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On the Reconstruction of Palaeo-Ice Sheets: Recent Advances and Future Challenges

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Abstract:

Reconstructing the growth and decay of palaeo-ice sheets is critical to understanding mechanisms of global climate change and associated sea-level fluctuations in the past, present and future. The significance of palaeo-ice sheets is further underlined by the broad range of disciplines concerned with reconstructing their behaviour, many of which have undergone a rapid expansion since the 1980s. In particular, there has been a major increase in the size and qualitative diversity of empirical data used to reconstruct and date ice sheets, and major improvements in our ability to simulate their dynamics in numerical ice sheet models. These developments have made it increasingly necessary to forge interdisciplinary links between sub-disciplines and to link numerical modelling with observations and dating of proxy records. The aim of this paper is to evaluate recent developments in the methods used to reconstruct ice sheets and outline some key challenges that remain, with an emphasis on how future work might integrate terrestrial and marine evidence together with numerical modelling. Our focus is on pan-ice sheet reconstructions of the last deglaciation, but regional case studies are used to illustrate methodological achievements, challenges and opportunities. Whilst various disciplines have made important progress in our understanding of ice-sheet dynamics, it is clear that data-model integration remains under-used, and that uncertainties remain poorly quantified in both empirically-based and numerical ice-sheet reconstructions. The representation of past climate will continue to be the largest source of uncertainty for numerical modelling. As such, palaeo-observations are critical to constrain and validate modelling. State-of-the-art numerical models will continue to improve both in model resolution and in the breadth of inclusion of relevant processes, thereby enabling more accurate and more direct comparison with the increasing range of palaeo-observations. Thus, the capability is developing to use all relevant palaeo-records to more strongly constrain

deglacial (and to a lesser extent pre-LGM) ice sheet evolution. In working towards that goal, the accurate representation of uncertainties is required for both constraint data and model outputs. Close cooperation between modelling and data-gathering communities is essential to ensure this capability is realised and continues to progress.

1. Introduction

The first paper published in the newly-launched Quaternary Science Reviews in 1982 was 'On the Reconstruction of Pleistocene Ice Sheets: A Review' by John T. Andrews. His paper highlighted a range of topics in Quaternary science that require accurate reconstructions of the area, volume and chronology of palaeo-ice sheets (e.g. global sea level, marine oxygen isotopes, plant and animal migrations, glacial stratigraphy and chronology), but noted that such reconstructions were difficult to produce, and often tackled separately by sub-disciplines. The evidence used to reconstruct palaeo-ice sheets was reviewed, and he emphasised the need to reconcile empirical evidence with results from numerical ice sheet models, which were pioneering at the time (e.g. Mahaffy, 1976; Budd and Smith, 1981; Denton and Hughes, 1981), but still very much in their infancy. An important discussion focussed on the mismatch between the over-simplified single-domed ice sheets generated by modelling and the more dynamic and multi-domed configurations inferred from glacial geological evidence (e.g. Shilts, 1980). Andrews (1982: p. 26) concluded that the future of ice sheet reconstructions "does not rest with a single approach, but with a multiple approach of reconstructions based on all available field and theoretical data".

Over the last few decades, palaeo-ice sheets have also assumed increasing importance as analogues for assessing recent changes observed in modern ice sheets. The Antarctic and

Greenland Ice Sheets have, overall, continued to retreat since the early Holocene (Anderson et al., 2002; Alley et al., 2010), and the rate of mass loss has increased in recent decades in response to both oceanic and climatic warming (Shepherd et al., 2012). However, observations of present-day ice sheets are often restricted to a few decades and there is a need to understand the longer-term significance of their recent changes. Numerous workers have, therefore, recognised the potential that palaeo-ice sheets offer in terms of assessing the duration and magnitude of ice sheet instabilities that could occur (e.g. Kleman and Applegate, 2014) and their contributions to sea-level rise (Carlson and Clark, 2012). Moreover, our confidence in future predictions of ice sheet mass balance and sea level rise will benefit from numerical ice sheet models that have been rigorously tested against palaeo-data.

Since Andrews' pioneering (1982) review, glaciological numerical modelling of palaeo-ice sheets has evolved from hand-tuning models against a few constraints to Bayesian methodologies involving thousands of observational constraints and dozens of calibrated model parameters (Tarasov et al., 2012). In parallel with numerical model development, there has been a rapid expansion in the size and qualitative diversity of datasets for constraining ice sheet models (e.g. glacial geological records from both onshore and offshore, relative sea-level records, glacio-isostatic data, ocean sediment records, etc.). Thus, it has become increasingly necessary to forge interdisciplinary collaboration between sub-disciplines concerned with ice sheet reconstructions and, in particular, to link numerical modelling with observations from both terrestrial and marine records. This has been one of the goals of the international MOCA project (Meltwater routing and Ocean-Cryosphere-Atmosphere response: www.physics.mun.ca/~lev/MOCA), previously funded as a joint network of INQUA PALCOM (Paleoclimate) and TERPRO (Terrestrial processes) commissions. MOCA workshops have enabled modellers and empiricists to collaborate and elucidate what goes into their reconstructions and the associated uncertainties. This synthesis has helped to identify new constraints and scrutinize hidden assumptions, with the aim of providing more robust ice sheet reconstructions. Workshops have also highlighted recent advances in the techniques and methods to reconstruct ice sheets, identified the remaining challenges, and also illuminated directions for future research. These methodological insights motivate this paper. Our aims are to highlight some important advances in reconstructing palaeo-ice sheets from: (i) glacial geological evidence in terrestrial and marine settings (Section 2); (ii) improvements in dating methods and approaches (Section 3); and (iii), numerical modelling (Section 4). Note that we do not cover every aspect of reconstructing palaeo-ice sheets, but focus on methods that are primarily targeted at constraining the extent, thickness and dynamics of ice sheets at the regional to continental scale (Figure 1). We then discuss some key challenges for future work (Section 5), emphasizing the need to link terrestrial and marine records with numerical modelling. In doing so, we evaluate the extent to which Andrews' (1982) invocation for interdisciplinary ice sheet reconstructions has been met.

2. Advances Based on Glacial Geological Observations in Terrestrial and Marine Settings

2.1. Formalising Glacial Inversion Techniques Using Ice Sheet Flow-Sets

Subglacial processes beneath palaeo-ice sheets created and preserved landforms that are the basic ingredients for reconstructing their flow patterns and extent. The technique of inverting the bed imprint to extract ice sheet properties is known as 'glacial inversion' (Kleman and Borgström, 1996; Kleman et al., 2006). The burgeoning availability of aerial photography, remote sensing and Geographical Information Systems (GIS) in the latter half of the 20th century permitted the systematic mapping of ice sheet beds that yielded benchmark reconstructions of the last mid-latitude ice sheets (e.g. Boulton et al., 1985; Dyke and Prest, 1987). The prevailing paradigm was that the mapped ice flow indicators (e.g. drumlins) formed in a radial pattern close to the ice margin and mostly during deglaciation (e.g. Boulton et al., 1985). As such, these reconstructions often depicted a radial pattern of flow from one or more relatively stationary ice domes (Figure 2a, b). This paradigm was challenged when Boulton and Clark (1990a, b) recognised that the Laurentide Ice Sheet (LIS) bed was a 'palimpsest' (relict) landscape of flow patterns of different ages that were not all formed in an ice marginal environment (Figure 3). They grouped coherent patterns of glacial lineations into discrete mapped units, termed 'flow-sets' (Figure 3), and noted how they typically cut across each other, allowing them to assign their relative age based on principles of superimposition. In contrast to previous reconstructions (Figure 2a), they revealed a highly mobile ice sheet with ice divides and dispersal centres shifting by 1000–2000 km during the last glacial cycle, as hinted at in previous records of erratic dispersal trains (Shilts, 1980).

The identification of cross-cutting flow-sets permits detailed reconstructions of changes in ice sheet flow patterns, but flow-sets can also preserve important information with regard to the glaciological conditions that formed and preserved them. Kleman and Borgström (1996) highlighted the importance of the basal thermal regime and subglacial hydrology when deciphering the glacio-dynamic context of flow-set formation, and identified several different fan (flow-set) types that could be used to create a time-slice sequence of ice sheet evolution (e.g., those formed during a surge, or those formed during warm-based deglaciation). An exemplar of this inversion methodology is provided in a reconstruction of the Fennoscandian Ice Sheet (FIS) from 115 to 9 ka (Kleman et al., 1997). Similar to the LIS, their reconstruction depicted ice sheet configurations with shifting ice divides during build-up and decay (Figure 4).

Implicit in these reconstructions of the LIS (Boulton and Clark, 1990a, b) and FIS (Kleman et al., 1997) is that some flow-sets record ice flow patterns and properties that date to the last glacial cycle, but predate the Last Glacial Maximum (LGM) and were protected by subsequent cold-based ice. These frozen-bed patches span a wide range of spatial scales that can be used to glean information about ice sheet dynamics and configuration (Kleman and Glasser, 2007). In places, they occur as a mosaic of isolated patches or 'islands' in upland landscapes, but it is also clear that large frozen-bed areas are prevalent at the ice sheet scale (Kleman and Glasser, 2007). Thus, ice directional indicators from earlier stages in the last glacial cycle can be pieced together to reconstruct the inception and build-up of ice sheets prior to the LGM (Kleman et al., 1997). Compared to the record of deglaciation, pre-LGM ice flow indicators and associated flow-sets are obviously more fragmentary, but they can be integrated with other evidence (e.g. till stratigraphy, chronological data) to provide tentative reconstructions of minimal ice sheet outlines and dispersal centres that provide potentially powerful constraints for numerical modelling of ice sheet build-up (Kleman et al., 2010; Stokes et al., 2012).

Similar inversion methods have been used to reconstruct the pattern and timing of the British-Irish Ice Sheet (BIIS) from ~30 to 15 ka (Clark et al., 2012). A key conclusion was that different sectors of the BIIS reached their maximum positions at different times, and that the initiation of ice streaming and calving may have been an important factor in explaining the retreat of marine-based sectors. Clark et al. (2012) compared their reconstruction with numerical modelling of the ice sheet (Boulton and Hagdorn, 2006; Hubbard et al., 2009) and found that, whilst there were broad similarities in terms of the extent and the position of ice divides, there were marked discrepancies in the timing of maximum extent and retreat phases, which they suggested may be due to the palaeo-climate forcing, interpreted from the Greenland ice core record. They highlighted the need for further data-model integration, and an improved understanding of both calving dynamics and the links between interior thinning and marginal retreat.

One deficiency in the glacial inversion method (Section 2.1) is that our process understanding of how various landforms are produced remains somewhat limited. However, recent advances in geophysical observations have allowed bedforms to be imaged beneath existing ice sheets at depths of up to 2 km below the ice surface (King et al., 2007; 2009; Smith et al., 2007). Specifically, drumlins appear to be recorded in the onset zone of Rutford Ice Stream, West Antarctica, where ice velocities accelerate from 72 to >200 m a^{-1} in the transition from slower ice sheet flow to more rapid stream flow (King et al., 2007). Further down the ice stream, where velocities increase to around 375 m a^{-1}, King et al. (2009) reported the presence of mega-scale glacial lineations, which are indistinguishable from those reported on palaeo-ice stream beds. These observations confirm the inference from palaeo-glaciology that bedform elongation is related to ice velocity (Clark, 1993).

In summary, new remote sensing and GIS products have led to major advances in our ability to map changing flow patterns on palaeo-ice sheet beds. Glacial inversion methods have allowed workers to develop a more formal and explicit methodology to invert the bed record. The resultant reconstructions have allowed us to decipher major changes in ice sheet basal thermal regime and ice divide location and migration, both during build-up and decay phases. These empirically-derived reconstructions are an improvement on their predecessors, but there remain some important discrepancies when compared to numerical modelling and secure links between landforms and palaeo-ice dynamics (velocity, thickness, basal shear stress) are difficult.

2.2. Identification of Palaeo-Ice Streams

Given the importance of rapidly-flowing ice streams to ice sheet mass balance and stability, ice sheet reconstructions that omit their activity are likely to be deficient (Stokes and Clark, 2001). The dynamic behaviour of palaeo-ice sheets has often been linked to the activity of ice streams (Denton and Hughes, 1981; Kleman et al., 1997; Hemming, 2004; Clark et al., 2012), and it has been known for some time that ice streaming is the most likely explanation for the low ice surface slopes in the marginal areas of some ice sheets (Fisher et al., 1985). Therefore, it is perhaps surprising that explicit recognition of ice streaming in glacial inversion techniques is a relatively recent development (Kleman and Borgström, 1996; Stokes and Clark, 1999).

Early attempts at incorporating ice streams in large-scale ice sheet reconstructions were largely based on topographic inference (Denton and Hughes, 1981; Dyke and Prest, 1987). Indeed, in relation to Denton and Hughes' (1981) hypothesised ice streams, Andrews (1982: p. 25) noted that "it is not known whether or where ice streams existed within the Laurentide Ice Sheet". However, many of these early inferences were later supported by studies that identified geological evidence of ice stream activity, such as distinctive erratic dispersal trains with abrupt lateral margins (e.g. Dyke and Morris, 1988), convergent flow-set patterns (Punkari, 1995), and highly elongate subglacial bedforms termed mega-scale glacial lineations (Clark, 1993). Based on the evidence of hypothesised ice streams, and an understanding of their activity in present-day ice sheets, several diagnostic geomorphological criteria were proposed to aid their identification (Stokes and Clark, 1999).

These criteria provide an observational template for identifying palaeo-ice streams based on their landform record, but other evidence includes large-scale topographic features (such as cross-shelf troughs and trough mouth fans: Section 2.3). As a result of new satellite and digital elevation products, there has been a huge increase in the number of inferred ice streams (Figure 5), such that we now have a good knowledge of their location in most palaeo-ice sheets (Clark et al., 2012; Livingstone et al., 2012a; Margold et al., 2015). Moreover, recent work has revealed abrupt changes in the trajectory of palaeo-ice streams, known as 'flow switching' (Dowdeswell et al., 2006; Winsborrow et al., 2012), which has also been reported from observations of present-day ice sheets (Conway et al., 2002) and numerical modeling (Payne and Dongelmans, 1997). Thus, reconstructions of palaeo-ice stream activity provide new insights into the long-term behaviour of ice streams (and their potential forcing) and often help reconcile complex cross-cutting flow-set patterns (discussed in Section 2.1) (Dyke and Morris, 1988; Stokes et al., 2009).

Despite much progress identifying soft-bedded ice streams, understanding the evolution of bedrock landforms in both inter-stream and ice stream areas is more challenging, with features probably developing over multiple glacial cycles (Roberts and Long, 2005). Nonetheless, recent progress has been made in identifying the landform assemblage of 'hard-bedded' ice streams, with rock drumlins, mega-flutes and mega-lineated terrain interpreted to reflect accelerated abrasion and quarrying of bedrock under rapidly-flowing ice (Eyles, 2012).

