 are predominantly within (temperate) high mountain and high latitude regions and that seasonality in thermal stratification is very different in the tropics.

Water temperature-density relationships in proglacial lakes are complicated, not least due to spatio-temporal scales and external (catchment, climate) and internal factors (lake geometry and physic-chemical attributes). However, glacial meltwater entering a proglacial lake will be generally near $0^{o}C$ in temperature. In contrast, hillslope or groundwater streams can be at least $10^{o}C$ warmer. In response to these temperature differences and to seasonally changing water sources and meltwater volumes, proglacial lakes typically develop a well-mixed layer of warmer and thus lower density water at the top of the water column; in high latitude and (temperate) high mountain proglacial lakes this occurs during summer. There can be an accompanying sharp decrease in temperature at the base of the lake. In the autumn, high latitude and (temperate) high mountain proglacial lake surface waters that are usually cool, become denser and consequently sink. Eventually the vertical temperature profile of such a typical proglacial lake can overturn, perhaps twice a year for lakes that are distal from an ice margin (Ashley, 1995; Richards et al., 2012).

Density stratification in proglacial lakes, and especially in ice-contact lakes, is due not only to water temperature differences, but also to suspended sediment concentration (Table 6). Meltwater streams with high sediment loads (> 1 g l^{-1}) can develop density stratification where density increases with depth. This vertical density gradient will likely be more gradual than that of temperature (Ashley, 1995). Ashley (1995) lists published suspended sediment concentrations of over 200 mg l^{-1} for proglacial lakes; although more recent publications suggest much lower values (e.g. Geilhausen et al., 2012; Liermann et al., 2012).

Overall, if there is a significant difference in density, as is typical for ice-contact lakes due to summer inputs of sediment laden meltwater, inflows will likely maintain integrity as a plume and being denser than ambient lake water will sink as an underflow current, or a gravity, density or turbidity current (e.g. Gilbert and Crookshanks, 2009). Such density currents produce graded, rhythmically-stratified sediments that infill depressions on a lake basin floor. In contrast, distal proglacial lakes can receive lower-density water from a glacial, groundwater or a hillslope source. This lower-density water rises to the surface as an overflow. Water entering a distal proglacial lake with the same density as that of the lake will circulate as an interflow (e.g. Smith et al., 1982; Francus et al., 2008). Overflow-interflow circulation is typical of distal proglacial lakes and produces rhythmically-laminated thin bottom sediments that drape topography.

iv) Chemical and biological characteristics

Chemical processes within proglacial lakes are very under-studied, perhaps because of the over-riding importance of physical processes. This lack of study is perhaps surprising because chemical characteristics and biological material within proglacial lake sediments are important for providing both relative and absolute geochronological dates as well as proxy evidence for environmental conditions at the time of sedimentation (e.g. Lamoureux and Gilbert, 2004; Etienne et al., 2006). For example, Fortner et al., (2005) showed that chemical budgets can identify changing water source contributions through time.

Biological disturbance of sediment by surface or near-surface lacustrine (or marine) dwelling organisms i.e. 'bioturbation' in distal proglacial lakes is usually well-developed since sedimentation rates are very low due to fast mixing; incoming flows are of a similar temperature and suspended sediment concentration as lake water (Table 6; Smith et al., 1982). The wide range of structures produced by bioturbation processes are trace fossils. The range of species of these fossils permits reconstruction of environmental conditions such as water depth, water temperature, turbidity and salinity, and the organic component of the fossils permits absolute geochronological dating (Andrews et al., 1996; Bujalesky et al., 1997; Uchman et al., 2009).

v) Sedimentation

It is absolutely crucial to understand sedimentation within proglacial lakes (Table 7) for (i) deciphering the Quaternary record of glaciation and glacier dynamics and (ii) distinguishing proglacial lakes from other water bodies. Sedimentation within proglacial lakes is primarily controlled by proximity to the ice-margin and by lake water density stratification. However, most mountain regions are subject to seismic activity, paraglacial processes and other geomorphological controls that also affect sediment dynamics within lakes. Specifically, sediment concentration within input streams and the vertical position of input streams relative to the total water depth are key controlling factors on sedimentation. To a lesser extent; and usually only in very large lakes, winds can set up epilimnial currents, the Coriolis force can direct currents, and the duration of ice cover on a lake can affect circulation patterns and strength via thermal stratification development (Ashley, 1995; Richards et al., 2012). This exposure of proglacial lakes to atmospheric conditions sets those sediments apart in compositional character from subglacial lakes (Livingstone et al., 2012). In all proglacial lakes, the spatial and temporal pattern of sedimentation (Table 7) can be interrupted due to cyclical and episodic or abrupt water level changes (e.g. Hicks et al., 1990; Lewis et al., 2002).

Sedimentation within all proglacial lakes tends to be dominated in volume and in areal extent by that from suspension settling (Table 7). However, sedimentation within ice-contact lakes can be dominated by the action of jets, where both bedload and suspended load are of very high volume, or from hyperconcentrated flows, or from traction carpets similar to those in turbid subaerial rivers (Francus et al., 2008; Fanetti et al., 2008; Benn and Evans, 2010). Ice rafting can be a very effective mechanism for dispersing sediment over large distances in ice-contact lakes. Ice-contact sedimentation is dominated by ice-contact deltas, subaqueous fans and submerged ramps (Tables 6, 7) but perhaps also by mass movement processes. Ice-cliff margins and submerged ice ramps are unlikely to be preserved in any coherent state. With distance from the ice margin, rhythmically laminated bottom sediments will develop (Ashley, 1995). Diamict 'moraine' material is likely to be deposited on the up-ice side of grounding line fans and ice-contact deltas. The apex of an ice-contact delta marks the position of a discrete and sustained input of meltwater, but multiple apexes could either reflect multiple inputs or else a single source that moved. The thickness of topset (subaerial) sediments on a delta is closely related to the vertical range of meltwater inputs; the difference between maximum lake level and the depth of fluvial scour by input streams.

Distal glacilacustrine sedimentation is far more likely to be preserved in the geological record and thus is far more useful as a palaeo-environmental archive than ice-contact sediments (e.g. Leonard and Reasoner, 1999; Francus et al., 2008). The physical separation of a distal proglacial lake from a glacier or ice sheet margin means that there is no moraine, minimal if any ice-rafted debris (IRD), decreased suspended load and decreased clast sizes in comparison to that within ice-contact lakes. Distal glacilacustrine sediments are characterised by deltaic sediments; fine-grained, prograding and gently-dipping foresets, and by rhythmic bottom sediments and sand, silt and clay (e.g. Loso et al., 2004). Following deglaciation, sedimentation in proglacial lakes can be dominated by paraglacial activity (Ballantyne, 2003) or hillslope processes (e.g. Gilbert and Desloges, 2005). This progression of sedimentation style is especially evident within small lake basins and in mountainous areas will predominantly comprise slumping of valley-side glacigenic deposits to generate large subaqueous debris flows (Desloges, 1994; Martini and Brookfield, 1995; Dirszowsky and Desloges, 1997; Johnsen and Brennand, 2006). Rapidly-altered sedimentation styles or rates and thus regional (non ice-contact) ice retreat or advance can be inferred by abrupt changes or unconformities in the sequence of grain size, mineralogy and biological composition (e.g. Schiefer and Gilbert, 2008). It should be noted that turbidity/gravity/density currents in proglacial lakes may only last a few minutes, but can deposit several centimetres of sediment, whereas suspension settling of fine-grained material may only deposit a few millimetres or less over many months (Palmer et al., 2007). Gilbert and Crookshanks (2009) have demonstrated and quantified the spatial and temporal variability in sedimentation rates from turbidity currents. It is clear that laminated sediments can represent depositional cycles on many timescales and the term ‘varve’ is best restricted to where a seasonal cycle can be demonstrated (c.f. Leemann and Niessen, 2004).

Identification of the vertical succession of deposits related to ice-contact, ice-distal and paraglacial conditions has enabled inference of very rapid sedimentation during glacier retreat (e.g. Leonard and Reasoner, 1999; Loso et al., 2004), slowed depositional rates after deglaciation, and subsequent sedimentation of substantial volumes of sediment through fluvial influx of reworked glacigenic sediment (Ballantyne, 2003; Etienne et al., 2006; Schiefer and Gilbert, 2008). This overall pattern is the same as those reconstructed for Late Pleistocene ice sheets (e.g. Mullins et al., 1990; Desloges and Gilbert, 1991; Gilbert and Desloges, 1992; Seltzer, 2008; Murton and Murton, 2012), and as reconstructed for past ice sheet margins, outlet glaciers (e.g. Anderson and Archer, 1999; Charlet et al., 2008) and for mountain glaciers (e.g. Gilbert et al., 1997; Lamoureux, 1999; Lewis et al., 2002; Blass et al., 2003, 2005; Loso et al., 2004; McCulloch et al., 2005; Thomas et al., 2010). There are very few studies of contemporary sedimentation within proglacial lakes, but a few selected examples are given Table 1.