The marine geological record also preserves evidence of palaeo-ice streaming in the form of episodic layers of ice-rafted detritus (IRD), the most conspicuous of which were deposited during Heinrich events (Heinrich, 1988; Hemming, 2004; Section 2.7). Andrews and Tedesco (1992) attributed the carbonate-rich IRD layers associated with the two most recent Heinrich events (H1 and H2) to a source area eroded by an ice stream in Hudson Strait (Andrews and MacLean, 2003). Ironically, terrestrial glacial geologic evidence for the existence of this ice stream is relatively scarce compared to other ice streams (Margold et al., 2015), but it is often implicated in Heinrich events, despite uncertainty over the precise mechanisms through which debris becomes entrained within icebergs and subsequently released (Hemming, 2004). That said, numerical modelling has been able to reproduce the episodic activity of this ice stream at the appropriate time-scales (MacAyeal, 1993; Marshall and Clarke, 1997; Calov et al., 2010), and coupled iceberg transport/melt and climate modelling can generate IRD layers of sufficient thickness (Roberts et al., 2014), assuming relatively small (0.04 Sv over 500 years) iceberg discharge from Hudson Strait. The cause of Heinrich events, however, remains open to debate (see Section 2.7).

Recent advances in sediment provenance techniques highlight the potential to further constrain episodes of ice stream activity using ocean sediment records (Darby and Bischof, 1996; Andrews and Eberl, 2012). Some studies have focussed on potential correlative events from smaller ice streams at the eastern margin of the LIS, such as those draining into Baffin Bay (Andrews et al., 1998; 2012) or the continental shelf off Nova Scotia (Piper and Skene, 1998). Further afield, IRD events have also been detected in the Arctic Ocean (Darby et al., 2002; Darby, 2003) and attributed to source areas in the Canadian Arctic Archipelago (Stokes et al., 2005) and Eurasia (Spielhagen et al., 2004). Although obtaining precise ages for Arctic Ocean IRD events remains difficult because of generally low sedimentation rates, it has been suggested that some events sourced from the Canadian Arctic Archipelago are broadly correlative with the North Atlantic's Heinrich events (Darby et al., 2002). The possibility of large-scale ice sheet reorganisations during Heinrich events has also been hinted at from records of terrestrial ice streams at the southern margin of the LIS (Mooers and Lehr, 1997), but this concept has received little attention. Numerous studies have also examined potential correlations between LIS IRD events and meltwater and IRD events from other mid-latitude ice streams bordering the North Atlantic (Lekens et al., 2006; 2009), although there is much uncertainty about whether IRD events from the eastern LIS have correlatives from other ice sheets (e.g. Fronval et al. 1995; Dowdeswell et al., 1999; Hemming, 2004; Rashid et al., 2012; see also Section 2.7).

To summarise, recent advances have permitted the identification of numerous palaeo-ice stream tracks, and inventories of their location during deglaciation are probably close to complete for some ice sheets (e.g., for the Laurentide Ice Sheet, where >100 ice streams have been identified: Margold et al., 2015). These records attest to major changes in their spatial extent during deglaciation, but there have been few attempts to compare reconstructions of palaeo-ice stream activity with output from numerical ice sheet models. In contrast, modelling has been used to simulate the behaviour of individual ice streams, especially in relation to the IRD record, which provides a valuable tool to explore the history of ice streams prior to the LGM.

2.3. Offshore Geophysical Evidence of Ice Sheet Extent and Dynamics

A major development in ice sheet reconstructions has been the increased use of geophysical techniques (seismic, sidescan sonar, swath bathymetry) to investigate the marginal areas of palaeo-ice sheet beds that are now submerged beneath sea level (Ó Cofaigh, 2012). Imagery from continental slopes has revealed sedimentary depocentres (trough mouth fans: Vorren and Laberg, 1997; Batchelor and Dowdeswell, 2014), the architecture of which often indicates rapid, episodic sedimentation by ice streams (Dowdeswell et al., 1996; Dowdeswell and Elverhoi, 2002; Nygård et al., 2007). Indeed, dating of sediment packages is an important constraint on ice stream activity that is not easily available from terrestrial records, and some marine records extend back through several glacial cycles (Nygård et al., 2007).

Sediment depocentres are often associated with major troughs carved across the continental shelf (e.g. Batchelor and Dowdeswell, 2014). Swath bathymetry data from within these troughs (Figure 6) commonly reveals geomorphology which fulfils the criteria for palaeo-ice streams, similar to those described in terrestrial settings (Anderson et al., 2002; Ottesen et al., 2005, 2008; Livingstone et al., 2012a). Together with sub-bottom profiling and seismic investigations, geophysical techniques have the added advantage of being able to map seafloor morphology and changes in sediment thickness across large areas (Dowdeswell et al., 2004). The seaward extent of glacial till and sub-ice morphology has also been used to constrain the thickness of ice sheets at their marine margins (Polyak et al., 2001; Jakobsson et al., 2010; Dowdeswell et al., 2010; Niessen et al., 2013).

A major contribution of submarine geophysical evidence has been the recognition that palaeo-ice sheets were more extensive than previously thought, with almost all of the mid-latitude Northern Hemisphere ice sheet maximal margins now known to have reached the continental shelf edge, with particular attention focussed on the Arctic Ocean (Figure 7). In the last decade alone, there are cases where margins have been revised and extended to the edge of the continental shelf for the Laurentide and Innuitian Ice Sheets, the Eurasian Ice Sheet (including the Fennoscandian, Barents Sea and British-Irish Ice Sheets), and the Greenland Ice Sheet (e.g. Ottesen et al., 2005; Shaw et al., 2006; Bradwell et al., 2008; England et al., 2009; Ó Cofaigh et al., 2013). In particular, advances in high-latitude seafloor mapping in the 1990s, especially high resolution multibeam mapping, have helped elucidate the glacial history of the Arctic Ocean (reviewed in Jakobsson et al., 2014). A series of investigations have documented evidence for extensive erosion of the Arctic Ocean seafloor caused by ice at modern water depths of up to 1,000 m, as well as glacial landforms on individual ridge crests and plateaus where water depths are shallower (Vogt et al., 1994; Jakobsson, 1999, Polyak et al., 2001). These data, together with chronological information retrieved from sediment cores, have been taken to imply that the most extensive ice shelf complex existed in the Amerasian basin of the Arctic Ocean during Marine Isotope Stage (MIS) 6 (Jakobsson et al., 2010; 2014).

Large seabed scour marks have also been reported at modern water depths of ~600 m along the northern Svalbard margin and across the Yermak Plateau (Vogt et al., 1994; Dowdeswell et al., 2010) as well as at depths of ~1,000 m on Morris Jesup Rise (Jakobsson et al., 2010) (Figure 7). Similar scour marks and, in places, glacial bedforms such as flutings and mega-scale glacial lineations have also been used to invoke the presence of extensive and thick ice sheets and ice shelves along the Chukchi Borderland (Jakobsson et al., 2005; Dove et al., 2014) and southern Mendeleev Ridge (Niessen et al., 2013), which require revisions of glacial Arctic ice cover (Jakobsson et al., 2014). More localised ice shelves have also been hypothesised, often in association with ice streams (e.g. Hodgson, 1994).

Taken together, geophysical evidence of ice sheet and ice shelf extent across continental shelves and submarine ridges has led to major advances in our understanding of the dynamics of palaeo-ice sheet margins, which numerical modelling can now target. Reconstructions of palaeo-ice shelves have seen some major advances, but they are more difficult to constrain, especially in terms of their timing and spatial extent (see also Section 4.3).

2.4. Subglacial Hydrology of Ice Sheets and Subglacial Lakes

Glacial geological features relating to subglacial hydrology can provide useful information relating to, for example, ice sheet basal thermal regime and the distribution and drainage of water at the bed. Eskers are particularly useful in glacial inversion techniques (Section 2.1) and are generally thought to form time-transgressively within a few tens of kilometres of the retreating ice margin (Kleman et al., 1997, 2006, 2010; Mäkinen, 2003; Storrar et al., 2014a). Clark and Walder (1994) showed that eskers from the LIS were concentrated over the impermeable crystalline bedrock of the Canadian Shield, where subglacial meltwater was more likely to carve R-channels into the base of the ice. In contrast, over softer permeable sediments, meltwater was more likely to drain into and across the till, carving much shallower channels or ‘canals’ in a more distributed system (Sjogren et al., 2002). This broad pattern is also seen in the FIS (Boulton et al., 2009) and, while eskers can clearly form over softer sedimentary beds, they are typically less common and depict a more chaotic and fragmentary pattern (Shilts et al., 1987; Storrar et al., 2014a). More recently, Storrar et al. (2014b) showed that the number of eskers increased during deglaciation of the LIS across the Canadian Shield, coinciding with increased rates of ice margin retreat during climatic warming. This is reminiscent of the seasonal evolution of drainage systems in much smaller valley glaciers (Hubbard and Nienow, 1997) and implies that drainage systems (and therefore subglacial lubrication) evolve over millennial time-scales.

An important advance in our understanding of the subglacial hydrological system beneath ice sheets has been the discovery of numerous subglacial lakes. They were first identified beneath the Antarctic Ice Sheets in the 1960s (Robin et al. 1970), but several hundred have now been detected (see Wright and Siegert, 2011). They are thought to be an active component of the subglacial hydrological system, with the potential to fill and drain, and exert an important influence on ice dynamics (Fricker et al., 2007; Smith B.E. et al., 2009; Wright and Siegert, 2011).

Until recently, subglacial lakes had mainly been investigated under present-day ice sheets and examples of putative palaeo-subglacial lakes were rare (e.g. McCabe and Ó Cofaigh, 1994; Munro-Stasiuk, 2003; Christoffersen et al., 2008). This is, perhaps, surprising because access to the sediments and landforms associated with palaeo-subglacial lakes is relatively easy compared to extant ice sheets. However, despite recent attempts to formulate diagnostic criteria to identify their presence (e.g. Bentley et al., 2011; Livingstone et al., 2012b) it is difficult it to distinguish their geological signature from former proglacial (ice-marginal or ice-fed) lakes. Nonetheless, there is strong theoretical support for their existence (e.g.

Shoemaker, 1991; Livingstone et al., 2012b, 2013) and palaeo-ice surface and bed topographies can be used to calculate 3-D hydraulic potential surface (Shreve, 1972). Evatt et al. (2006) were the first to consider this method at the ice-sheet scale and predicted where subglacial lakes might have formed under the LIS at the LGM. A similar approach was adopted by Livingstone et al. (2013), who used an ensemble of ice-sheet model outputs to explore the likelihood of subglacial lake formation under the Cordilleran, Laurentide and Innuitian ice sheets (Figure 8).

Subglacial lakes can drain rapidly towards the ice margin (as jökulhlaups), sometimes on sub-annual timescales (Smith B.E. et al. 2009). Knowledge of the geomorphological and sedimentological impact of such drainage events is still in its infancy but, for example, the ‘Labyrinth’, an anastomosing network of channels in the Dry Valleys (Antarctica), is thought to have formed by periodic subglacial lake drainage events (Denton and Sugden, 2005). Gravel-boulder outwash fans at the mouth of tunnel valleys in North America have also been related to large magnitude subglacial meltwater outbursts (Cutler et al., 2002). There is also an extensive literature and long history on proposed subglacial (mega-) flood tracts (e.g. Shaw, 2002), which have been linked to large palaeo-subglacial lake outburst events (Shoemaker, 1991, 1999). However, in some cases, the use of subglacial bedforms (e.g. drumlins, ribbed moraine) to delineate flood pathways is more controversial (e.g. Clarke et al. 2005). The pattern of tunnel valleys, subglacial meltwater channels, and other meltwater-related features, are useful for constraining palaeo-subglacial lakes and elucidating how they interact with the subglacial hydrological system. More generally, the imprint of meltwater drainage recorded on the bed of former ice sheets is a potentially useful test of numerical models that predict the configuration of the subglacial hydrological system (e.g. Hewitt, 2011; Werder et al., 2013), but which has generally been under-used.

In summary, meltwater landforms are, perhaps, under-used in terms of understanding the source, routing, and storage of meltwater associated with palaeo-ice sheets. Eskers provide important information about subglacial drainage patterns at the ice sheet scale and how it evolves through time, but meltwater channels carved into bedrock are more enigmatic. Numerical ice sheet models have proved an important tool for identifying the potential location of subglacial lakes, but there have been few attempts to integrate a palaeo-glaciological understanding of subglacial hydrology into numerical modelling of ice sheets.

2.5. Proglacial Hydrology of Ice Sheets and Proglacial Lakes

Ice sheet reconstructions have used ice-marginal meltwater channels, spillways, glacial lake shorelines and deltas, to trace former ice margins and reconstruct ice retreat patterns (Kleman, 1992; Jansson, 2003; Kleman et al., 2006; Greenwood et al., 2007; Margold et al., 2013a, b). These landforms are particularly important in areas of formerly cold-based ice that prevented the formation of glacial lineations or eskers (Kleman, 1992). Improved spatial resolution of satellite imagery and Digital Elevation Models, and better access to high-resolution data in readily available platforms such as Google Earth, have facilitated investigation of proglacial meltwater landforms over large areas (Margold and Jansson, 2012) and the incorporation of glacial lake shorelines and deltas into ice sheet reconstructions (Jansson, 2003; Clark et al., 2012).

In relation to ice marginal meltwater features, it has long been known that large proglacial lakes formed around some margins of palaeo-ice sheets as they retreated (e.g. Upham, 1895; Leverett, 1902). The routing of lake overflows and meltwater is important because of the potential impact on ocean circulation and climate (Rooth, 1982; Teller et al., 2002; Barber et al., 1999; Spielhagen et al., 2004; Mangerud et al., 2004; Peltier et al., 2006; Carlson and

Clark, 2012). However, the discharge and location of water overflowing from these lakes varied as ice retreated from various basins, thereby opening new outlets (Teller and Thorleifson, 1983; Teller, 1987; Smith and Fisher, 1993; Smith, 1994; Mangerud et al., 2001; 2004). Records of pre-LGM proglacial lakes (and drainage) must have been equally complex, but most of that record has been eroded or is fragmentary and buried, and is likely to only be discernible in ocean sediment records (Nicholl et al., 2012).

Despite recent progress in reconstructing the history of proglacial lakes, many challenges remain, especially concerning one of the world's largest ice-marginal lakes fringing the decaying LIS: glacial Lake Agassiz (GLA). The extent of the lake varied through time, covering a total area >1 million km^2 during its ~6,000 year history, and overflowing at various times to the Gulf of Mexico, the North Atlantic, the Labrador Sea (including its final drainage via Hudson Bay), and the Arctic Ocean (Teller, 1987; Teller and Leverington, 2004) (Figure 9). GLA serves as a useful case study because it highlights the potential difficulty of constraining the continental scale drainage re-routings of proglacial lakes.