Several specific sedimentary units and structures are very important to describe and understand because they can be used to both indicate and to diagnose glaci-lacustrine processes and environmental conditions. Sedimentary evidence of proglacial lakes most obviously includes rhythmites, which may include varves; however, IRD is unequivocal, or diagnostic, of a glacilacustrine (or glacimarine) environment. IRD can be identified in both massive and laminated sediments, which in proglacial lakes may only be a few tens of centimetres thick and in the latter case by the deformation or penetration of underlying laminae. Individual clasts of IRD are termed 'dropstones' and where significant numbers of dropstones are deposited in space a 'dropstone diamicton' can be recognised. The rate of deposition of IRD is a function of the debris content of the ice and the frequency of iceberg passage, which reflects the calving rate and the distance from the ice margin. Thus clast-rich dropstone diamictons are usually associated with proximal environments (Hambrey, 1994). Dropstone clast morphology is inherited from the parent glacial material; i.e. it depends on the glacial transport pathway and mechanism; active subglacial transport versus inactive supraglacial transport. Thus no systematic differences in dropstone clast morphology are evident between proximal and distal glacilacustrine environments (Hambrey, 1994). Dropstone clast orientation is preferentially vertical, due to its passage vertically through the water column, but the preservation of this orientation depends on the bottom sediment characteristics.

Iceberg grounding structures and sediments are also diagnostic of glacilacustrine (and glacimarine) environments. A key piece of diagnostic evidence of a palaeo proglacial lake is the presence of linear and curved grooves within the sediments that have ridges immediately alongside them. These iceberg keel scours and plough marks may be entirely in parallel, or may cross over each other and can therefore be used to infer the number of glacial advances and the direction of those advances. Iceberg scouring can be so intense that all primary depositional structures are destroyed leaving a chaotic massive structureless diamicton. However, more commonly distinctive folding and faulting structures can create erosional grooves and constructional berms by single iceberg grounding or ploughing events (Woodworth-Lynas and Guigné, 1990; Eyles et al., 2005).

Sedimentation volumes and rates, as obtained by the thickness of deposits and by the relative and absolute dating of deposits (usually varves) respectively, are used to infer i) water sources and dynamics, ii) proximity to an ice margin and hence glacier advances and retreats, iii) climatic conditions driving these glacier margin changes. Since the work of Jopling and McDonald (1975), it has been clear that, away from the point of influx, glacilacustrine sediment thins as accumulation rates are much reduced. It is also clear that ice-contact lakes have higher sedimentation rates than distal proglacial lakes. The rate at which sediment is delivered from a glacier into an ice-contact lake is a function of ice melt rate, glacier type and of sediment concentration within the ice.

vi) Landforms

Glacilacustrine landforms (Table 7) permit reconstructions of glacier margin position, glacier margin thickness, glacial meltwater conduit positions and types, glacier behaviour and, by inference, glacier mass balance and climate. Glacier margins in large and deep proglacial lakes are less stable than those on land because calving and surging processes mean that it is difficult for a glacier or an ice sheet to establish a fixed boundary of equilibrium. This ice margin instability, alongside the high potential for wave erosion, means that well-defined moraine ridges are unlikely to be preserved in proglacial lake basins. Thus the history of ice retreat across very large and deep proglacial basins is not very well known (e.g. Murton and Murton, 2012). Subaqueous moraines may have linear or sinuous crests and particularly in the case of DeGeer moraines occur in fields of relatively narrow sub-parallel ridges (e.g. Golledge et al., 2008). Subaqueous moraine ridges frequently display heavily glacitectonised sediment structures that are indicative of compression; i.e. ice margin pushing during advance (e.g. Bennett et al., 2000). DeGeer moraines can more simply formed by intermittent glacier margin retreat (e.g. Lindén, and Möller, 2005) as well as by brief ice margin still-stands or minor advances, and thus they have considerable implications for understanding deglaciation (Larsen et al., 2006). Subaqueous moraines can mainly consist of coalescing grounding line fans, as is the case for the Salpausselka Moraines in southern Finland (Fyfe, 1990), the eastern Maine moraines (Ashley et al., 1991) and the Oak Ridges Moraine in Ontario, for example (Russell and Arnott, 2003; Sharpe et al., 2007).

Grounding line fans are fed by flows entering a lake basin at the base of the water column, rather than at the top, as is the case for deltas. Grounding line fans can thus be used to infer ice thickness. Grounding line fans are also therefore commonly associated with specific points along esker systems where they mark former glacier margin positions. If a glacier advances, till deposition may occur as well as glacitectonic thrusts and folds.

Deltas form by a combination of fluvial aggradation above water level and progradation on the delta front where bedload is rapidly deposited when the incoming stream decelerates with still lake water interaction. Delta-front profiles are usually concave, reflecting decreasing sediment clast size due to reduced flow competence and reduced sedimentation rate. Deltas can be used to reconstruct water levels and ice margin configurations. However, most importantly quantification of delta volumes can provide valuable estimates of erosion rates within glaciated catchments (e.g. Østrem et al., 2005). It is usual to distinguish between ice-contact deltas formed within ice-contact lakes and glacier-fed deltas formed within distal proglacial lakes. This is not least because the morphology and sedimentology of deltas reflects the gradient of the feeder river and the water depth, but also possibly temporal changes in glacier coverage within the catchment (e.g. Dirszowsky and Desloges, 2004). Rising water depth can produce stacked delta sequences or else deltas can become incised in the case of falling levels. Changing water depth can also be recorded within the internal sedimentology of lake deltas (e.g. Østrem et al., 2005; Russell, 2007), as can climatic-catchment processes, glacial processes and associated water and sediment loading (e.g. Brandes et al., 2011), or lake drainage.

3. Quaternary importance of proglacial lakes

Proglacial lake evolution, character and behaviour, glacier and ice sheet growth and decay, meltwater and sediment fluxes directly affect geological systems and these interactions are summarised in Table 6. Furthermore, proglacial lakes demonstrate complex system interdependencies and feedback mechanisms on global to local scales (Fig. 6). These interdependencies are discussed in the following sections in order to (i) better understand the Quaternary record by using recent quantitative studies of modern analogues, and (ii) enable future studies to make a judgement of the geological importance of the evolution and development of contemporary proglacial lakes around the world.

3a) Proglacial lake influences on ice dynamics

Glacier margin morphology (section 2a), physical stability and dynamics (section 2b) are affected by the presence, character and behaviour of an ice-marginal lake. Indeed, the growth of an ice-marginal lake (section 2c) can affect the longitudinal stress balance of a glacier and therefore glacier velocity and mass balance (Fig. 7). This interaction between ice-marginal lakes and glacier ice dynamics has been measured for modern mountain glaciers (e.g. Kirkbride and Warren, 1999; Diolaiuti et al., 2006; Röhl, 2006) and has been modelled for Quaternary ice sheet margins (e.g. Cutler et al., 2001). Although Price et al., (2008) did not explicitly consider ice-marginal lakes, they have shown in a modelling study of a modern part of the western Greenland ice sheet that ice sheet margin conditions fundamentally affect ice sheet dynamics up to the equilibrium line. Overall, ice-marginal lakes can control glacier ice dynamics via eight mechanisms; (i) raising the water table, (ii) raising water temperature, (iii) raising subglacial water pressure, (iv) increasing ice surface gradient, (v) promoting flotation, (vi) promoting calving, (vii) intense ice flexure and fracture during rapid draining or filling, (viii) intense flushing of sediment from beneath a glacier and intense aggradation of sediment at a glacier terminus during rapid draining (Fig. 7). The Quaternary record has been used to suggest that it is also possible that ice-marginal lakes can encourage development of ice-streams (e.g. Stokes and Clark, 2003, 2004; Demidov et al., 2006).

Water at the base of a glacier is fundamental to controlling processes of ice motion via the longitudinal stress balance and specifically via decreased friction, enhanced basal sliding and saturation of underlying sediments promoting deformation, for example. The depth of water at an ice-margin determines (i) the distance 'up-ice' that water propagates, (ii) vertical extension of a basal hydrological system (Harper et al., 2010) via basal water pressure (e.g. Tsutaki et al., 2011) and (iii) calving rates (e.g. Pelto and Warren, 1991) and these factors encourage faster ice velocity and heighten mass loss (Table 6; Fig. 7). It is also plausible that where large volumes persist at the bed of a glacier; and perhaps up into a glacier (Harper et al., 2010), basal water pressures not only promote sliding if they approach the overburden pressure (Tsutaki et al., 2011), but could result in hydro-fracturing and basal crevasse development and propagation (c.f. van der Veen, 1998). As an ice-marginal lake increases in size and mass loss from a glacier snout progresses, a glacier ice surface must steepen thus increasing driving stresses and glacier velocity in a positive feedback loop (Fig. 7).