Its outlet chronology has been reconstructed using: (1) the location of the changing LIS margin (Dyke, 2004; Lowell et al., 2009); (2) the dating of beaches that define the lake's outline (e.g. Teller et al., 2000; Lepper et al., 2013); (3) data-calibrated numerical ice sheet modelling (Tarasov and Peltier, 2006); (4) the palaeo-topography of the basin inferred from isostatic rebound (e.g. Leverington et al., 2000; 2002; Rayburn and Teller, 2007); and (5) the dating of meltwater events in the different outlet channels, and in lakes and oceans beyond the channel mouths (Section 2.6), including the use of deep sea oxygen isotope records and distinctive sediment discharge to identify meltwater pulses (Fisher, 2007; Hillaire-Marcel et al., 2008; Lowell et al., 2009; Lewis et al., 2012; Fisher and Lowell, 2012; Teller, 2013).

Whilst it is generally agreed that GLA initially drained south into the Gulf of Mexico (along the Mississippi River), and finally drained north-east into Hudson Bay (Barber et al., 1999;

Clarke et al., 2004), its drainage history between these two end-points has been the subject of much debate, particularly in association with the abrupt cold reversal of the Younger Dryas (YD). It has been argued that overflow from GLA may have triggered the YD stadial, but there is uncertainty as to whether it overflowed to the east, to the northwest, or did not overflow at all (Lowell et al., 2009; Fisher and Lowell, 2012; Lowell et al., 2013; Teller, 2013). Opinions differ about precise timing and routing of GLA overflow during the YD, because different approaches often yield different interpretations (e.g. deVernal et al., 1996; Tarasov and Peltier, 2005; Carlson et al., 2007; Murton et al., 2010; Not and Hillaire-Marcel, 2012; Cronin et al., 2012; Carlson and Clark, 2012). In a recent review, Carlson and Clark (2012) favoured an easterly route, but numerical modelling suggests a large influx of freshwater into the Arctic Ocean was more likely (although not necessarily from GLA: Tarasov and Peltier, 2005), and would have had a more significant impact on ocean circulation (Condron and Winsor, 2012). Several studies of deep-sea Arctic oxygen isotopes (Spielhagen et al., 2004), faunal assemblages (Hanslik et al., 2010, Taldenkova et al., 2013), and other proxies (reviewed in Carlson and Clark, 2012) provide evidence for significant meltwater discharge through the Mackenzie River at the onset of the YD (see also Section 2.6), as does an OSL-dated sequence at the mouth of the Mackenzie River (Murton et al., 2010). However, Fisher and Lowell (2012) argued that, in the Agassiz basin and headwaters of the Mackenzie River (e.g. near Fort McMurray, Alberta), field data are lacking to support Agassiz water being routed to the northwest at this time.

A major impediment to understanding the history of proglacial lakes (including GLA) has been the lack of dating control on shorelines, although optically stimulated luminescence dating (Section 3.2) has seen success (e.g., Lepper et al., 2013). The dating of lake spillway channels can also help to constrain the timing of lake overflow (e.g. Fisher, 2003; 2007; Fisher et al., 2008). Difficulty remains in obtaining maximum ages for spillways in recently deglaciated areas, and in situations where floods may occur subglacially or cut through stagnant ice (Clarke et al., 2004). Nonetheless, shore-line records can be a powerful constraint for numerical modelling of meltwater runoff from ice sheets (as well as providing further constraint on deglacial ice sheet evolution) in models that explicitly resolve proglacial lakes (see Fig. 9 in Tarasov et al., 2012).

Over and above changes in baseline runoff related to changes in melting of the LIS and precipitation, there were occasional short-term hydrological spikes that may have also impacted on ocean circulation (Rooth, 1982; Barber et al., 1999; Condron and Winsor, 2012). Any catastrophic lake outbursts should also leave a geomorphological imprint, and there are well-documented examples from North America, Fennoscandia and Russia (Bretz, 1923; Teller and Thorleifson, 1983; Baker and Bunker, 1985; Smith and Fisher, 1993; Murton et al., 2010; Margold et al., 2011). In some distal locations, one might also expect to find dateable material, such as regional erosional surfaces associated with lag deposits (e.g. Murton et al., 2010). In places, these surfaces and associated geomorphology may also extend offshore (Blasco et al., 1990).

In summary, ice marginal meltwater landforms (meltwater channels, spillways, shorelines) are an important ingredient for glacial inversion techniques, especially where other evidence is scarce. In recent years they have taken on added importance for identifying the extent of large proglacial lakes and meltwater routing, and their associated impacts on the ocean-climate system. Despite much progress, however, it remains difficult to precisely date marginal drainage or outburst floods or even continental scale drainage re-routings; and so it has often proved difficult to reconcile the terrestrial and marine records of meltwater routing. Numerical modelling offers an important constraint on the likely volumes of water delivered by ice sheets, but the resolution of ice margin chronologies is not yet capable of resolving the precise routing through various spillways and there remains a challenge in deciphering what fraction of a given lake is released via baseline drainage or through major outburst floods.

2.6. Detection of Glacial Lake Outburst Events in Near-Shore Marine Records

Since suggestions that the addition of freshwater from glacial lake outbursts may have disrupted ocean circulation (e.g. Rooth 1982; Broecker et al., 1989), palaeoceanographers have been searching marine sediment records for evidence of freshwater discharge events that might complement terrestrial records (Section 2.5). Their detection is, however, complicated by several factors, most notably marine sediment processes, the proximity of core sites to ice sheet margins (and their associated glacial lakes), and the proxies used to infer hydrologic changes (mainly in salinity and temperature) (de Vernal and Hillaire-Marcel 2006). Along continental margins, close to where these pulses of water entered the ocean, sedimentary and hydrological processes include a mix of turbidity currents, surface and intermediate meltwater plumes and IRD. In more distal open-ocean regions, meltwater plumes or only IRD will be recorded (Ó Cofaigh and Dowdeswell, 2001; Eyles and Lazorek, 2007). The sensitivity of microfaunal and isotopic proxies will also vary with proximity to meltwater sources.

This complexity means that it is often difficult to unequivocally identify glacial lake discharge in open ocean sediment records due to low sediment accumulation rates and uncertainty surrounding oceanographic response. For example, there is clear lithological evidence (red clay layer, spikes in detrital carbonate) for drainage of the final phase of GLA ~ 8.5-8.2 ka (glacial Lake Ojibway: Figure 9) through Hudson Strait (Andrews et al. 1995, 1999, Kerwin 1996, Barber et al. 1999, Hillaire-Marcel et al. 2007; Jennings et al., 2015). The lithological signature is lost to bioturbation in distal settings, but oxygen isotopic evidence for this event extends to the Laurentian Fan and continental slope north of Cape Hatteras (Keigwin et al., 2005). In contrast, in the deep western Labrador Sea there is a detrital carbonate spike but no isotopic signature (Hillaire-Marcel et al., 2007, 2008).

Recent studies of near-shore regions provide more concrete evidence for hydrological signals of abrupt drainage of glacial lake water than those from open ocean sites. In the outer St Lawrence Estuary, there is microfaunal and geochemical evidence for hydrological changes, possibly caused by GLA drainage, near the onset of the Younger Dryas ~ 13 ka (Rodrigues and Vilks, 1994; Keigwin and Jones, 1995, de Vernal et al., 1996; Carlson et al. 2007). In the central St. Lawrence Lowlands and the Lake Champlain Basin to the south (regions that are closer to the LIS margin and mid-continental proglacial lakes), the evidence is even more convincing. Geomorphological evidence from isostatically uplifted lake and marine shorelines and high-resolution (up to ~ 0.1 to 0.8 cm a^{-1}) lacustrine and marine stratigraphic records (Franzi et al., 2007) from glacial Lake Vermont and Champlain Sea sediments provide a nearly-continuous record of LIS retreat from ~13.5 to ~10 ka. Evidence points to six regional lake drainage events from these lakes between 13.3 and 10 ka, originating in the St. Lawrence/Champlain and Ontario-Erie lowlands (Lakes Vermont, Iroquois), modern Lake Huron (Lake Algonquin), and the interior plains (Lake Agassiz) (Rayburn et al., 2005, 2007; Cronin et al., 2008; 2012). These events involved volumes of lake waters ranging from hundreds to thousands of km^3 (Teller et al., 2002) and at least three events correlate with abrupt millennial scale climate events recognized in Greenland ice cores and other proxy records: the Intra-Allerød Oscillation (~13.3 ka), the YD (12.9–13.1 ka), and Preboreal Oscillation (~11.2–11.5 ka) (Rayburn et al. 2011, Katz et al., 2011).

As noted above (Section 2.5), GLA drainage through the Mackenzie River at the onset of the Younger Dryas has been proposed as an alternative route to the St. Lawrence on the basis of modelling (Tarasov and Peltier 2005) and dating of Mackenzie delta sands and upstream gravels and erosional channels (Murton et al. 2010). Here too, there is a contrast between the temporal resolution of central Arctic marine sediment records and those along Arctic continental margins. Sedimentation rates on the Lomonosov, Mendeleev and Northwind Ridges average roughly 0.5-2 cm ka^{-1}, while those on continental margins are 2 to 3 orders of magnitude higher. Consequently, central Arctic isotopic records of meltwater induced hydrological changes are ambiguous (Poore et al. 1999), whereas in more proximal regions of the Chukchi and Beaufort Seas off North America, a number of independent proxy methods provide clearer evidence for deglacial hydrological changes (Andrews and Dunhill 2004, Polyak et al. 2007). There is also geophysical (submarine incised valleys) and stratigraphic evidence on the Chukchi Margin for deglacial ice-rafting and lake drainage, possibly from Alaska's Glacial Lake Noatak, that cannot be accounted for by river discharge or precipitation changes (Hill and Driscoll 2008). Off Siberia, sediments from the Laptev Sea indicate major hydrological changes during the last deglaciation (around 12.9 ka), perhaps from Lena River discharge (Spielhagen et al., 2005, Taldenkova et al., 2013).

In summary, these few examples illustrate the huge potential to utilise a variety of evidence from continental shelves and slopes to constrain ice sheet dynamics and runoff. This includes the integration of submarine geophysical, chronological (e.g. radiocarbon dating, varves), sedimentological (e.g. IRD), microfaunal (e.g. foraminifera, ostracodes dinoflagellates), and geochemical proxies (usually oxygen isotopes). Among the many remaining challenges in detecting specific abrupt discharge events, chronology ranks highest because it remains difficult to distinguish short-lived, catastrophic outbursts, most evident in spikes in stable isotope records, from hydrological changes due to more progressive influx of glacial meltwater (i.e longer-term baseline flows).

2.7. Heinrich Events and Associated Meltwater Plumes

Glacial intervals are characterised by Heinrich events that, sensu stricto, are identified by a sudden increase in the coarse lithic fraction, a dominance of the polar planktonic foraminifera Neogloboquadrina pachyderma(s), and a lowering of inferred sea-surface salinity (Heinrich, 1988; Bond et al., 1993). For the past two decades, these events have highlighted the limits of our understanding with respect to reconstructing ice sheet dynamics and their links to the ocean-climate system (Hemming, 2004).

The original definition of Heinrich layers was based on IRD in the central North Atlantic (Ruddiman 1977; Heinrich, 1988), but their sedimentary signature in the Labrador Sea changes with distance from Hudson Strait, and is significantly different from that observed distally. Close to Hudson Strait, the layers are up to several metres thick. IRD abundance shows high-frequency internal variations within the Heinrich layers, with two abundance maxima at the base and the top separated by cyclic deposits of mud turbidites, meltwater plume deposits (rapid deposition of hemipelagic sediment) and IRD beds (Rashid et al., 2012). Such internal variations are less striking moving southwards, as the thickness of Heinrich layers decreases and IRD abundance shows less variation. The succession of turbidites, meltwater plume deposits and IRD beds are organized as couplets (Figure 10) (Hesse and Khodabakhsh, 1998) and are similar to seasonal cycles recognised in the Bay of Biscay by Zaragosi et al. (2006). Dispersed drop clasts and pellets within the turbidite layers suggest that ice calving occurred at the same time as meltwater supply. High resolution sedimentary records from the Bay of Biscay also show a similar pattern for meltwater events, with turbidites attributed to meltwater during warm periods, and IRD beds topping the turbidites during cold periods (Zaragosi et al., 2006; Toucanne et al., 2009; Roger et al., 2013). Thus, Heinrich events appear to involve seasonal meltwater discharge and iceberg calving, similar to that seen at modern glacier outlets.

The volumetric importance of meltwater plume deposits in proximal Heinrich layers means that Heinrich events cannot be interpreted only in terms of iceberg supply and rates of iceberg melting (Andrews and MacLean, 2003; Roche et al., 2004). Indeed, the IRD layers may represent only a small portion of the Heinrich events (Figure 10), and the proportion varies between different events, implying variability in the proportion of icebergs to meltwater (Rashid and Piper, 2007). For example, Heinrich event 3 (H3) is the most prominent event in the Labrador Sea for the past 40 ka, but IRD sourced from the Hudson Strait Ice Stream is barely recorded in the associated Heinrich layer.

The geographic distribution of IRD depends not only on supply, but also on iceberg melting rates and transport distance, which is further controlled by oceanic and atmospheric temperatures, sea ice, winds, and ocean currents. Because the volume of IRD is not representative of the event, IRD layers from the North Atlantic cannot be used as the sole proxy for rapid and extreme calving events (Rashid et al., 2012). Furthermore, Andrews et al. (2012) have been able to identify at least two sources for the Heinrich layers in the Labrador Sea based on source rock signatures. The complex internal structure of Heinrich events as well as the variability in the sediment source (Tripsanas and Piper, 2008b; Rashid et al., 2012; Roger et al., 2013) indicates a complex series of ice stream catchments and tributary redistribution for each event, which is consistent with the activation of a number of different ice streams (see Section 2.2).

Because meltwater pulses were associated with high sedimentation rates, long stratigraphic records are difficult to obtain close to ice stream outlets. Nevertheless, similar patterns from the Eastern Canadian margin and European margin are inferred from 14 to 45 ka (Figure 11). Sedimentary records suggest that meltwater pulses started shortly after or around H4 (Lekens et al., 2006; Toucanne et al., 2009). The amount of meltwater increased toward the LGM, with a climax from 15 to 28 ka (Figure 11) (Lekens et al., 2006; Haapaniemi et al., 2010;

Roger et al., 2013). Where ice persisted, sedimentation rates remained high until the beginning of the Holocene (Piper et al., 2007). Meltwater pulses did not occur randomly, and sometimes occurred in cycles that may be related to millennial-scale variability (e.g. Dansgaard-Oeschger (DO) cycles). These are recorded in turbidite records, particularly in the North Sea fan, the Orphan Basin, and the Laurentian fan, where thick sequences of turbidites were interrupted every 1–2 ka by hemipelagic sediments (Lekens et al., 2006; Piper et al., 2007; Tripsanas and Piper, 2008b; Roger et al., 2013). Recent work, however, suggests that the existence of individual D-O oscillations does not rely upon meltwater outputs to force them, but may arise from self-sustained nonlinear oscillations of the coupled atmosphere-ocean-sea ice system that are 'kicked' into action by preceding Heinrich events (Peltier and Vettoretti, 2014).

On both the European and eastern Canadian margins, a drop in the meltwater input is inferred to have begun around 20-23 ka, and glacigenic debris flows related to readvance of ice streams were common (Lekens et al., 2006; Tripsanas and Piper, 2008a). These readvances coincided with weak Atlantic Meridional Overturning Circulation (AMOC) between 19–23 ka (van Meerbeeck et al., 2011). More generally, the timing and amplitude of the meltwater pulses coincides with major changes in the AMOC during the last glacial cycle (van Meerbeeck et al., 2011).

Despite recent advances, there is no clear consensus about the timing and amount of meltwater and sediment flux during Heinrich events (Alley and MacAyeal, 1994; Johnson and Lauritzen, 1995; Hemming, 2004; Hulbe et al., 2004; Marshall and Koutnik, 2006; Marcott et al., 2011). Insufficient data have been collected near the different glacial outlets (Section 2.6), so that use of marine records has overemphasised Heinrich events and the post-Heinrich event 1 deglaciation. Freshwater flux is difficult to quantify and is not necessarily correlated with IRD flux (the latter being easier to quantify from sediment cores). This difficulty is well illustrated with the multiple attempts at modelling Heinrich events (Hemming, 2004; Marshall and Koutnik, 2006; Roberts et al,. 2014). The data show that iceberg calving represented only part, and in the case of H3 a very small part, of a full Heinrich event; and the freshwater was largely provided by seasonal meltwater discharge. This aspect of the Heinrich events is critical because it can help test numerical modelling. For example, the 'Binge and Purge' model from MacAyeal (1993) is able to reproduce the IRD layers at the base and top of the Heinrich events, but fails to generate seasonal meltwater deposits. The jökulhlaup model from Johnson and Lauritzen (1995) is able to reproduce large amounts of meltwater, but because of its catastrophic approach, iceberg calving and meltwater outburst are not distinct processes and occur over too short a time compared to a typical Heinrich event. Models that assume that Heinrich events were the result of ice-stream reaction to oceanic forcing (Marcott et al., 2011; Alvarez-Solas et al., 2012) may not account for the meltwater derived from both Hudson Strait and from other coastal regions throughout Heinrich events, but rather emphasise processes that produce abundant icebergs from a Hudson Strait Ice Stream and ice shelf. Some glaciological models driven by climate forcing (Marshall and Koutnik, 2006) obtain results very close to what is observed in the sediment record, as they are able to differentiate meltwater input from coastal regions, ice sheets, and Heinrich events from the LIS. Nonetheless, such models encounter difficulties with the Heinrich events, because they restrict the events to ice calving events and therefore overestimate the IRD flux and miss the rhythmic layers deposited from meltwater.

In summary, recent work on sediment sources have shown a great variability even within the same ice sheet, suggesting a complex redistribution of ice stream and tributaries for ice calving and/or meltwater events. Sedimentary records suggest involvement of seasonal meltwater discharge and iceberg calving during ice sheet collapse. Modelling direct meltwater inputs into deep water have been made (discussed in Section 4.6), but the volumetric importance of such flows remains to be assessed from sedimentary record. Insufficient data have been collected near palaeoglacial outlets and, therefore, the timing, duration and significance of seasonal meltwater pulses during ice sheet collapse remain to be assessed.

3. Recent Advances in Dating Ice Sheet Extent

The value of terrestrial glacial geological evidence described above is increased if it can be dated. This section highlights recent advances in the application of cosmogenic nuclide dating (Section 3.1), luminescence dating (Section 3.2) and radiocarbon dating (Section 3.3) to ice sheet reconstructions. A recent in-depth review of dating methods, specific to the Arctic region, can be found in Alexanderson et al. (2014).

3.1. Cosmogenic dating

Cosmogenic nuclide dating has developed into an established chronological tool for ice sheet reconstructions (Bierman, 2007; Balco, 2011). It has enabled direct exposure dating of glacial landforms and deposits, and the number of studies applying cosmogenic nuclide dating for reconstructions of ice sheets has grown rapidly over the last decade. These studies tend to focus on: (i) exposure dating of ice sheet extent (both laterally and vertically) and (ii) determining the effect of subglacial erosion and preservation.

Several recent studies have focussed on constraining the extent and thickness of extant ice sheets in Greenland (Roberts et al., 2008; Briner et al., 2014; Young et al., 2013) and Antarctica (Stone et al., 2003; Bentley et al., 2010; Mackintosh et al., 2011) since the LGM. Similarly, studies of the last mid-latitude ice sheets have tended to focus on changes since the

LGM (e.g. Rinterknecht et al., 2006; Balco and Schaefer, 2006; Stroeven et al., 2010, 2011), but dating of more extensive pre-LGM glaciations has been applied to the north-eastern FIS (Linge et al., 2006), the northern Cordilleran Ice Sheet (Ward et al., 2007; Stroeven et al., 2014) and the Patagonia Ice Sheet (Kaplan et al., 2007; Darvill et al., in press).

Cosmogenic dating has limitations that are closely related to geomorphological uncertainties. An ideal sample for dating the deglaciation of an ice sheet (bedrock or boulders exposed after deglaciation) has had no exposure to cosmic rays prior to glaciation (no inheritance) and full exposure to cosmic rays (no shielding) after deglaciation (Heyman et al., 2011). Often, however, at least one of these requirements cannot be met, resulting in exposure ages that can be either younger or older than the actual deglaciation age. If a surface has been exposed to cosmic rays before the last glaciation, experienced limited or no glacial erosion while ice covered, and full exposure after deglaciation, the surface will yield exposure ages that are older than the deglaciation age due to prior exposure. If, on the other hand, a surface was not exposed before the last glaciation and has only been exposed during a part of the post-glacial time, the surface will yield exposure ages that are younger than the deglaciation age due to incomplete exposure. Several samples from a surface with an expected distinct deglaciation age commonly display scattered exposure ages indicating that the problem with prior and/or incomplete exposure is common (Putkonen and Swanson, 2003; Kaplan et al., 2007; Heyman et al., 2011).

To address the issue of prior and incomplete exposure, several strategies have been applied to interpret sets of exposure ages; firstly, scattered exposure ages can be interpreted as a result of prior exposure with the youngest exposure age of the group interpreted as closest to the actual deglaciation age. Secondly, scattered exposure ages can be interpreted as a result of incomplete exposure with the oldest exposure age closest to the actual deglaciation age. Thirdly, the average of a set of exposure ages can be taken as the deglaciation age, based on the assumption (typically implicitly) that the effects of prior exposure and incomplete exposure will yield equally large errors of opposite character (too old and too young). Fourthly, a scatter in a set of exposure ages can be evaluated using numerical modelling of geomorphic processes to identify the most likely deglaciation age (Applegate et al., 2012). Finally, a statistically robust approach is to only use groups of exposure ages that are so well clustered that the scatter can be explained by measurement error alone, and to accept the mean exposure age of that group as the most likely deglaciation age (Balco, 2011). An example of how exposure age interpretations can lead to different ice sheet reconstructions is the deglaciation of the southern FIS. Giving preference to either the average or the older exposure ages has led to reconstructions diverging with up to some thousand years (Rinterknecht et al., 2006; Houmark-Nielsen et al., 2012).

When glacial erosion has been limited, and nuclides accumulated during a previous period of exposure remain, differing nuclide decay rates of multiple radiogenic isotopes can be used to quantify burial durations under ice (Fabel et al., 2002; Stroeven et al., 2002). For quartz minerals, pairs of ^{10}Be (half-life 1.4 Ma) and ^{26}Al (half-life 0.7 Ma) have been used to infer burial durations of hundreds of thousands of years (Bierman et al., 1999; Fabel et al., 2002; Stroeven et al., 2002). This information helps constrain the glacial erosion and quantify the cumulative duration of ice coverage over multiple glacial cycles. For quantifying shorter burial events, a promising nuclide is in-situ produced ^{14}C, which has a half-life of only 5,700 years, and which therefore decays rapidly enough to be significantly altered when covered by ice for just a few thousand years. Miller et al. (2006) and Briner et al. (2014) used ^{14}C, ^{10}Be, and ^{26}Al, to show the limited glacial erosion on upland Baffin Island and to quantify the duration of Holocene ice cap coverage.

Cosmogenic exposure dating is rapidly evolving with refinements in both measurement and calculation techniques (Balco, 2011). An important improvement in measurement accuracy was achieved for the most commonly used nuclide, ^{10}Be, when Nishiizumi et al. (2007) accurately determined Be ratios of several different standards used to measure ^{10}Be concentrations. An outcome of this study was that reported ^{10}Be concentrations from earlier measurements should be adjusted by up to 17% (Nishiizumi et al., 2007). The production of cosmogenic nuclides is also being re-evaluated. Several recent ^{10}Be production rate calibration studies have reported 5-15% lower reference ^{10}Be production rates (e.g. Balco et al., 2009; Young et al., 2013; Heyman, 2014) compared to the original CRONUS production rates (e.g. Balco et al., 2008), implying that many exposure ages are likely to be thousands of years older than previously reported. For example, Ballantyne and Stone (2012) recalculated the exposure ages of 22 boulders from moraines in northwestern Scotland based on new locally calibrated ^{10}Be production rates, increasing the original mean exposure age by 6.5-12%, and suggesting that the ice retreated much earlier and did not persist throughout the Lateglacial Interstadial. Furthermore, in a recent paper from the CRONUS-Earth project, the production rates have been evaluated for ^{10}Be, ^{26}Al, ^{3}He, ^{36}Cl, and ^{14}C (Borchers et al., in press) for an updated CRONUScalc online exposure age calculator. The geographical (and temporal) scaling of production rates has also seen recent advances, with Monte Carlo simulations originally aimed to estimate exposure to cosmic ray flux for aircrew during flights (Sato et al., 2008) being developed into a model for cosmogenic production scaling on Earth (Lifton et al., 2014).In summary, cosmogenic dating has enabled direct dating of glacial landform surfaces and has led to several major advances regarding palaeo-ice sheet history and dynamics, including dating of ice sheet margins and verification of surface preservation under non-erosive ice sheets. The method is now widely used and will continue to develop into an established chronological tool for ice sheet reconstructions with further calibration (e.g., with AMS C-14) and refinements to production rates and scaling (Bierman, 2007; Balco, 2011). A key point regarding cosmogenic dating is that exposure ages are only correct if all geomorphological uncertainties are understood and correctly addressed; when that is not the case, exposure ages may lead to erroneous chronological constraints.

3.2. Luminescence dating

The most commonly used approaches in luminescence dating are Thermo-Luminescence (TL, using either quartz or feldspar), Optically Stimulated Luminescence (OSL, quartz), and Infrared Stimulated Luminescence (IRSL, feldspar). TL dating is now rarely used, but both OSL and IRSL are frequently applied in glacial geological reconstructions. The advantage of OSL is that the signal is known to bleach rapidly when quartz grains are exposed to light and to be stable for millions of years once shielded. However, some quartz has been unsuitable for OSL dating due to either low signal levels (Preusser et al., 2006) or signal instabilities (Steffen et al., 2009). In such contexts, feldspar IRSL often provides an alternative, although this signal is known to suffer from anomalous fading, causing age underestimation (Wintle, 1973; Huntley and Lamothe, 2001). To circumvent the problem of fading, new approaches involve using a thermally-assisted IRSL measurement (post-IR IRSL; Thomsen et al. 2008; Buylaert et al., 2012). However, this signal is known to bleach less rapidly and can cause age-overestimation in proglacial environments (Blomdin et al., 2012; Lowick et al., 2012). An advantage of IRSL is the higher saturation level, which allows dating further back in time than OSL.

Initial attempts to date proglacial sediments using TL had little success due to problems with incomplete bleaching of the signal prior to deposition (Kronborg, 1983; Jungner, 1983). This problem was also observed in modern proglacial sediments (e.g., Gemmell, 1985; 1994). Even though both OSL and IRSL are much more light-sensitive than TL, early studies using these techniques also struggled with incomplete bleaching when using multi-grain approaches

(Duller, 1994; Rhodes and Pownall, 1994). However, as pointed out by Duller (1994), sediments will likely contain grains with various bleaching histories and levels, including those that experienced complete resetting. The major methodological breakthroughs to date proglacial sediments were, therefore, the introduction of single aliquot and single grain methodologies. In these approaches, several equivalent dose (D_e) measurements from the same sample provide information about the level of bleaching, and different statistical approaches have been proposed to extract the well-bleached fraction from the distribution (Galbraith et al., 1999; Galbraith and Roberts, 2012).

Many studies have stressed the importance of sampling suitable glacial sediment facies (Fuchs and Owen, 2008; Thrasher et al., 2009). For the FIS, for example, OSL ages have been produced for a wide variety of deposits related to ice sheet (de)glaciation (e.g. Alexanderson and Murray, 2012a; Johnsen et al., 2012), although they typically contain ages that overestimate the expected age range, presumably due to incomplete bleaching. Several recent studies have investigated the luminescence properties of quartz and feldspar from modern depositional environments, thereby testing the significance of incomplete bleaching in different glacial settings (Alexanderson, 2007; Alexanderson and Murray, 2012b). Alexanderson and Murray (2012b) showed that the risk for incompletely bleached grains is largest in subglacial till and proximal glaciofluvial sediment, and least in distal glaciofluvial and lacustrine sediment.

OSL ages for sediment associated with the Eurasian Ice Sheet seem to be in relatively good agreement with geological interpretations and radiocarbon ages (e.g. Mangerud et al. 2001; Murray et al., 2007), and some studies have successfully dated sediments older than the LGM. For example, Mangerud et al. (2001) sampled beach and shoreface deposits from Glacial Lake Komi, northwestern Russia. All sampled sand was transported and deposited by waves at or close to the palaeo-shoreline, implying that grains were likely to have been exposed to sufficient sunlight to reset OSL. In North America, early work applied TL to both relict and modern proglacial sediment (Berger and Eyles, 1994), whereas more recent work has applied both fading-corrected IRSL and OSL dating on a variety of deposits (mainly glaciolacustrine and postglacial aeolian) associated with proglacial lakes (Lepper et al., 2013) and the deglaciation of the Laurentide (Balescu et al., 2001) and the Cordilleran ice sheets (Demuro et al., 2012).

In summary, OSL and IRSL dating are currently the most reliable methods to determine depositional ages of glaciofluvial and aeolian sediments related to deglaciation. They are often complementary to cosmogenic and radiocarbon dating in underpinning palaeoglaciological reconstructions, but incomplete bleaching remains an issue in some glacial sedimentary environments and they typically have much larger error bounds compared to radiocarbon dating.