Glacier mass loss and glacier velocity are also augmented by an ice-marginal lake because lake water delivers heat to a glacier and thus causes thermally-induced melting. Thermal melting can cause notches to develop at the water line and this thermal undercutting can apparently control calving (e.g. Kirkbride and Warren, 1999; Diolaiuti et al., 2006; Röhl, 2006), especially for glaciers with velocities, low calving rates and low surface gradients (Röhl, 2006). Quantification of such thermal melting is very rare, but subaqueous melting of 9 $m.yr^{-1}$ has been recorded by Hochstein et al., (1998), for example. Furthermore, heat delivered from ice-marginal lakes to glaciers controls thermal erosion far beyond the immediate extent of the ice-marginal lake itself by producing ponded meltwater beneath and within a glacier, particularly in crevasses (Fig. 7). An ice-marginal lake will therefore cause glacier margin fluctuations, glacier velocity and glacier mass balance to be at least partially decoupled from climate (Kirkbride, 1993). In New Zealand, for example, glaciers terminating in lakes have termini that have retreated the farthest and moved at the fastest speeds in comparison to all other types of glaciers (Chinn, 1999) and over the last century have shown more variability in recession and far lower mean elevation changes (Chinn, 1996). In contrast, the relative stability of the margin of the Moreno glacier in Patagonia has in part been attributed to calving processes and to the bottom topography of a proglacial lake (Rott et al., 1998).

Where ice-marginal lake water is sufficiently deep, relative to the ice thickness, buoyancy will cause flotation of an ice margin and rapid calving, snout retreat and surface lowering (e.g. Naruse and Skvarca, 2000; Boyce et al., 2007; Tsutaki et al., 2011), or sudden glacier lake drainage, i.e. a 'jökulhlaup' (e.g. Huss et al., 2007). If calving generates ice bergs at a rate greater than iceberg melt, the resultant accumulation of ice on a lake surface can buttress an ice-margin and stabilise a glacier snout (Geirsdóttir et al., 2008). In contrast, when glacier ice is grounded with lake water, that ice is often in tension and near fracture and consequently unstable (e.g. Pelto and Warren, 1991; Kirkbride and Warren, 1999; Diolaiuti et al., 2006; Röhl, 2006).

Whenever ice-marginal lakes drain rapidly or fill rapidly, which often occurs in cycles (Evans and Clague, 1994; Russell et al., 2011), the water mass releases/exerts stress on the ice margin that is manifest in ice flexure (e.g. Walder et al., 2006) or fracture. Jökulhlaups that drain along the front of an ice-margin can cause mechanical erosion of that ice margin (e.g. Sugden et al., 1985) causing temporarily unstable ice cliffs. Jökulhlaups that drain subglacially can temporarily increase basal water pressure over a broad region; promoting sliding, inhibiting cavity closure and blocking drainage (e.g. Anderson et al., 2005), producing glacier velocity changes (e.g. Anderson et al., 2005; Huss et al., 2007; Sugiyama et al., 2007). The Quaternary record suggests that glacier velocity can accelerate due to the drainage of ice-dammed lakes (Meinsen et al., 2011; Lovell et al., 2012) and thus for glacier dynamics to be decoupled, at least temporarily, from climatic perturbations.

Modern subglacial jökulhlaups have been noted to be highly effective in flushing sediment from a glacier bed (Russell et al., 2006) and at redistributing it to glacier margins. Sediment aggradation at a glacier meltwater portal will promote ice-contact delta and braid-plain formation (section 2fv;

Fleisher et al., 2003) and thereby alter the grounding-line dynamics for water-terminating glaciers (Lindén and Möller, 2005; Benn, 2008). Enhanced sediment accumulation in water at glacier termini has the potential to physically buttress those glacier termini and thus to remove them from both the immediate effects of both lake level change and climate change.

Quaternary ice-marginal lakes have been highlighted as key factors in the location and initiation of ice streams (e.g. Stokes and Clark, 2003, 2004; Demidov et al., 2006) and hence in the onset and progression of ice sheet deglaciation. The mechanisms by which a proglacial lake could control ice streams are rather the same as already discussed for ice-marginal lakes controlling glacier dynamics; specifically that if an ice sheet with a steep surface gradient and consequent high driving stresses entered a proglacial lake, that was deep relative to ice thickness, then an increase in velocity through calving would propagate fast ice flow up-glacier by altering the longitudinal stress balance through a series of thermo-mechanical feedback mechanisms (Stokes and Clark, 2003).

3b) Proglacial lake influences on meltwater, sediment fluxes and landforms

Proglacial lakes interrupt the passage of meltwater from a glacier through a proglacial zone and reduce flow velocities sufficiently to cause sedimentation. They therefore act as a trap for sediments which would otherwise be transported into a proglacial zone and beyond. This trapping and storage is well-recognised and there is a plethora of work on both Quaternary and modern lake sedimentation (e.g. Hicks et al., 1990; Kirkbride, 1993; Marshall and Clarke, 1999; Hasholt et al., 2000; Lewis et al., 2002; Schomacker and Kjaer, 2008; Fujita et al., 2009; Liermann et al., 2012). Such sediment sinks have great potential to record relatively high-resolution evidence of sediment delivery from the catchments in which they are situated; they can provide information on medium to long-term sediment yields on a catchment scale (e.g. Liermann et al., 2012) and chronologies of lake sediments can be used as proxy evidence for regional hydrological changes and climate variability (Desloges and Gilbert, 1998). Sediment fluxes in glacial environments are usually considered to be highest immediately following deglaciation due to over-steepened relief and paraglacial slope readjustment (e.g. Hallet et al., 1996; Leonard, 1997; Ballantyne, 2000; Orwin and Smart, 2004). However, there is potential for larger glaciers to produce higher sediment yields than smaller glaciers or retreating glaciers. This complication as to whether higher sediment yields represent deglaciation or increases in the size of ice masses has important implications for using sediment records in the context of glacier and climate reconstructions. Recent observations of the southern outlets of Vatnajökull in Iceland indicate that proglacial lakes are acting fluvial sediment traps, reducing sediment input into the Atlantic coastal system (Syverson, 1998; Schomacker and Kjær, 2008). Proglacial lakes are also traps for aeolian sediment. For example, development of proglacial lakes at the margins of the Patagonian ice field resulted in the trapping of fine-grained aeolian sediment in those lakes in the early phases of the last glacial retreat and thus stopped the ice age dust flux to Antarctica (e.g. Ackert, 2009).

Proglacial lakes can act as sediment stores as a consequence of episodic processes that are seemingly independent of climate change. For example, the eruption of Eyjafjallajökull in 2010 caused rapid melting of glacier ice resulting in debris-rich jökulhlaups, which completely infilled a moraine-dammed lake at the snout of the glacier Gígjökull. This sediment has the potential to be re-mobilised by further high-magnitude jökulhlaups triggered by future eruptions of Eyjafjallajökull. In such cases, proglacial lake evolution and sediment flux are linked to cycles of volcanic activity rather than glacier retreat, although ultimately climate change may be implicated in intensification of subglacial eruptive activity (e.g. Carrivick et al., 2009; Tuffen, 2012; McGuire, 2013).

Episodic events exert profound control on meltwater and sediment fluxes in glaciated catchments over very short time periods. For example, hydrological diversion or re-routing and even reversal of rivers, and consequent geomorphological activity including gorge incision, valley formation and sedimentation has been attributed to GLOFS or jökulhlaups soon after the Last Glacial Maximum (LGM). Global examples are given by Baker (2003); examples in northern Russia are explained by Astakhov et al. (1999), Mangerud et al. (2001) and Krinner et al. (2004); and Murton and Murton (2012) give many examples from the Quaternary in the UK. Some of these proglacial lake drainages produced jökulhlaups so large that the resultant influx of freshwater and sediment affected ocean circulation and climate; this is addressed in section 3c.ii. Modern outburst floods from ice-, moraine- and landslide-dammed lakes in the Himalayas, Canada, New Zealand and Tien Shan have caused similar hydrological and geomorphological impacts(e.g. Desloges and Gilbert, 1998; Cenderelli and Wohl, 2003; Korup et al., 2006).