3.3. Radiocarbon Dating and Pan-Ice Sheet Margin Chronologies

For many years, radiocarbon dating has been used to determine ice margin chronologies, with several hundred dates available for most ice sheets (e.g. Hughes et al., 2011) and thousands in some cases (Dyke et al., 2003). Accelerator Mass Spectrometry (AMS) dating of terrestrial macro-fossils and marine micro-fossils has greatly enhanced our ability to constrain the timing of events, including ice retreat. In some locations, bulk sample radiocarbon ages of lake-bottom sediments are now accepted as having errors of at least 1-2 ka and up to 10 ka (Grimm et al., 2009). This is due to problems such as dissolved carbonates in water or detrital carbon reworked from carbon-bearing rocks and sediments. In Switzerland, for example, Andree et al. (1986) found an 800-year offset in radiocarbon age between bulk sediment and AMS plant macrofossil ages for the late-glacial interval, while offsets of up to 8 ka occur in basal lake sediments in central North America (Grimm et al., 2009). In another example, the initiation of 13 lakes/bogs across a landscape previously covered by the south-eastern LIS provided a tight cluster of AMS ages between 16 and 15 ka (Peteet et al., 2012), which is 5-9 ka later than the time of ice retreat based upon the extrapolation of bulk chronologies (Dyke et al., 2003), varves (Ridge, 2004), and cosmogenic dates (Balco et al., 2009).

Perhaps the most important advance in terms of radiocarbon dating is the synthesis of dates from across an ice sheet bed to generate pan-ice sheet margin chronologies. Whilst various dating techniques have been deployed in specific localities, only rarely have these been compiled across larger areas. Building on the sequence of maps depicting deglaciation of North America (Dyke and Prest, 1987), Dyke et al. (2003) assembled a chronological database of mainly radiocarbon dates, supplemented with varve and tephra dates, which constrain ice margin positions and shorelines of large glacial lakes. Dates on problematic materials (e.g. marl, freshwater shells, lake sediment with low organic carbon content, marine sediment, bulk samples with probable blended ages, and most deposit feeding molluscs from calcareous substrates) were excluded. Marine shell dates, a major component, were adjusted for regionally variable marine reservoir effects based on a large new set of radiocarbon ages on live-collected, pre-bomb molluscs from Pacific, Arctic, and Atlantic shores. The resultant database contains ~4,000 dates, as well as interpolated margin position maps, which can be readily imported into GIS software for further analysis and integration with numerical modelling. Indeed, the new chronology has provided a powerful constraint for numerical modelling (Tarasov and Peltier, 2004; Tarasov et al., 2012)..

More recently, the DATED project has synthesized the available published evidence for the timing and spatial extent of the Eurasian Ice Sheet (between 25 and 10 ka) (Hughes et al., accepted). All relevant dates for both the growth and decay of the Eurasian Ice Sheet (~5,000 dates) were critically evaluated and entered into a database, together with metadata to interpret them that includes the location, site type, dated material, dating method, sample number, and stratigraphic context and setting. Each date was classified in terms of its stratigraphic context and assessed in terms of likely reliability. To facilitate comparison and create an internally-consistent dataset, all radiocarbon dates were calibrated using the same calibration data, with a uniform reservoir correction for marine samples, and all terrestrial cosmogenic nuclide exposure ages were recalculated using common schemes for production rates and scaling models. The chronological database was coupled to a GIS and relevant glacial geomorphic indicators from the literature were compiled in digital geo-referenced form. Motivated by the requirement of ice sheet models for uncertainty estimates on input data, ice margin isochrones for every 1,000 years between 25 and 10 ka were interpreted based on the spatial observations, together with the classified and calibrated chronological data. Importantly, three ice margin positions were reconstructed for each time slice: a maximum, minimum and most-credible (Figure 12). These capture the end-members of possible ice marginal positions ('error margins') that satisfy the chronological constraints including dating precision, stratigraphic and spatial correlations, and gaps in dating (or ages).

Other on-going assimilations of existing, and newly collected, geological evidence to form the basis of ice sheet reconstructions also include the BRITICE Project for the British-Irish Ice Sheet and the RAISED project for Antarctica (Reconstruction of Antarctic Ice Sheet Deglaciation: see overview in Bentley et al., 2014). The MOCA project is also updating and revising the Dyke et al. (2003) North American chronology in areas where new dates have been obtained, and through the addition of min-max bounding isochrones for each time-slice.

It is clear that much can be gained by synthesising existing dates into pan-ice sheet chronologies, especially in terms of assessing the synchronicity of the maximum extent of different ice sheets and their rates of deglaciation. Pan-ice sheet chronologies with quantified errors also provide a robust test for numerical ice sheet models.

4. Recent Advances Based on Numerical Modelling of Ice Sheets

4.1. Model Resolution, Parameterisation and Uncertainty

Three issues cut across most aspects of modelling complex environmental systems such as palaeo-ice sheets. First, such modelling is subject to limited model grid resolution and the approximation of relevant processes (if not their complete lack of inclusion) due to computational expense. Secondly, the above, along with uncertainties in model inputs, induce significant uncertainties in model outputs that need to be quantified. Thirdly, in order to reduce and quantify these uncertainties, large and diverse observations (palaeo-datasets) are required to constrain the models. However, the quantity and quality of such datasets are generally inadequate to provide complete constraint and suffer from their own uncertainties, especially in the context of palaeo-ice sheet reconstructions.

Ice sheet models are run at the highest grid resolution possible for the given context and available computational resource. For large glacial cycle ensembles of continental ice sheets, this is typically in the range of 20 to 50 km. However, with parallelized models able to efficiently distribute the modelling of a single ice sheet over hundreds of processor cores, model runs down to 5 km grid resolution are now possible (Golledge et al., 2012; Seguinot et al., in review). In contrast, climate models are computationally much more expensive, with current Earth-system Models of Intermediate Complexity (EMICS) running at effective resolutions of ~500 km. As ice sheet models need a climate forcing to evolve, this mismatch in climate model resolution is a major problem for palaeo-ice sheet modelling. Furthermore, even the representation of core dynamical processes that are resolvable at modelled scales (in both climate and ice sheet models) are generally subject to simplification to enable computational tractability within research timescales. For ice sheet modelling, the Shallow-Ice (SIA) and Shallow-Shelf approximations (SSA) are standard reductions of the full 3D stress balance ("Stokes" equation) for ice sheets, streams, and shelves, respectively (Blatter et al., 2011; Kirchner et al., 2011; Schoof and Hewitt, 2013).

Processes that cannot be explicitly resolved according to known physics (due to model grid size and/or computational expense) must therefore be parameterized or ignored. Parameterized processes generally involve poorly-defined parameters whose values must then be constrained through some combination of physical reasoning, observational comparisons, and high-resolution modelling that explicitly resolves the process.

Modelling uncertainties also arise from those present in various inputs to ice sheet models. Some are observational, such as the basal topography of present-day ice sheets (Fretwell et al., 2013). Other uncertainties are related to past conditions such as the basal topography at the time of ice sheet inception (Tarasov et al., 2012). The largest uncertainties concerning inputs to ice sheet models are in the atmospheric and oceanic components of the climate representation (Pollard, 2010). Near-surface temperatures and rates of snow accumulation are the primary controls on regional terrestrial ice thickness, at least for regions that are not streaming (e.g. Seguinot et al., 2014). Ocean temperatures and currents have a strong impact on grounding line stability (influencing ice calving and sub-shelf melt at marine boundaries) and are major sources of uncertainty, although recent advances have been made (see Section 4.4). Basal processes (drag, hydrology, and sediment production/transport/deposition) determine fast flow conditions and also have large uncertainties, due both to their small scales, complex interactions, and limited accessibility for direct scientific study.

The presence of significant uncertainties necessitates a probabilistic approach to reducing and quantifying uncertainty. This is best understood through Bayes Theorem, in which the posterior probability, given a set of constraints, is proportional to the product of the likelihood and the prior probability. The ‘prior’ is the initial probability distribution for a set of poorly-defined model parameters and inputs. The likelihood function specifies the probability of a value of model output being in agreement with a set of observations (which were not used to generate the prior), given associated uncertainties. The determination of a full posterior probability distribution of model predictions will generally require drawing samples of parameter and input data sets from the prior, computing resultant model predictions, and comparing those predictions against observations with the likelihood function (Rougier, 2007). The likelihood function, in some sense, acts as a metric or, more crudely, as a measuring stick. Posterior probability determination, therefore, adds another order of computational load because repeated model runs are required. A key innovation in this regard has been the introduction of statistical emulators of complex models that enable large sampling within available computational resources (e.g. Tarasov et al., 2012).

As noted, ice sheet modelling needs to be constrained by observations from the palaeo-record, which can be relatively scarce. For example, present-day climate system modelling benefits from terabytes of daily data retrieved from satellites. This is in sharp contrast to the limited set of records for the last glacial cycle, and their uneven distribution in both space and time. Further uncertainties are due to the largely indirect nature of most proxy records in relation to the quantities/characteristics of interest, as well as the inherent dating uncertainties associated with the proxy record. Given the resolution and approximations, ice sheet models will not, in the foreseeable future, freely recreate inferred margin chronologies with even 100 km (except for margins subject to strong topographic controls such as continental shelf breaks) and 500 year accuracy. Uncertainties in local climate forcing and controls on fast flow necessitate the use of some form of applied nudging of modelled margins positions towards geologically-based inferences (Tarasov et al., 2012). Dated ice margin reconstructions are, therefore, an important constraint, but have their own uncertainty. The explicit quantification of these uncertainties in the form of maximum and minimum isochrones for each time-slice is an important step forward (see Section 3.3). However, the long-term goal is the calibration of glaciological models against the direct marginal constraint data that have been used to construct the geologically-inferred isochrones, especially for regions where marginal constraints are sparse.

4.2. Constraining Ice Sheet Thickness and Palaeotopography Using Glacial-Isostatic-Adjustment Data

Except for cosmogenic nuclide dating of nunataks and trimlines (Section 3.1), there is little direct evidence to constrain the thickness distribution and related surface topography of palaeo-ice sheets. These constraints are largely provided from records of glacio-isostatic-adjustment (GIA), which can be inverted to reconstruct ice sheet configuration. Pure 'GIA-based' reconstructions invoke iterative tuning of ice load chronologies to fit relevant GIA data, subject to geologically-inferred ice margin chronologies (e.g. Peltier and Andrews, 1976). Traditionally, the GIA-based method was largely reliant on relative sea level (RSL) data, which have no coverage for the large regions of palaeo-ice sheets that are presently ice covered and/or were not exposed to post-glacial submarine conditions. The recent large expansion of coverage of measurements of present-day vertical velocities from continuous or repeat GPS (Argus et al., 2014; Peltier et al., in press) has significantly improved the body of constraints available from this measurement system for deglaciated regions lacking RSL data. The ICE-NG series of deglacial ice load reconstructions (Tushingham and Peltier, 1992 (ICE-3D); Peltier, 1994; 1996 (ICE-4G); Peltier, 2004 (ICE-5G)) have been the most widely used examples of the GIA-based approach. The most recent model in this series, ICE-6G_C (VM5a), uses the largest set of space geodetic constraints currently available to both constrain and validate the reconstruction (Argus et al., 2014; Kierulf et al., 2014; Peltier et al., 2015). Except for the glaciologically-derived Greenland component (Tarasov and Peltier, 2002) of ICE-5/6G, these models are not subject to any explicit glaciological constraints. To enforce semblance to realistic ice-sheet profiles, the GIA-based reconstruction of Lambeck and colleagues (e.g. Lambeck et al., 2010) invoke equilibrium parabolic glaciological flow-line approximations.

GIA-based reconstructions of ice sheet thickness variations require forward modelling to predict the geophysical observables that serve as constraints. This methodology requires two primary inputs. The first input is the glaciation history itself, which is inferred (if non-uniquely) by adjusting the glaciation history so that the RSL and vertical velocity predictions of the model fit the observations. For the non-glaciological components of the ICE-NG series, this essentially (although not trivially) involves moving around conceptual blocks of ice until the best fits are achieved. The second is the chosen representation of the internal viscoelastic structure of the planetary interior which, in turn, is used to generate the aforementioned RSL predictions for a given ice loading history. How the theoretical structure is able to separate error in the targeted glaciation history from error in the viscoelastic structure warrants discussion.

The elastic component of the viscoelastic structure is well-constrained by the almost spherically symmetric distribution (properties are a function only of depth in the Earth) of elastic Lame parameters of the Hookean elastic model of the planetary interior that is constrained by body wave and free oscillation seismology. By the elastic component of the rheology of Earth material, we mean that property which would determine the material response to an applied stress if liquid-like ‘flow’ were impossible. The only additional parameter required is to specify the viscosity component of the viscoelastic structure. Although it is well understood that this viscosity should be laterally heterogeneous in a medium in which the solid state convection process is occurring that is required to understand plate tectonic phenomenology, it is unclear as to the horizontal scale on which such lateral heterogeneity should be significant. Therefore, the dominant models in the ICE-NG series continue to be based upon the assumption that the viscosity may be adequately represented as spherically symmetric. In the ICE-NG (VMX) series of models, the depth dependent profile of viscosity VMX is constrained by demanding that the same profile be capable of reconciling the GIA observations from every region that has undergone the crustal rebound process, following the most recent deglaciation. In constructing this model, use is made of the fact that ice sheets of increasing lateral scale are sensitive to the viscosity of the Earth over an ever increasing range of depths. The data employed to infer the depth dependence of viscosity (wavelength dependent relaxation times) are chosen in such a way that they are only weakly dependent upon the deglaciation history employed to fit the observational constraint. Note, however, that this weak dependence breaks down in ice marginal regions (thus local variations in the earth rheology for these regions are much harder to constrain). The VM2 viscosity profile of the ICE-5G model, and the simple multilayer fit to this profile provided by the VM5a model in ICE-6G_C (VM5a), provides an excellent fit to the majority of the GIA-related observations. In the most recent work, however, a further refinement to this viscosity profile has been shown to be necessary to incorporate the additional constraints provided by relative sea level history data from the region of postglacial forebulge collapse outboard of the LIS along the eastern and western seaboards of the continental United States (Roy and Peltier, 2015). Thus, the process of model improvement is an iterative one in which one starts with an assumed known depth variation of mantle viscosity determined on the basis of observations that are relatively independent of the thickness of glacial ice that, when removed during deglaciation, was responsible for inducing the time dependent uplift of the land that is recorded in radiocarbon-dated RSL histories. One then adjusts the time dependence of ice sheet thickness within the inferred deglaciation isochrones (when available) to obtain the amplitude(s) of observed sea level fall at sites that were once ice covered. One then invokes additional data, such as GPS observations of present day rates of vertical motion of the crust, to further constrain the ice sheet loading history, and still further data to further refine the radial profile of mantle viscosity, until a fully converged model is obtained.

Although the dominant models of the GIA process continue to those based upon the spherically symmetric ansatz, it is important to note that significant current effort is being expended to investigate the extent to which lateral heterogeneity of the internal mantle viscosity structure may be influencing the conclusions concerning deglaciation history to which such GIA analyses have led. Recent examples of such work include van der Wal et al. (2013), which focused upon an attempt to infer 3D rheological Earth properties for Fennoscandia; and that of Austermann et al. (2013), which focused upon the rebound process in the Caribbean. In the former, the authors found that the expanded parametric range of laterally varying Earth rheology could improve RSL fits of the ICE-5G (VM2) ice loading chronology of Peltier (2004). The latter study found that lateral heterogeneity of viscosity in the vicinity of Barbados could significantly perturb the fit of the same ICE-5G (VM2) model to the Peltier and Fairbanks (2006) record that is employed to constrain the net eustatic increase in sea level across the most recent glacial interglacial transition. In neither of these studies was attention focused upon an analysis of the uncertainties concerning the conclusions to which the authors were led. A research priority for the community is a clear specification of regional uncertainties in Earth rheology along with the quantification of the impact of these uncertainties on inferred deglacial ice sheet chronologies.