Quantification of the contribution of such episodic sediment pulses to long-term sediment budgets is difficult because of problems in obtaining direct measurements (e.g. Korup et al., 2004). However, this quantification needs addressing, because glacifluvial, aeolian and marine sediment transfer systems will be increasing affected by formation and drainage of moraine- and rockslide-dammed lakes that is predicted to accompany deglaciation due to climate change (e.g. Richardson and Reynolds, 2000; Chikita and Yamada, 2005). The Quaternary record might offer help in this quantification; (i) it should be possible to infer meltwater volumes and meltwater generation rates from the spatial and temporal patterns in proglacial lake evolution; regional contrasts in lake evolution have been attributed to describe climatic variability (e.g. Gardelle et al., 2011); (ii) sedimentation rates in neighbouring catchments with and without a history of outburst floods could be compared. We note that identification of storage elements within catchments such as proglacial lakes is critical to the effective study of sediment budgets (Reid and Dunne, 1996).

Proglacial lakes most obviously influence terrestrial geomorphology within very large continental basins, which were occupied by exceptionally large Quaternary proglacial lakes (Fig. 1). These lake basins received and retained large volumes of glacial and glacifluvial sediment during deglaciation from the LGM. The sediments, still well-preserved, are an invaluable (i) geochronological archive (section f) and (ii) support modern-day Arctic and sub-Arctic peatlands that are very important for the global carbon balance. Glacilacustrine sediments can dominate moraine facies, for example as for the Salpausselkä moraines that in part determined by lake water depth (Fyfe, 1990). Furthermore, whilst lake water level dynamics, and in particular lake 'transgressions' (c.f. Teller, 2001) controls beach development and character, it is these palaeo-lake shorelines that have a geological importance because they can be used to measure regional isostatic uplift (Teller, 2001). Quaternary proglacial lakes influenced terrestrial geomorphology more widely by promoting continental hydrological drainage reversals; examples have been reported in Siberia and mainland Europe (e.g. Mangerud et al., 2004; Svendsen et al., 2004) and in the UK (e.g. Murton and Murton, 2012).

Quaternary GLOFs or jökulhlaups have indirectly extended the geomorphological influence of proglacial lakes to marine settings. Terrestrial canyons (e.g. Baker and Bunker, 1985; Herget, 2002) and submarine canyons (e.g. Thieler et al., 2007; Elliot and Parson, 2008) have been excavated by drainages of Quaternary proglacial lakes. In terms of deposition, Quaternary jökulhlaups have emplaced giant submarine fans onto continental shelves (e.g. Brown and Kennett, 1998; Geirsdóttir et al., 2002). Sedimentological analyses in Alaskan fjords have suggested that marine impacts of Quaternary jökulhlaups may be more widespread than previously thought, because many deposits previously interpreted as 'glacial' could in fact be products of jökulhlaups (Willems et al., 2011).

3c) Proglacial lake and climate interactions

The relationships between proglacial lakes and climate have been researched extensively; it is therefore our intention to only outline the most salient elements of these relationships. Proglacial lakes are intrinsically linked to wider geological and environmental systems; namely weather and climate, meltwater and sediment fluxes and glacier and ice sheet dynamics. These interactions are briefly discussed below, highlighting interdependencies and feedback mechanisms.

i) Proglacial lake influences on local and regional weather

In comparison to bedrock, soil or vegetation covered land and glacier ice, (unfrozen) proglacial lakes have a lower albedo (e.g. Sakai et al., 2000) and a higher thermal heat capacity. Proglacial lakes thereby affect the local surface energy balance and in particular summer ablation, which is the major component of annual mass balance for most alpine glaciers (Oerlemans and Reichert, 2000). Specifically, on a local scale ice-marginal lakes can absorb relatively high amounts of incoming shortwave radiation during summer, freeze in winter and hence reflect most incoming shortwave radiation (Björnsson et al., 2001), retain relatively cold meltwater and have a thermal heat capacity greater than that of surrounding land, which combined lead to relatively cool summer air temperatures and relatively warm autumnal air temperatures.

On a regional scale, substantial lakes can directly influence weather patterns (e.g. Long et al., 2007 Brown and Duguay, 2010). Krinner et al., (2004) and Mangerud et al. (2004) discuss the feedback effects between ice-dammed lakes, climate and glacier fluctuations in northern Eurasia during the earliest stages of the last glaciation there (90 - 80 ka). Ice sheet growth over the Barents and Kara seas expanded south onto the Russian mainland and led to the damming of north flowing rivers forming ice-dammed lakes. Modelling results suggest that these lakes remained cool in summer, not only because they were in receipt of large quantities of water from the ice sheet margin but also as a consequence of the thermal heat capacity of water being far greater than that of land. The resultant summer cooling, which was strong over the lakes themselves, also extended onto the ablation zone of the ice sheet (Mangerud et al., 2004), and in turn reduced summer melt across the margin of the ice sheet and increased ice mass balance. It is thought that such 'coupled lake-ice sheet systems' could be self-sustaining and play an important role in climate modification (Krinner et al., 2004). This relationship is finely balanced; if ice sheet decay triggers ice-dammed lake drainage, de-coupling occurs. Numerical modelling by Peyaud et al. (2007) that was designed to further illuminate the impacts of Quaternary ice-dammed lakes on ice sheets also supports these findings. It is clear that further work on the relationships between proglacial lakes, ice sheet and climate will assist in environmental change reconstruction and the prediction of the impacts of current global deglaciation. If proglacial lakes modify regional climate, then indirectly they may assist in the preservation or degradation of ice sheets (Fig. 6). For example, work by Hostetler et al. (2000) simulated the effects of Lake Agassiz on the climate of central North America at 11 ka using a high-resolution regional climate model nested within a general circulation model and demonstrated that the lake cooled the climate at the southern edge of the Laurentide Ice Sheet.

ii) Proglacial lake influences on global climate

There is well-established evidence that the drainage of large Quaternary proglacial lakes affected ocean circulation and thereby changed climate due to the sudden influx of very large (~ 1 M m^3s^{-1}) meltwater pulses into the oceans and there is a substantial body of work on these events and their implications (e.g. Barber et al., 1999; Marshall and Clarke, 1999; Clarke, et al., 2001; Mangerud et al., 2001; Teller et al., 2001; Fisher et al., 2002;Teller et al., 2002; Clarke, et al., 2003; 2004, 2009; Rayburn et al., 2011). Late Pleistocene megafloods from the margins of the Laurentide ice sheet in Canada and North America have been extensively studied; the fluctuations of the ice sheet resulted in the complex evolution and drainage of the basins that now contain the Great Lakes (e.g. Kehew and Lord, 1987; Teller and Kehew, 1995). Particular attention has been paid to the outflow of large quantities of freshwater into Arctic Ocean from the subglacial drainage of 'superlake' Agassiz. The timing, routing and duration of catastrophic meltwater discharges from Lake Agassiz and other proglacial lakes has been debated at length (e.g. Broecker, 2006; Lowell et al., 2005; Tarasov and Peltier, 2005; Lepper et al., 2007; Fisher et al., 2008; Murton et al., 2010; Rayburn et al., 2011), not least because the final drainage of Lake Agassiz into Hudson Bay could have triggered the Younger Dryas cold event - a millennial-scale reversal to glacial conditions during the last glaciation - as a consequence of the disruption of the salinity gradient that drives meridional overturning circulation (c.f. Teller, 2012).

Development and drainage of huge proglacial lakes in Eurasia could also have perturbed regional hydrological conditions and thereby indirectly affected global climate. For example, in Russia, ice sheets developing in the north blocked large northwards flowing rivers creating Lake Komi, the Baltic lake and lakes in the White Sea Basin and the West Siberian Plain (Astakhov et al., 1999; Mangerud et al., 2001; Krinner et al., 2004; Margold et al., 2011); these lakes would have drained into the Arctic Ocean, influencing sea ice formation and oceanic circulation (Mangerud et al., 2001). Margold et al., (2011) have proposed that flooding from Glacial Lake Vitim in Siberia caused climatic and environmental impacts as a consequence of large freshwater influx into the Arctic Ocean, further suggesting that the drainage of Lake Vitim was responsible for a freshwater spike at - 13 ka inferred from Arctic Ocean sediments (Speilhagen et al., 2005).

It is clear that further research on the sensitivity of ocean circulation to freshwater inputs and the extent to which the drainage of large proglacial lakes can modify the behaviour of ocean circulation and trigger climate oscillations will be increasingly important, given the development of proglacial lakes concomitant with current deglaciation.

4. Summary and conclusions

In summary, geological evidence from the Quaternary shows that proglacial lakes were a major component of deglaciation. At present, proglacial lakes are increasing in number and size around the world. Ice-marginal or 'ice-contact' lakes exert a control on mountain glacier and ice sheet margin dynamics via the longitudinal stress balance. Glacier margin morphology, physical stability and ice dynamics are all affected by the presence, character and behaviour of an ice-marginal lake. Growth of an ice-marginal lake can have a positive feedback with glacier dynamics and glacier mass loss, due to the controls of; (i) raising an englacial water table, (ii) raising englacial water temperature, (iii) raising subglacial water pressure, (iv) increasing ice surface gradient, (v) promoting ice margin flotation, (vi) promoting calving, (vii) intense ice flexure and fracture during rapid draining or filling, (viii) intense flushing of sediment from beneath a glacier and intense aggradation of sediment at a glacier terminus during rapid draining. Proglacial lakes interrupt routing of ablation-fed meltwater, episodically released meltwater and sediment from a glacier system to a proglacial zone. Sedimentation in proglacial lakes is one of the most important; continuous and high-resolution geochronological archives. Proglacial lakes have been documented to impact continental hydrological drainage, and via jökulhlaups can create spectacular terrestrial and submarine landforms including canyons and extensive fans. Proglacial lakes can affect weather and climate via albedo and thermal heat capacity controls, and via sudden drainage to produce jökulhlaups that abruptly deliver freshwater and sediment in sufficient quantity to modify ocean circulation and climate.

Scientific advances in understanding of the Quaternary can be brought about by (i) drawing terrestrial and marine records together, and (ii) considering a range of approaches and methodologies across multiple scientific disciplines. However, there are hitherto untapped opportunities for utilising emerging technologies to understand proglacial lake processes and to revisit or extend proglacial lake geochronological archives. For example, autonomous underwater vehicles (AUVs) with capability to simultaneously measure bathymetry, water currents and water quality parameters in three-dimensions and at high repeatability could offer great insights to modern processes of sedimentation, water circulation and indeed subglacial water routing and discharge regimes. Water surface-deployed acoustic systems such as 'sediment echo-sounders' or 'sub-bottom profilers' that are capable of imaging sediment stratigraphy through a water column should offer unprecedented information on both modern Quaternary sedimentation styles and rates.

In conclusion, our knowledge of the character, behaviour and geological importance of proglacial lakes has improved markedly over the past few decades at least in part due to a shift from mono-disciplinary studies to more integrated studies. However, whilst there is much more holistic work to be done, and this study is a start, it is abundantly clear that proglacial lakes are intrinsically linked to climate change and to wider geological systems through glacier mass balance and hence glacial meltwater and sediment fluxes. Proglacial lakes could be an integral part of deglaciation by affecting the stability and character of glacier margins. They can disengage glacier behaviour from climatic perturbations. Proglacial lakes interrupt the delivery of meltwater and sediment to proglacial zones and ultimately to oceans. Sedimentation within proglacial lakes is a very important geochronological archive recording short-term inter-seasonal patterns, inter-annual patterns, and long-term patterns of glacier-derived meltwater fluctuation and hence, by proxy, glacier mass balance. Overall, proglacial lakes are exceptionally useful for understanding (i) patterns, character and behaviour of mountain glaciers and glaciations, (ii) ice sheet dynamics and driving processes, iii) past deglaciation, iv) the geological importance of present-day proglacial lakes, and (v) future deglaciation in both mountain and ice sheet environments. Looking forwards, proglacial lakes will continue to increase in number size due to mountain glacier and ice sheet mass loss. Ice-marginal lakes will de-couple mountain glacier and ice sheet margin changes from climate and will accelerate glacier mass loss in the short-term. Outburst floods from proglacial lakes will therefore also increase in frequency and perhaps also in magnitude, thereby constituting a persistent hazard.

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Figure 7. Summary schematic of the influence of an ice-marginal lake on glacier dynamics. Forces are in italicised text and with black arrows. Processes are in black text and with grey dashed lines to denote interactions.

Region	Selected recent references
Greenland	Russell et al., 1990; Hasholt, 1993 ; Hasholt et al., 2000
Iceland	Roberts et al., 2005; Schomacker, 2010
European Alps	Haeberli et al., 2002; Huggel et al., 2002; Huss et al., 2007
Tibetan Plateau	Richardson and Reynolds, 2000; Chen et al., 2007; Yao et al., 2010; Wang et al., 2011
Himalaya	Fujita et al., 2009
Alaska	Fleisher et al., 2003; Loso et al., 2004; Walder et al., 2005;Josberger, et al., 2010
South America	Harrison et al., 2006; Pasquini et al., 2008; Thompson et al., 2011
Caucasus Mountains, Russia	Stokes et al., 2007
New Zealand	Hicks et al., 1990; Kirkbride, 1993; Hochstein et al., 1998
Norway	Liermann et al., 2012
Canada	Lewis et al., 2002; Lamoureux and Gilbert, 2004

Table 1: Selected examples of studies on modern proglacial lakes

Region	Lake	Selected recent references
North America	LakeAgassiz	Teller et al., 2002; Clarke et al., 2003, 2004;
Baltic	Baltic ice lake	Bodén et al., 1997
North America	LakeBonneville	O'Connor, 1993
North America	Glacial LakeMissoula	O'Connor and Baker, 1992
West Siberia	Mansi, Komi	Komatsu and Baker, 2007
Central Asia	Khvalyn	Arkhipov et al., 1995
Southern Siberia	Lakes feeding the YeneseiRiver catchment	Krivonogov et al. 2005; Komatsu et al. 2009
South-central Siberia	Chuja-Kuray (Altai flooding)	Carling et al., 2002; Herget, 2005; Reuther at al., 2006
Kirgizstan	LakeIssyk-Kul	Grosswald and Rudoy, 1996
Siberia	Lake Baikal	Baker, 2007

Table 2: Selected studies on palaeo proglacial lakes.

Location	Example references
European Alps	Paul et al., 2007
Iceland	Schomacker, 2010
Svalbard	Schomacker and Kjær, 2008
Caucasus Mountains	Stokes et al., 2007
South America and Patagonia	Ames, 1998; Hubbard et al., 2005; Loriaux and Cassasa, 2012
Canada and North America	O'Connor and Costa, 1993; Clague and Evans, 2000
New Zealand	Chinn, 1996
Himalaya	Gardelle et al., 2011
Mt Everest region	Tartari et al., 2008
Bhutan	Komori., 2008
Tibet	Ageta et al., 2000; Chen et al., 2007; Komori, 2008; Wang et al., 2011

Table 3: Examples of studies documenting and analysing increased number and size of proglacial lakes (usually ice-marginal lakes) around the world.

Decreased water depth	Impact on glacier	Increased water depth
Reduced	Likelihood of calving Volume of calving	Enhanced
Convex	Ice surface profile	Concave
Increased	Likelihood of a frozen (cold) ice margin	Reduced
Increased	Overall: ice margin stability	Reduced (possible retreat, possible surging)

Table 4: Summary of the control of water depth of ice-contact lakes on ice margin character and behaviour.

Ice-contact (proximal)	Physicochemical property	Proglacial (distal)
higher	Suspended sediment concentration	lower
higher	Quantity and clast size of ice rafted debris, drop-stones	lower
higher	Sedimentation rate	lower
lower	Light penetration to depth	higher
lower	Water temperature	higher
lower	Animal and plant abundance	higher
lower	Taxonomic diversity	higher

Table 5: Summary of differences in physicochemical properties between ice-contact and other (distal) proglacial lakes. These properties determine the geochronological archive preserved within each type of lake.