A subsequent evolution has been the imposition of the available set of GIA-based constraints on 3D glaciological models. This has progressed from hand-tuned models (e.g. Tarasov and Peltier, 2002) to approaches that explore and quantify (to varying extents) uncertainties due to climate forcing and glaciological model components (Tarasov et al., 2012; Briggs et al., 2014) as well as uncertainties in the regional Earth rheology (Whitehouse et al., 2012; Lecavalier et al, 2014). The much expanded constraint of glaciological-self-consistency, however, can come with an expected cost. For regions of dense GIA-data coverage, such as North America, there is (to date) a trade-off of poorer fits to some of the GIA data (e.g. Tarasov et al, 2012). However, for regions under current ice-cover, such as Antarctica, GIA-based observatyions are sparse and provide relatively little constraint for deglacial evolution (Briggs et al., 2014). Glaciological modelling also offers a clearer path to implementation of Bayesian model calibration to generate a probabilistic distribution of deglacial ice sheet chronologies (Tarasov and Peltier, 2012) based on model fits to constraint data.

As a partial comparison of the two methodologies (pure-GIA versus a 'glaciological' ice sheet modelling approach), Figure 13 shows a set of 6 selected, but not necessarily representative, examples from hundreds of comparisons available (see Supplementary Figures 1 and 2). These are comparisons between relative sea level (RSL) observations and model predictions based upon application of the ICE-6G_C (VM5a) model and two of the overall best-fitting model runs detailed in Tarasov et al. (2012). The two model runs have similar overall RSL scores, but run nn9927 has a much better fit to marine limits, whereas nn9894 has a much better fit to strandline data. Both methodologies employ the same assumption as to the depth dependent viscoelastic structure of the Earth, namely the VM5a structure of Peltier and Drummond (2010). They both also employ the same dataset to describe the space time evolution of the Laurentide, Cordilleran and Innuitian ice sheet margin positions (the calibrated glaciological model has, however, a clearly defined uncertainty assessment for the ice margin, given in Tarasov et al, 2012), and the same database of relative sea level history constraints. The glaciological model is additionally constrained by strandline elevations (and associated age constraints). Parts of the Canadian

Arctic Archipelago (especially most of Ellesmere Island and Prince of Wales Island) are poorly fit by both nn9894 and nn9927 (Tarasov et al., 2012), whereas most RSL data envelopes are very well fit by the GIA-based model in this region. This high latitude misfit of the glaciological-based model runs to the RSL data is not shared by earlier models that have been produced using this methodology. It should also be noted that the plotted one-way error bars (the upward or downward “ticks”) for the limiting RSL data have been severely truncated for the sake of visual clarity. There are also sites where the glaciological models have better RSL fits than the GIA model (e.g. Melville Island sites 1253-4, NE Banks Island 1356, and Makkovik 1608). For the majority of the North American ice sheets the RSL fits are not dissimilar for the three ice chronologies.

Model differences around Hudson Bay (e.g. sites 1631-1641) illustrate a further characteristic difference between the predictions of the GIA-based model and those of the two glaciological models (Figure 13). Although both methodologies deliver reasonable fits to the RSL observations over the range of time over which such observations are available, the GIA-based model predicts a significantly larger net fall of sea level (land uplift), a consequence of the fact that the thickness of the ice sheet at the LGM is significantly greater in the GIA-based model than in either of those that have been delivered by the Bayesian calibration procedure in the glaciological approach.

In summary, both the pure GIA-based and the glaciological ice sheet model-based approaches have their strengths and weaknesses when reconstructing ice sheet palaeo-topography. A challenge for the pure GIA approach is that the space time thickness histories that it delivers will, in general, not be in accord with glaciological first principles. Furthermore, hand-tuning of pure GIA-based ice histories precludes any rigorous uncertainty assessment, although methodology is being developed that will enable this shortcoming to be eliminated. A challenge for glaciological modelling is that not only is a sophisticated ice sheet model required, but so too is a palaeo-climate model to drive the evolution of ice cover. The latter necessarily adds more degrees of freedom, but provides some physical constraint in terms of climatic self-consistency. By ignoring climate forcing and glaciological process issues, pure GIA methods offer a benchmark for quantifying the impact of climate and glaciological constraints.

4.3. Modelling Palaeo-Ice Shelves and Calving

As noted (Section 2.3), geophysical investigation of the ocean-floor has revealed that many palaeo-ice sheets and ice shelves were far more extensive than previously recognised. In the Arctic, for example, ice shelves likely covered extensive areas of the Amerasian Basin of the Arctic Ocean, the Chukchi Borderland, the Siberian Shelf, and the northern Svalbard margin (Section 2.3, Figure 7). When data supporting the existence of these ice shelf complexes began to emerge, the most commonly-used numerical models were unable to adequately capture the coupled ice sheet-shelf systems because they used the Shallow-Ice Approximation (SIA) (Blatter et al., 2011, Kirchner et al., 2011). Indeed, model results obtained with an SIA model for grounded ice will differ from their counterparts obtained from a coupled ice sheet-ice shelf model, because a SIA model is unable to account for the impact which an ice shelf may have on inland ice (e.g. through buttressing: see Section 4.4).

As the need for considering ice shelf dynamics was realized, models using a 'Shallow Shelf Approximation (SSA)' were developed (e.g. MacAyeal, 1989; Weis et al., 1999). However, simple coupling of SIA and SSA models across the grounding line was difficult and so alternative approaches explored how ice dynamics in the transition zone could be modelled without having to employ the more computationally intense 'Stokes' equations (Chugunov and Wilchinski 1996; Hulbe and MacAyeal 1999; Schoof, 2007; Schoof and Hewitt, 2013).

Earlier recommendations to employ second order SIA and SSA models had to be rectified (Kirchner et al., 2011) and have led to the development of new adaptive, error-based couplings schemes (Ahlkrona et al., submitted), while so-called ‘hybrid’ models (Pollard and DeConto, 2012, Bueler and Brown, 2009) make use of an heuristic approach of coupling an ice sheet and ice shelf. Hybrid models can also be used for simulations of ice sheet-shelf systems, including grounding line migration, at comparatively low computational cost (e.g. Pollard and DeConto, 2009; 2012).

In addition to the development of coupled ice sheet-shelf models, the modelling of calving has gained prominence. Calving is usually included in a coarse parametric fashion, introducing considerable uncertainty to model results. Recent years, however, have seen the implementation of physically-based calving laws to ice flow models (Benn et al., 2007a, b; Nick et al., 2010). Using an extension of the crevasse depth formulae of Nye (1955; 1957), the criterion defines the calving front as where surface and basal crevasses penetrate the full ice thickness. By calculating crevasse penetration depth from the first derivative of glacier velocities (longitudinal strain rates), calving is linked directly to ice dynamics. Incorporating the effects of ponded meltwater (allowing deeper crevasse penetration) also allows calving rates to be driven by climate fluctuations via changes in surface runoff (e.g. Nick et al., 2013). These physically-based calving mechanisms have begun to be implemented in numerical modelling (e.g., of Antarctica) (Albrecht et al., 2011; Levermann et al., 2012; Pollard and DeConto, 2012; Briggs et al., 2013; Albrecht and Levermann, 2014; Pollard et al., 2015), but remain particularly challenging in palaeo-applications where the intermittent and often rapid nature of the calving process has to be properly accounted for on long timescales.

In summary, numerical modeling of ice shelves and calving is mostly focussed on predicting extant ice sheet dynamics. Thus, the modeling of palaeo-ice shelves and calving is very much in its infancy. Apart from the long timescales involved (causing mostly numerical challenges) two major issues will need to be overcome. First, the accurate description and implementation of ice sheet-ocean interactions will require a rigorous treatment not only of calving, but also melting and refreezing processes beneath an ice shelf, for which very simple parameterizations are currently used. Given the dependence on sub-shelf ocean circulation and water temperature, accurate representation of this process for palaeo-timescales will likely pose a long-term challenge. Therefore, the second issue is the need for a rigorous assessment of model sensitivity to calving rates and sub-ice shelf melt rates so that uncertainty related to these processes can be quantified. Given the scarcity of data, this aspect of model calibration will be challenging and will also require an accurate treatment of the grounding line (see Section 4.4).

4.4. Improvements in Modelling Marine Ice Sheet Grounding Lines

Recent years have seen a focus on improving the modelling accuracy of marine ice-sheets, where the bed is substantially grounded below sea level (Hindmarsh, 2006; Schoof, 2007; Katz and Worster, 2010; Nick et al., 2010; Docquier et al., 2011; Drouet et al., 2013; Feldmann et al., 2014). The grounding line is where ice, flowing from its source areas, begins to float. Grounding lines circumscribe much of the grounded ice in Greenland and nearly all of Antarctica, and grounding lines were present in substantial parts of the margins of palaeo-ice sheets (Figure 7). The most challenging aspect in these models is the representation of the grounding line and its migration.

At the simplest level, the mechanics of grounded ice and ice-shelves are rather different. In both cases, their motion is conceptualised as being due to a driving stress; in simple ice-shelves, this is proportional to the thickness of the ice, while in grounded ice it is proportional to the product of ice thickness and surface slope. This led to the idea of the 'marine ice-sheet instability' or 'grounding-line instability', first put forward by Weertman (1974). At the grounding line, close to the ice-shelf, the appropriate driving stress is the 'ice-shelf' driving stress. Following mathematical analysis, the rate at which ice spreads at the grounding line is proportional to the ice thickness raised to a power of around three (van der Veen, 2013). This is a consequence of the rheological properties of ice and the assumption that the appropriate mechanical description is the same as for floating ice-shelves. Evidently, if the grounding line retreats on an adverse slope (deepens inland), the thinning rate will increase, because the thickness of the ice at a retreating grounding line must increase. This gives rise to the possibility of a grounding-line instability. While the idea of marine ice-sheet instability was largely accepted by the glaciological community, there were doubts about the rigour of Weertman's theory because it was not a full solution of the equations describing mechanics and mass conservation.

Recently, Schoof (2007) confirmed Weertman's original supposition and provided the first theory of grounding line dynamics in a formula that related flux of ice across the grounding-line to the ice-thickness. This permitted ice-sheet modellers to compare the flows calculated by their models with the Schoof flux formula, and thus to understand the requirements for calculating flux across the grounding-line to sufficient accuracy. The particular computational issue is the accurate calculation of flow in a 'boundary layer', where the ice flow changes from sheet flow, resisted by basal shear stress, to extending flow characteristic of ice shelves. The accelerating flow in this boundary layer has to be captured properly for accurate calculation of grounding-line advance or retreat (Schoof, 2007). The extent of the boundary layer for ice sheet flow is typically 10-20 km (Hindmarsh, 2006), while for a perfectly slippery stream where all the resistance comes from the lateral margins, the boundary layer approximates the ice-stream width (Hindmarsh, 2012).

Alternative mathematical approaches (Katz and Worster, 2010), numerical experiments, and model inter-comparisons (Docquier et al., 2011; Pattyn et al., 2012; Drouet et al., 2013) have confirmed the basic accuracy of the Schoof (2007) formula and approach. However, models incorporating the full set of appropriate mechanical equations predict flow across the grounding line up to twice as fast as given by the Schoof formula (e.g. Drouet et al., 2013). These aspects of the mechanical problem are discussed by Nowicki and Wingham (2008) and Fowler (2011), drawing on earlier work by Chugunov and Wilchinsky (1996). A recent review of these issues and ice-sheet dynamics, more generally, is provided by Schoof and Hewitt (2013).

In addition to expressing the relationship between ice flux and thickness, the Schoof formula shows how flux depends on the viscous properties of ice and the basal resistance. Most importantly, the Schoof (2007) formula includes a parameter that represents buttressing from an ice shelf. Buttressing is crucial because it opposes the tendency of the ice to spread and thin at the grounding-line, and can inhibit retreat and augment advance (Goldberg et al., 2009). Again, numerical experiments indicate that the Schoof (2007) formula is accurate in the presence of specified buttressing (Drouet et al., 2013), but it is also true that representing buttressing by a parameter diverts attention from the interplay between flow across the grounding line and the buttressing from an ice shelf. For example, increased flow associated with thinning might induce increased resistance from ice shelves. This has been demonstrated in numerical experiments to the extent that grounding lines on reverse slopes can be stabilized; in other words achieve steady positions resistant to perturbation (Gudmundsson et al., 2012). Related to this, is the notion that radial grounding lines associated with radially spreading shelves can buttress and stabilise grounding lines (Pegler and Worster, 2012). Changes in buttressing are also likely to be the primary link through which ocean melting of ice-shelves, which is strongly associated with ice-sheet thinning in modern glaciers (Pritchard et al., 2012), affects the grounding line (Goldberg et al., 2012). Changes in ice-stream flux from surging can also cause advance or retreat of grounding-lines (Robel et al., 2014).

The accuracy of the Schoof (2007) formula led Pollard and DeConto (2009) to incorporate it directly into their palaeo-marine-ice-sheet modeling of the time-dependent behaviour of the Antarctic ice-sheet (AIS) over the last 5 Ma, with the grounding line contracting and expanding over 41 ka obliquity, and 100 ka eccentricity cycles. The alternative to their approach is to solve the equation in the boundary layer numerically, which must account for the much greater accuracy required there. This requires more advanced techniques, such as adaptive nested grids (Gladstone et al., 2010; Cornford et al., 2013; Feldmann et al., 2014) or higher-order methods. These methods have yet to be published in full, but results of an initial inter-comparison can be found in Pattyn et al. (2012). They are significantly more computationally intensive than use of the Schoof formula, and thus are unlikely to be used in glacial cycle modelling for the foreseeable future.

The Schoof (2007) formula also highlights the parameters whose specification requires improvement; principal among these are buttressing, strongly affected by the ocean melt-rate near the grounding-line, and bed resistance. The latter, in areas currently occupied by ice-streams, is at best taken to be the same as present-day values inferred from glaciological inversions (Joughin et al., 2006), or else assumed or taken as a tuning parameter, constant in time. Sub-shelf melt is not yet computed with oceanographic modelling in palaeo-contexts (see Section 4.3), and this situation is likely to persist owing to the ocean models' high computational expense and the inability of limited observational records to even validate modelling of present conditions. For example, the Pollard and DeConto (2009) model does incorporate and require varying back-pressure from ice shelves. This varying back-pressure is indirectly forced through varying the ocean melt at the base of ice-sheets, using a highly parameterized representation of the ocean.

An important aspect of grounding-line modelling of large sectors of ice-sheets is using knowledge of the thickness of the ice sheet in the past to constrain grounding line positions, and thereby ice volume. Bentley et al. (2010) and Le Brocq et al. (2011) have explored this with respect to the Weddell Sea sector of the West Antarctic Ice Sheet (WAIS). They calculated ice sheet elevations consistent with quasi-steady grounding line positions, concluding that the Weddell Sea grounding-line did not advance to the continental shelf edge at the LGM. However, Hillenbrand et al. (2013) concluded that, on the basis of marine geological evidence, the grounding line extended to the continental shelf margin, possibly only briefly. More systematic views of matching data with models are provided by Whitehouse et al. (2012) and by Briggs and Tarasov (2013). Whitehouse et al. (2012) and Argus et al. (2014) considered the data/modelling matching problem for all of Antarctica, with an emphasis on GIA interactions; whereas Briggs and Tarasov (2013) made a systematic analysis of how to consistently incorporate data sources in inverse modelling procedures.

The impact of GIA on ice dynamics has been a long-lived theme in ice sheet modelling (Oerlermans, 1980; Tarasov and Peltier, 1997). Gomez et al. (2010, 2012) have recently expanded this with the addition of self-gravitational effects (i.e. the effect of changing mass distribution on the geoid) and analysis of the resulting combined impact on grounding line stability. Retreat from glacial maxima is likely to have been into much deeper basins, which will have subsequently shallowed through GIA, adding considerably to the potential complexity of the evolution. Moreover, loss of ice-volume leads to instantaneous lowering of the raised geoid around ice-sheets as explained in terms of the gravitationally self-consistent Sea Level Equation that is fundamental to the understanding of the GIA process (Clark et al, 1978; Peltier et al, 1978). Thus, in certain cases, the grounding-line does not retreat into deeper water, even though it lies on a reverse slope. Bradley et al. (2015) present geophysical evidence for GIA-related advances of the grounding line in certain sectors of the Antarctic ice-sheet.

To summarise, until recently, modelling inadequacies were a bigger constraint on understanding the evolution of palaeo-marine ice-sheets than data inadequacy. The opposite is probably true now, although there remains plenty of scope for the practice of marine ice-sheet modelling to further improve. The ruling hypothesis about current marine ice-sheet retreat, particularly in Antarctica, is that it is due to warming oceans. If this holds for palaeo-ice-sheets, the representation of this in models is the next computational challenge; and data to constrain the ocean temperature on continental shelves are likely to become increasingly important. Modelling can also answer a related question: what ocean conditions are needed to drive the observed retreat of marine ice-sheets?

4.5. Modelling the Impact of Iceberg and Meltwater Events on the Ocean-Climate System

As discussed in Section 2.7, there is abundant evidence of meltwater and IRD pulses from palaeo-ice sheets during the last deglaciation. The impact of these pulses occurs via massive meltwater input at the coast and its subsequent mixing within the oceanic system or through the release of icebergs that acts as a mobile heat sink and freshwater source. Although all these effects are important for the ocean-atmosphere system, no modelling studies have so far included all these effects with the adequate physical and spatial resolution.

In numerous modelling studies, meltwater fluxes have been applied to perturb the Atlantic Meridional Overturning Circulation (AMOC) (Stommel, 1961; Bryan, 1986; Stouffer et al., 2006) and the associated northward heat transport, thus creating abrupt climate events (Paillard and Labeyrie, 1994; Ganopolski and Rahmstorf, 2001). Often, the North Atlantic has been used as the region where the freshwater flux was applied since it is: (i) advectively close to the main north Atlantic convection sites (where deep sinking of denser surface water occurs), thus ensuring an effective response of the simulated oceanic system; and (ii), where the main marine deposits of IRD are found (see Section 2.7) i.e. in the so-called 'Ruddiman belt' between 45 and 55°N (Ruddiman, 1977; Heinrich, 1988; Hemming, 2004). Consequently, most models have been run with an input of freshwater, termed 'hosing experiments', Ganopolski and Rahmstorf, 2001; Stouffer et al. 2006) in the Ruddiman belt, with different sets of magnitudes used to evaluate the sensitivity of the models to a freshwater perturbation (see Kageyama et al., 2010, for a review).

Model inter-comparisons show that not all models are sensitive to freshwater hosing in the same manner (Stouffer et al., 2006; Kageyama et al., 2010). For example, while all models show a shutdown of the thermohaline circulation under a 1.0 Sv hosing (Stouffer et al., 2006), they differ in response when the forcing is closer to what is currently estimated for Heinrich events, i.e. 0.1 to 0.4 Sv (Hemming, 2004). Moreover, when a different climate state is used (e.g. Ganopolski and Rahmstorf, 2001; van Meerbeeck et al., 2009; Kageyama et al., 2013), the response of the climate models to freshwater forcing is also different (it becomes more or less sensitive depending on the details of the boundary conditions and the model used). Recently, it has become clear that state-of-the-art GCMs tend to overestimate the oceanic stability with respect to freshwater forcing (Valdes, 2011). However, because the actual sensitivity of the climate system is unknown, and could be different between different climate states, an assessment of the most likely behaviour of the system is not possible.

The effects of melting icebergs have also been (explicitly or implicitly) simulated as freshwater input in climate models (Ganopolski and Rahmstorf, 2001; Otto-Bliesner and Brady, 2010). However, icebergs not only freshen the surface ocean through melting, but also through cooling the ocean by the heat uptake needed to melt the ice, and by having a distribution that is modified by climate itself through the winds and surface ocean currents. Hence, modelling them only as simple freshwater fluxes in traditional 'hosing' experiments potentially neglects important processes. As such, recent thermo-dynamical iceberg models have been coupled to climate models. One of the first studies to undertake this approach was Levine and Bigg (2008), who accounted for both the distribution of icebergs and their freshwater release. With icebergs, they found a greater freshwater release is required to have the same effect on the AMOC than the traditional freshwater hosing approach. They attributed this to the more localized freshwater input from icebergs. They also found that the length of the pathway that the icebergs travel to the main deep oceanic convection sites is an important influence on their eventual impact. A similar conclusion was reached by Jongma et al. (2009), who showed that the main freshwater input from icebergs originating from Hudson Strait (see Section 2.2) was in the western North Atlantic, thus diminishing the impact on the Nordic Seas' deep convection when a comparable amount of freshwater is used to those in the traditional hosing experiments.

Using a more complete modelling approach, including the latent heat exchanges, Jongma et al. (2013) showed that the effect of an iceberg armada from Hudson Strait on the AMOC strength was relatively similar to the traditional freshwater hosing approach, albeit with a different time evolution and through rather different physical mechanisms. Indeed, they showed that, contrary to Levine and Bigg (2008) and Jongma et al. (2009), icebergs were much more efficient at suppressing deep convection when the latent heat is taken into account. This counter-intuitive result (cooling the surface ocean should generate denser waters) is explained by a 'sea-ice facilitation effect', whereby the cooling effect of icebergs promotes sea-ice formation. Although sea-ice formation increases the salinity and density of the surface waters through brine rejection, a denser sea-ice cover also strongly reduces the loss of oceanic heat to the atmosphere, leading to a lower surface water density. The two effects are thus competing, but the insulating impact is stronger, and the net result found is a large reduction in dense water formation. In comparison to the traditional approach, Jongma et al. (2013) found that freshwater 'hosing' is not a good representation of the Heinrich events, because simulating the iceberg distribution induces important east-west differences in the North Atlantic (fresher and colder western north Atlantic and saltier eastern north Atlantic) that leads to substantial differences in the AMOC recovery after the Heinrich events. If one intends to compare modelling results with North Atlantic proxy data, the freshwater hosing approach is probably not a valid approximation. The geographical envelope of iceberg paths modelled in Jongma et al. (2013) is consistent with the spread of IRD observed from ocean cores drilling in the North Atlantic (Hemming, 2004; see Section 2.6) as seen on Figure 14. A similar geographical spread of the melting was found recently by another study that uses the IRD distribution of the North Atlantic Ocean as a constraint for the model (Roberts et al., 2014).

Besides icebergs as a source of meltwater, it is also important to account for the changing drainage of meltwater from ice sheets in both space and time. Drainage events have been associated with cooling events during deglaciation (Section 2.5) and it is important to analyse the modelled climatic response, not only to different volumes and rates of meltwater release, but also to the location of the freshwater input in the oceans. In the context of analysing the likely causes of the YD cold period, Peltier et al. (2006) analysed the effect of routing a large freshwater input into the Arctic Ocean as proposed by Tarasov and Peltier (2005). Peltier et al. (2006) found that the effect of an Arctic Ocean input for the freshwater flux is very similar to the same input in the North Atlantic, albeit with a small delay of a decade, and some slight differences in the time evolution of the strength of the AMOC. Consequently, in their low-resolution coupled climate model, a surface freshwater flux affects the AMOC even if the forcing is not directly located in the North Atlantic region, close to the deep convection sites. More recently, the first ocean modelling study to examine meltwater transport with a high enough resolution to permit ocean eddies (Condron and Winsor, 2012) found that, for an intense flood over a 3-year period, a much larger fraction of freshwater discharge into the Arctic was likely to reach deep water formation sites than the same discharge into the Gulf of St Lawrence. It is, however, unclear what the effect would be after a few decades and if this result is linked to the sensitivity of one particular model.

Since many other river outlets or ice-sheet margins may provide freshwater input during the last deglaciation (e.g. Marshall and Clarke, 1999; Tarasov and Peltier, 2006; Not and Hillaire-Marcel, 2012), Roche et al. (2010) set up a low-resolution modelling study to examine the response of one particular climate model to the input of freshwater of different magnitude and systematically tested all locations. They showed that the effect of freshwater input on the AMOC can be predicted on the basis of an 'advective distance' to the main sites of deep convection. The closer the input to the main deep convection sites, the easier it is to disrupt the AMOC; the further away (in an advective sense), the more likely the freshwater will undergo some mixing along the way and its effect will be reduced. They also demonstrated that, in sea-ice covered regions, the freezing of a large part of the freshwater that is then transported as sea-ice (in accordance with the Arctic Ocean scenario of Peltier et al., 2006) prevents the mixing of the waters and enhances the impact on the AMOC compared to a scenario without this sea-ice effect. Using a more complex (but still relatively low resolution) model, Otto-Bliesner and Brady (2010) also investigated the effect of variable magnitude freshwater fluxes in two different locations: the North Atlantic and the Gulf of Mexico. They confirmed the result obtained by Roche et al. (2010), showing that the AMOC reduction is less in the case of a freshwater input into the Gulf of Mexico than in the North Atlantic, due to mixing and re-circulation of part of the freshwater anomaly through the North Atlantic sub-polar gyre. A similar conclusion was reached for the 8.2 ka event by Li et al (2009), who evaluated the impact of different meltwater drainage routes on the AMOC response.

In summary, there is a large range of model responses to schematic freshwater forcing experiments in low-resolution ocean-climate models. When using models with a higher resolution and eddy resolving oceanic components, the sensitivity to freshwater forcing seems to be very sensitive to the location of freshwater injection. Some modeling experiments are now starting to include more realistic iceberg treatment, including the dynamics of freshwater release, latent heat and iceberg drift. Results obtained are showing an iceberg distribution in broad accordance with data constraints. No data-based constraints are available to characterize the actual sensitivity of the oceanic circulation to meltwater in different climate states.

5. Discussion and Conclusions

Given the recent advances and key challenges identified above, we outline some key methodological challenges for palaeo-ice sheet reconstructions, with an emphasis on how future work might integrate terrestrial and marine evidence with numerical modelling.

5.1. Improved Understanding of the Genesis of Subglacial Landforms

The burgeoning availability of remote sensing imagery and marine geophysical datasets has led to major advances in our ability to map the glacial geomorphology of palaeo-ice sheet beds. However, one major deficiency (see Section 2.1) is that our process understanding of how various landforms are created is incomplete. This is what Kleman et al. (2006) referred to as the ‘genetic problem’, i.e. deciphering the processes by which, and the conditions under which, particular landforms are created. Many of the landforms that are important in glacial inversion methods (e.g. ribbed moraine, drumlins, mega-scale glacial lineation) are least understood in terms of their genesis. Recent observations of landforms being created under modern ice masses hold much potential for further advances in glacial inversion techniques (King et al., 2009; Smith et al., 2007) because specific bedforms can be linked to specific glaciological conditions (e.g. ice thickness, velocity, subglacial hydrology). These observations provide the 'missing link' between glaciology and palaeo-glaciology, and will considerably refine glacial inversion methods and help to parameterize subglacial processes in ice sheet models (Bingham et al., 2010). If, for example, it could be demonstrated that fields of drumlins of a given amplitude and spacing were formed under specific ice dynamical conditions (velocity, thickness, basal temperature), they would provide additional constraints for palaeo-ice sheet reconstructions. Indeed, statistical descriptions of flow-set characteristics (see Hillier et al., 2013) could be compared to ice dynamics generated by numerical models. At the same time, one can expect that future technological advances will permit whole landscapes to be imaged beneath the Greenland and Antarctic ice sheets. Glacial inversion techniques that have been applied to palaeo-ice sheets will provide the only observational template to interpret what is likely to be an incredibly complex assemblage of cross-cutting landforms that reveal important aspects of the ice sheet's evolutionary history.

5.2. Capturing Ice Stream Dynamics in Numerical Ice Sheet Models

Once identified (see Section 2.2), the 'known' location of palaeo-ice streams provides a useful observational record to test ice-sheet models that generate ice streams (Boulton and Hagdorn, 2006; Hubbard et al., 2009; Stokes and Tarasov, 2010; Golledge et al., 2013). Given the current attention to improving ice-sheet models and capturing ice-stream dynamics, there is an urgent need to understand the properties (i.e., boundary conditions) that initiate, sustain or inhibit ice streaming (Winsborrow et al., 2010). This can be achieved through a combined approach that evaluates model predictions of ice streaming against independent geological evidence of their activity. Few studies have attempted this, but Stokes and Tarasov (2010) found generally good agreement for the LIS, despite the use of the Shallow-Ice-Approximation (see Section 4.3) and a relatively coarse grid-resolution. More recently, Golledge et al. (2013) presented a sector-by-sector comparison of modelled versus empirically-derived ice-stream locations for the Last Glacial Maximum in Antarctica, using a model with a much higher resolution (5 km) and that incorporated both the shallow ice and shallow-shelf approximations. At the continent scale, the flow characteristics of the modelled ice sheet is in good agreement with the present-day ice sheet, and they were able to capture the complex dendritic velocity pattern of ice streams fed by tributaries of intermediate velocity (Bamber et al., 2000) that extend well into the ice sheet interior (see Figure 15). There is, however, a clear need for further data-model calibrations with higher order models and with better treatment of subglacial hydrology, which may improve the ability to model terrestrially-terminating ice streams in the Northern Hemisphere ice sheets (Stokes and Tarasov, 2010).