	Controls	Character	Behaviour	Geological importance
Water depth	Ratio of inflow versus outflow, plus for ice-contact lakes; ice thickness, surrounding topography and dam stability	Highly variable spatially and temporally between lakes and within lakes	Can be static, linear, or cyclical in evolution	Deep water (relative to ice thickness) promotes calving, a concave glacier surface profile and thus ice margin retreat. Increasing water depth can cause ice margin flotation, flexure or fracture, and jökulhlaups. Deep water also probably promotes a distributed subglacial drainage system. Controls stratigraphy (evolution) of deltaic sediments and in particular deeper water (away from the ice margin) promotes laminations and couplet (varve) formation
Bathymetry	Glacially-subdued, scoured and smoothed terrain overlain by a drape of silts and clays	Flat floor and near-impermeable surface		One of the most obvious and ubiquitous indicators of past lake positions and extents. Encourages postglacial peatland development
Water density stratification	Temperature and suspended sediment concentration; i.e. type and balance and regime of water sources	Magnitude of density largely determined by suspended sediment. Density profile dominated by dynamics of water temperature	Seasonal over-turning of temperature profile. Basal sediment plumes common (density-currents)	Identification of density gradient type can help to distinguish between an ice-contact lake (density gradient dominated by suspended sediment) and a non ice-contact lake (density gradient dominated by temperature). Density gradient controld circulation and hence sedimentation style and patterns. Controls chemical and biological activity.
Sedimentation	Ice berg production / calving. Suspended sediment concentration of input streams. Distance from ice-margin. Stability of ice-margin and of glacial drainage network	Till, deltaic deposits, varves and other rhythmites, ice - rafted debris (IRD; dropstones), beach or shoreline deposits. Proximal ice-contact moraines, deltas, fans and ramps to ice-distal foresets to paraglacial (slump and hillslope) sedimentation.	Diurnal and seasonal cycles of sedimentation Spatial transition with distance from ice-margin.	Representation of control of glacier behaviour, ambient climate and hillslope hydrology, (Leemann and Niessen. 2004). Dropstones are diagnostic of ice berg and hence of a calving ice margin. Varves in particular are used to establish glacial and galcilacustrine chronologies. Denotes occurrence and style of deglaciation. Glacilacustrine sediments form extensive outcrops (Teller, 1987) and typically form valley-fills that are a major source of fluvially reworked sediment during the Holocene (Ballantyne, 2003 and references cited therein)
Landforms	Local topography, fluvial / glacio-fluvial processes, mass movements (e.g. Rubensdotter and Rosqvist, 2009)	Deltas, 'Delta moraines', 'De Geer moraines', 'washboard' or 'sub-lacustrine moraines', Wave-cut beaches; 'shorelines' or 'glacilacustrine terraces' Grounding line fans Ice berg scours and plough marks		(e.g. Larsen, et al., 2006; Golledge et al., 2008) For very large (Pleistocene) lakes deformed beach elevations used to reconstruct isostatic rebound (and hence ice thickness) Used to infer ice-margin position and thickness. Direction of ice advance and if cross-cutting number of ice advances

Table 6: Summary of the controls, physical characteristics, spatial and temporal dynamics and geological importance of proglacial lakes.

Sedimentological description	Inferred processes	Indicative landform
Well-sorted coarse sediments. Parallel bedded, steeply dipping foresets (20 – 33^{o}), normal and inverse graded beds of sand and gravel. Clasts parallel to bedding planes.	Mass movements (creep, slump, debris flows, grain flows, surge currents)	Coarse-grained (Gilbert-type) delta
Well-sorted medium to fine sediments. Sets of climbing ripple sequences, draped lamination occurring as low angle (< 20o) foresets of sand and silt. Deformational structures and dewatering structures common.	Rapid deposition from density underflows	Fine-grained delta
Proximal to distal open work gravel, chaotic bedding, convolute lamination, dewatering structures, climbing ripple sequences and drape lamination.	Deposition from a high velocity expanding jet	Sub-aqueous fan
Poorly-sorted, inter-bedded mass flow deposits, lacustrine mud and dropstones	Mass-movements (rockfall, debris slides) calving, squeezed debris at base	Ice cliff margin
Poorly-sorted stacked diamicts interbedded with sand beds and mud drapes	Mass movements (debris flow, creep, slumps) surge currents	Submerged ice ramp
Fine-grained, rhythmic, parallel-bedded sediments with occasional dropstones and biogenic structures.	Deposition from underflows, interflows and overflows and occasional ice bergs.	Lake bottom

Table 7: Ice-contact proglacial lake landforms and sediments adapted from Ashley (1995).

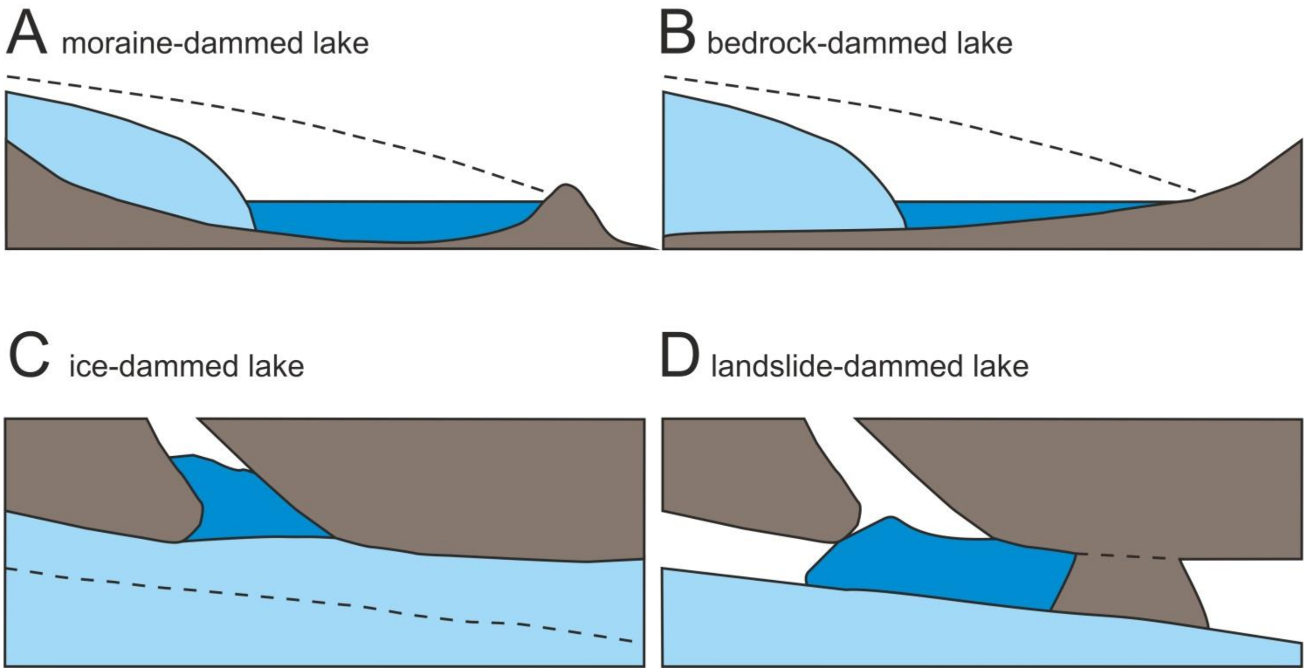

Figure 1: Proglacial lake evolution in response to ice advance and retreat, where dashed line indicates a previous ice margin, or slope margin in part D. Note that part A and B are in longitudinal (side) view, and parts C and D are in plan view. No spatial or temporal scale is implied.

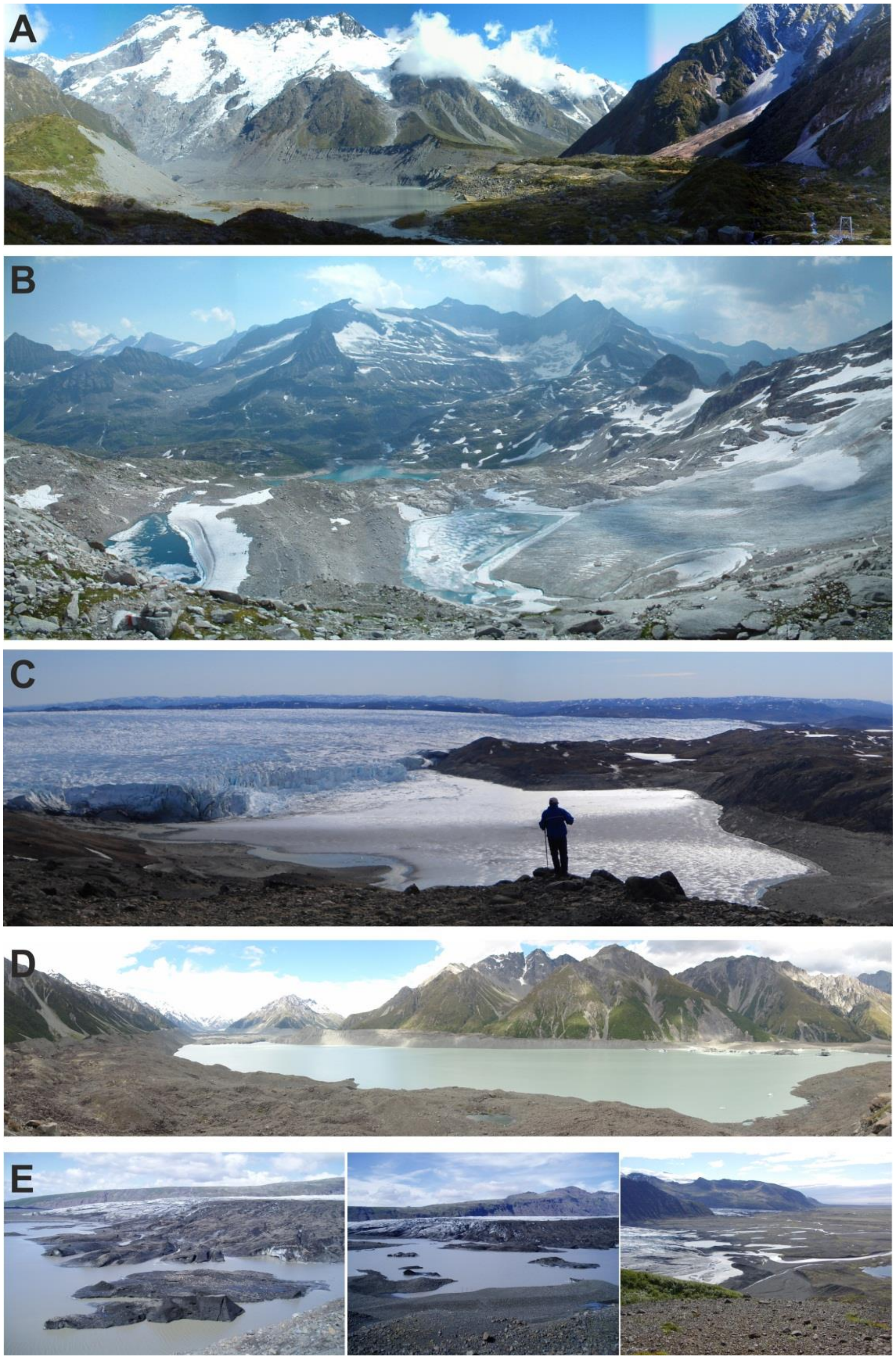

Figure 2: Examples of proglacial lake types; A: moraine-dammed lake at Mueller Glacier, Aoraki-Mt Cook, New Zealand; B: bedrock-dammed lake at Sonnblickkees, central Austria; C: ice-dammed lake at Russell Glacier, western Greenland; D and E: lake dammed by sandur as glacier surface downwasting of ice surface proceeds at Tasman Glacier, New Zealand and at Skaftafellsjökull, southern Iceland, respectively.

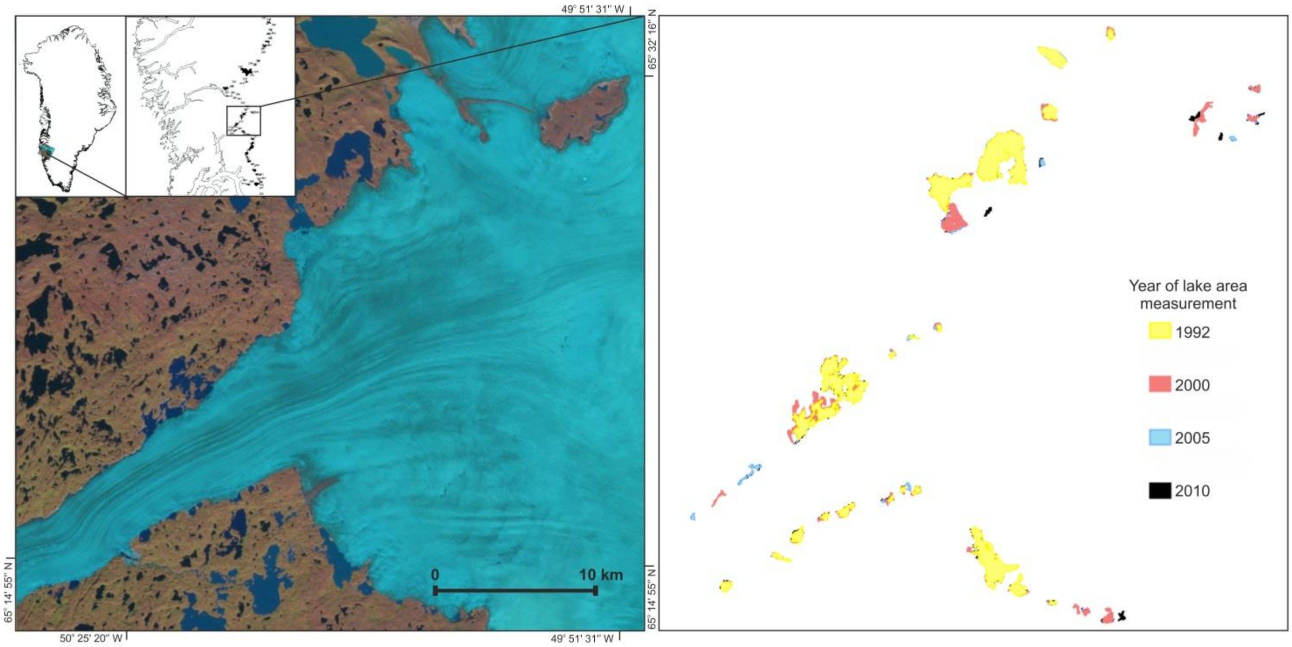

Figure 3: Illustration of the development of multiple proglacial lakes; in this case in the vicinity of Tasersuaq Glacier, west Greenland. Note that proglacial lakes in this area are increasing in number and in size and occur on the margins of the Tasersuaq outlet glacier, on the margins of the ice sheet itself and around nunataks.

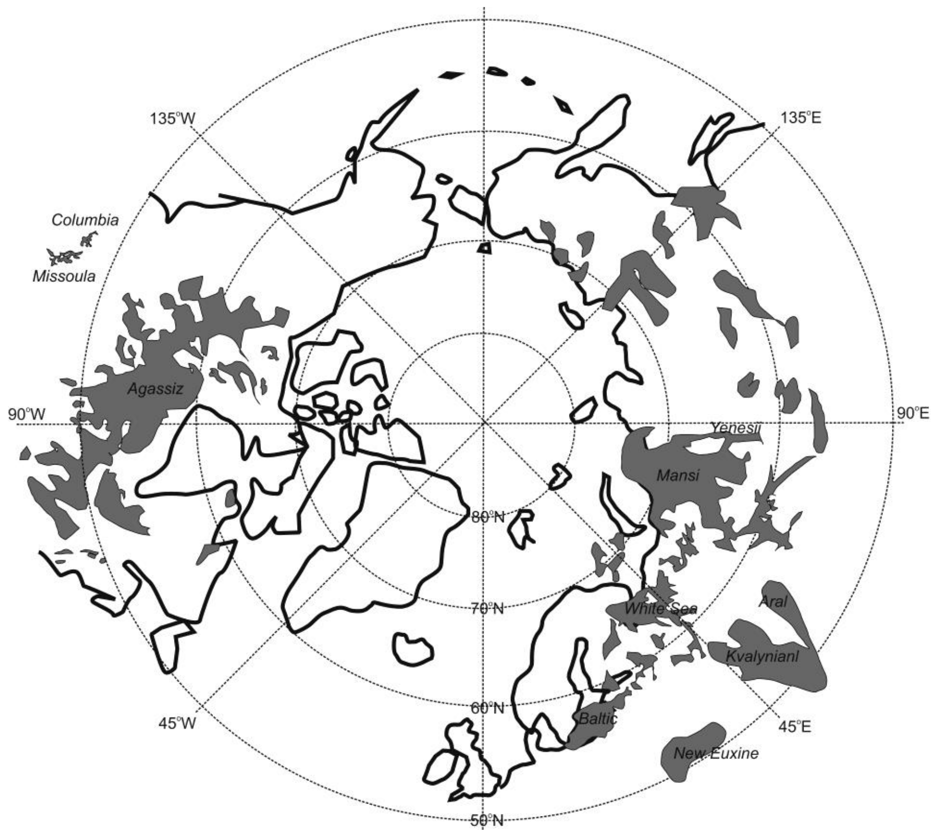

Figure 4: Schematic extent of major Quaternary proglacial lakes; i.e. land surface directly affected by major proglacial lakes. Note that not all lakes existed simultaneously and some lakes drained and re-formed several times. This figure is a compilation of reconstructions, predominantly those by Grosswald (1988), Teller (1987), Mangerud et al., (2004), Spielhagen et al., (2004), Baker (2007) and Astakov (2008). Note that the geological impact of some of these lakes is considerably more widespread than that depicted due to glacial lake outburst flood (GLOF) or 'jökulhlaup' erosion and deposition.

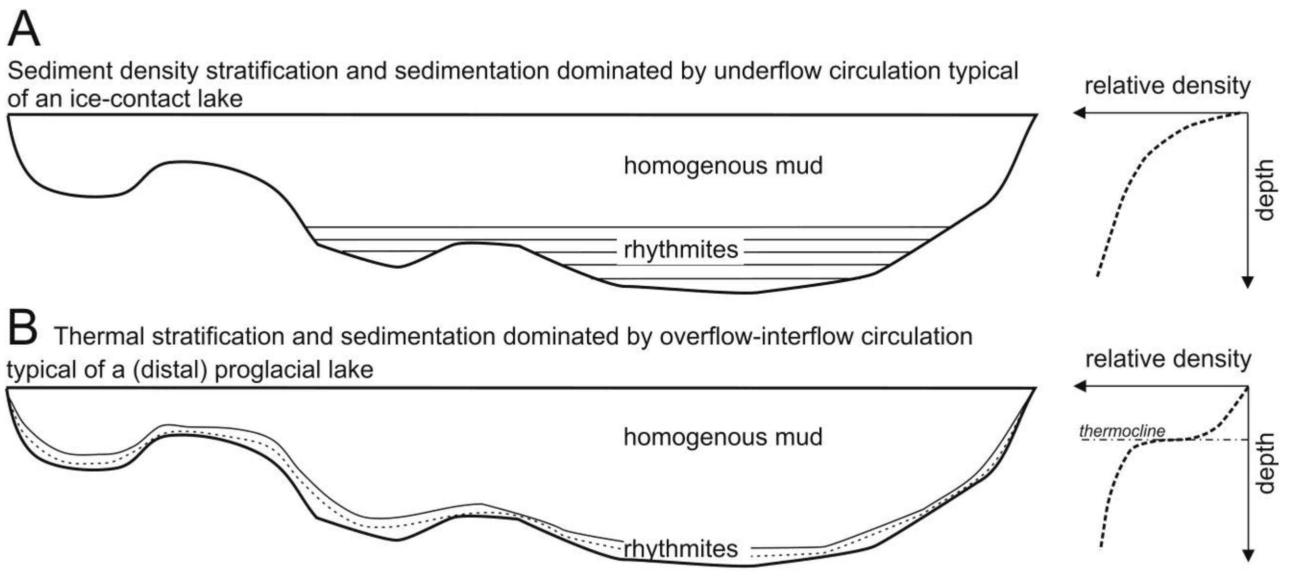

Figure 5: Spatial variations in lake sedimentation characteristics under different dispersal mechanisms, as driven by water density stratification and bathymetry. A: example for an ice contact lake; B: example for a proglacial lake. Adapted from Smith and Ashley (1985) and Ashley (1995).

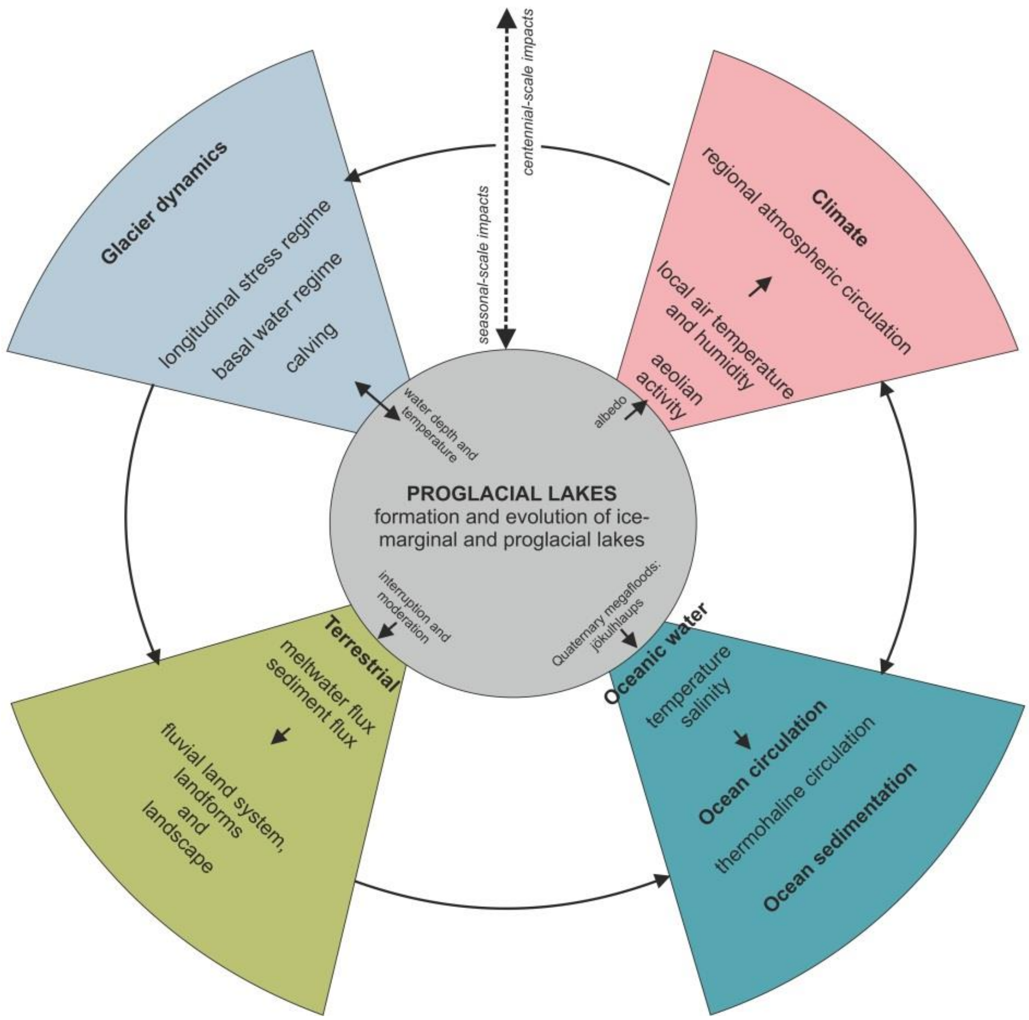

Figure 6: Summary schematic diagram of the influence of proglacial lakes on climatic, oceanic, terrestrial and glaciological processes.

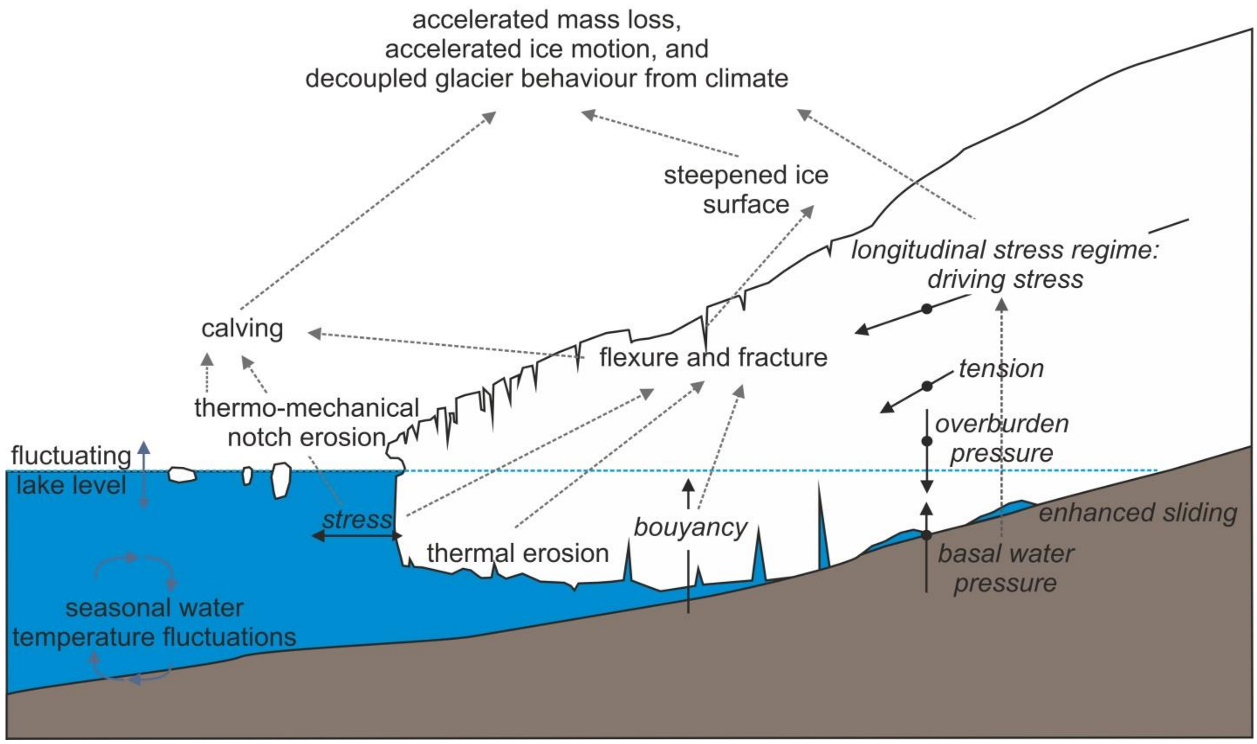

Figure 7: Summary schematic of the influence of an ice-marginal lake on glacier dynamics. Forces are in italicised text and with black arrows. Processes are in black text and with grey dashed lines to denote interactions.