A major limitation of using known palaeo-ice stream locations to test numerical models is, however, the current lack of dating control on ice stream flow-sets (Stokes and Tarasov, 2010; Margold et al., 2015). There is, therefore, a requirement for additional chronological constraints on palaeo-ice stream activity, including from marine IRD records (e.g. Andrews et al., 2012) which has much potential to test numerical models and refine ice sheet reconstructions (Mackintosh et al., 2011; Jamieson et al., 2012).

5.3. Capturing Subglacial Hydrology in Numerical Models

The imprint of meltwater drainage recorded on the bed of former ice sheets offers a potentially powerful constraint to develop, calibrate and test numerical models. Observations from modern ice sheets have highlighted the importance of water pressure and deformable wet sediments in lubricating the bed and controlling ice stream behaviour (Bell, 2008). However, subglacial hydrology models are still based on a priori assumptions about the nature of the drainage network and we lack the observational data at the spatial and temporal scales necessary to constrain and test them. A key challenge is to exploit the palaeo-meltwater record in a way that is useful to modellers at the spatial and temporal scales at which they operate. To do so, we require a better understanding of the origin of relict meltwater features, including the source of water, how quickly they were eroded/deposited, and where under the ice sheet they formed during the life-cycle of the ice sheet (Storrar et al., 2014a; Jansen et al., 2014). It also requires a more thorough characterisation of the whole drainage network, and so the identification of palaeo-subglacial lakes is also required. Likewise, in order to compare the relict subglacial imprint to drainage system morphology, subglacial hydrological models need to be extended to include the processes of sediment erosion, transport and deposition.

5.4. Capturing Proglacial Hydrology in Numerical Models

For a given probability distribution of deglacial ice-sheet evolution (including ice thickness, basal elevation and surface meltwater production), the accurate determination of regional meltwater discharge into ocean basins is reasonably straightforward on super-annual timescales. Water need only be routed down-slope until it either fills local depressions (while keeping track of water that is thereby locally stored) or reaches the ocean. Care is needed, however, in handling the sub-grid sensitivity of meltwater routing at choke points (Tarasov and Peltier, 2006) as surface drainage can otherwise get distorted with the grid-scale resolution of current numerical models. Routing sensitivity to the elevation of choke points also implies a need for accurate representation of GIA in the model (e.g. Tarasov and Peltier, 2002).

Even though meltwater discharge for a given ice-sheet history can be accurately computed in terms of surface runoff, the impact of these meltwater discharges are subject to large modelling uncertainties due to high sensitivities to uncertainties in reconstructed ice sheets, especially to the exact location of ice-sheet margins at choke points near spillways (Section 2.6). Models are unable to accurately predict past meltwater routing without palaeo-constraints on ice margin chronologies for these critical regions (Tarasov and Peltier, 2006). This sensitivity also implies that accurate estimation of margin-chronology uncertainty is critical (Section 5.8). The timing of regional discharge can change significantly with different uncertainty estimates for ice margin chronologies (Tarasov and Peltier, 2006). Given these ice margin uncertainties, there is therefore a need to constrain ice sheet records with palaeo-records that directly constrain meltwater routing or fluxes. Surface drainage also leaves geological and geomorphological indicators that are useful for constraining ice sheet models (Section 2.5). Fluvial erosion offers a measure of water fluxes (Jansen et al., 2014), but with complications arising from the non-linear response to water velocities. Strand-lines are an important direct measure of pro-glacial lake levels that can be directly compared against model output if their ages are well constrained (Tarasov and Peltier, 2006).

5.5. Freshwater Impacts on the Ocean-Climate System

State-of-the-art climate models are now frequently subject to ‘hosing’ experiments to test model response to freshwater fluxes (Section 2.5). However, most experiments are conducted at model resolutions that do not resolve the narrow high-velocity boundary currents of ocean basins and require major diffusive approximations that ignore the complex spatio-temporal structure of turbulent mixing and associated eddies. Furthermore, standard hosing experiments spread meltwater injection across the North Atlantic (e.g. Kageyama et al., 2013), as opposed to injection at the relevant coastal outlets. Higher resolution models (e.g. Condron and Winsor, 2012) are able to much better resolve these processes without diffusive approximations, but the higher horizontal grid resolution (18 km), along with 50 vertical levels, is computationally intensive and restricts the analysis to short time-scales. There is, therefore, a need to create alternative coarse resolution representations of eddy processes that are faithful to the underlying physics while permitting long model-time integrations. One potential route is the introduction of appropriate noise into the model equations to better represent the chaotic nature of turbulent eddies. This approach of stochastic sub-grid parameterization has been shown to significantly improve the representation of certain atmospheric processes (Berner et al., 2012). Ocean response is also highly sensitive to the location of meltwater injection (Condron and Winsor, 2012) and the spatial pattern of freshwater injection needs to be taken into account (Jongma et al., 2013). A related complication is the apparent sensitivity of the modelled climate system response to meltwater injection to the background state of the ocean-climate system (Kageyama et al., 2010; 2013). Finally, a further important question relates to the differing response to variations in the magnitude and duration of meltwater injections. A clear answer will again require high-resolution models that avoid implicit and explicit diffusive transports. Modelled freshwater injection rates are of moderate confidence on 100 year timescales, but except for lake drainage events, we have little confidence on shorter timescales given the time-resolution of available constraint data.

5.6. Fast and Accurate Palaeoclimate Models

Glaciological models require time-evolving palaeo-climate fields to compute the temperature of the ice and the (terrestrial and marine) surface mass-balance. The impacts of changing ice-sheet topography and meltwater fluxes on the ocean-climate system are also of high interest. Earth system Models of Intermediate Complexity (EMIC) provide representation of relevant Earth system processes at less than state-of-the-art levels of complexity to enable palaeo-ice sheet model integrations. Though a wide range of such models exists with varying emphasis on the complexity of individual components (Petoukhov et al., 2005), all have serious deficiencies. These range from missing processes, such as the dynamic evolution of the radiative effect of clouds and transport of sea-ice, to significant misfits in predicted monthly mean temperature and precipitation for the present-day climate. There is, therefore, an urgent need for the development of next-generation community EMICs that could complete glacial-cycle runs within a year of real time and that can handle the evolving area and bathymetry of the ocean (especially the opening and closing of gateways) and accurately respond to evolving ice sheets. Such a model would need fully coupled atmospheric, ocean, land surface, and sea-ice components. Given the importance of sea-ice and ocean feedbacks, a dynamic sea-ice model is required (i.e. that includes the stresses from wind and ocean currents) as well as an ocean GCM. The EMIC also needs to be well-documented, openly available, and designed for ease of coupling with new components and different users..

5.7. Quantifying Uncertainties in Ice Margin Chronologies

Dated (or age-bracketed) ice-margin positions are one of the most valuable ingredients for empirically-based ice sheet reconstructions and provide a powerful constraint for numerical modelling. Indeed, the compilation of pan-ice sheet margin chronologies has been pivotal in enabling data-calibrated numerical modeling (e.g. Tarasov et al., 2012). Ideally, empirical ice sheet reconstructions (e.g. of margin position or thickness) should clearly show the level of uncertainty or at least provide an estimate. The resolution of the margin positions is determined by both the amount of data available, and the chosen level (scale) of generalisation, which may leave large gaps (uncertainties) in some areas. Filling these gaps will be important if empirically derived ice-margin reconstructions are to keep pace with numerically modelled margin positions at 5 km resolution (Golledge et al., 2012; Pattyn et al., 2012; Seddik et al., 2012). Quantifying uncertainty is particularly important for numerical models using Bayesian calibration to produce probability distributions of the results (Tarasov et al., 2012).

It is also important that the data used to generate ice-margin chronologies are available for scrutiny. It is critical to distinguish between data (e.g. mapped moraine positions) and interpretations (e.g. interpolated ice margin positions where moraines are absent) and the interpretative rules used in the reconstruction. Our understanding of landform formation processes and thus how we interpret them may change (Section 5.1), but the distribution of landforms will not. In this regard, GIS is a particularly useful tool for providing data in a digital format, which allows for detailed scrutiny of the source data and facilitates comparison with other data or numerical model outputs (e.g. Napieralski et al., 2007; Li et al., 2007; Briggs and Tarasov, 2013).

5.8. Integration of Empirical Data and Numerical Modelling

An important long-term goal should be the rigorous specification of the probability distribution for past ice-sheet evolution. More practically, this should provide a meaningful envelope for past evolution that confidently captures 'reality'. In working towards that goal, an accurate representation of input and constraint data uncertainties is required. Modellers and those collecting empirical data need to define accurate error models for constraint and input data.

A key challenge is to effectively quantify the structural uncertainties in numerical models.

Structural uncertainties, such as those due to limited model grid resolution and missing processes, account for the residual discrepancy between the model and ‘reality’ with the best-fit set of model parameters. Within the climate and hydrological modelling, non-exclusive approaches include: (i) hierarchical models that parameterise the form of the structural error (Rougier, 2007; Hauser et al., 2011), (ii) extraction of variance components from multi-model ensembles (i.e. involving models with different parameterisations and numerical formulations: Sexton et al., 2012), and (iii) minimization of structural uncertainties through the introduction of parameterisations, new processes, and otherwise more complete models (e.g., Briggs et al., 2013). The interpretation of probabilities from ensemble-based results relies on the assumption that, in some sense, the ensemble contains model runs that are close to or at least bracket ‘reality’. However, it is unclear how to rigorously test such an assumption even when observations are fully bracketed (after accounting for uncertainties) by ensemble model results. A trend to increase the degrees of freedom in the model offers an obvious (though expensive) route towards this bracketing. However, given the nonlinearity of such systems, increasing model complexity can also make it more difficult to find model configurations that are close to ‘reality’. Therefore, significant misfits to at least a subset of observations are likely to persist (e.g. Tarasov et al., 2012).

One challenge that is relatively easy to address is the compilation of constraint data in an accessible and well-documented digital form that is easy to import for model use. A recent example is the initial version of the AntCal database of palaeo-data for Antarctica (Briggs and Tarasov, 2013). Other examples include the ALBMAP (LeBrocq et al., 2010) and BEDMAP II (Fretwell et al., 2013) data sets for Antarctica and Greenland. Likewise, the recent expansion of coverage of measurements of present-day vertical velocities from continuous or repeat GPS (e.g. Kierulf et al., 2014) has significantly improved constraint data for GIA-based ice sheet reconstructions (Argus et al., 2014; Peltier et al., 2015).

In terms of data-model integration, therefore, there are some key issues that ought to be explicit in both the modelling and empirical data-gathering communities. Studies using numerical ice sheet models ought to consider:

- What assumptions and approximations are made and what are the main sources of uncertainty?
- What is the uncertainty distribution for the input data, and for the output of the numerical model? Do the latter adequately address model approximations and associated uncertainties?
- What efforts have been made to better relate model output directly to palaeo-records?
- Given the above, how robust are the results? What has been done to ensure that the error bars (e.g. 2 sigma bounds) really envelop reality? To what extent has the full distribution of results been analyzed as opposed to excessive focus on a mean or single 'best-fitting' model run?
- Where to make the model output available for further scrutiny by the community, with appropriate embargos to allow full publication of results, etc.

Those providing empirical data should consider:

- Outlining the assumptions in terms of data interpretation and clearly specifying/quantifying errors in any resultant reconstruction.
- Introducing greater transparency in terms of data quality and quantity. Unfortunately, data quality often loses out to quantity. Clearly, there is a trade-off, but high quality data (i.e. with tight error bars and a strong signal) are generally more valuable than a large quantity of low quality data. Those collecting empirical data can also use the uncertainty maps from calibrated modelling studies to prioritize data collection and ensure that their efforts will offer the greatest constraint value for the integration of

data and modelling.

- Depositing data in centralized online data-bases. To ensure data-utilization, the maintenance and updating of online data-bases is very important. Servers such as the world data centres http://www.ncdc.noaa.gov/palaeo/icgate.html are a good example, but need to encourage detailed uncertainty specifications, and standardization of data formats.

An obvious issue in terms of data-model integration is that there is a large difference between the empirical datasets available for reconstructing ice sheets during their growth stages, as opposed to their decay stages. This is a particular challenge for constraining numerical models of ice sheet build-up, but a well-calibrated deglacial (post-LGM) record can be used as a target to help constrain the prior evolution of the ice sheet (Stokes et al., 2012).

5.9. Concluding Remark

Since Andrews (1982) pioneering review of the techniques to reconstruct ice sheets, there have been major technical advances in numerical modelling and a rapid expansion in the size and qualitative diversity of datasets for constraining such models. This paper has reviewed some of the major developments in techniques to reconstruct palaeo-ice sheets in order to evaluate the extent to which terrestrial and marine records have been integrated with numerical modelling. Whilst various disciplines have made important progress in our understanding of ice-sheet dynamics (both in terms of specific regions and specific processes), a well-constrained deglacial ice sheet chronology with adequately quantified uncertainties does not yet exist.The representation of past climate will continue to be the largest source of uncertainty for numerical modelling. Therefore, palaeo-observations will continue to be critical to constrain and validate modelling. Current state-of-the-art glaciological models continue to improve in model resolution and in the breadth of inclusion of relevant processes, thereby enabling more accurate and more direct comparison with the increasing range of palaeo-observations. Thus, the capability is developing to use all relevant palaeo-records to more strongly constrain deglacial (and to a lesser extent pre-LGM) ice sheet evolution.

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Figures:

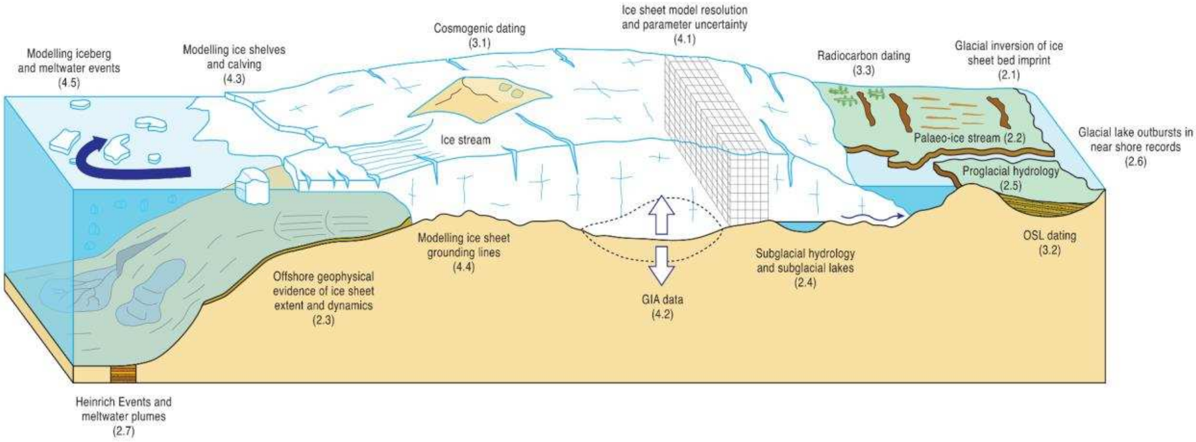


Figure 1: Simple schematic of an ice sheet (not to scale) showing some of the key domains and approaches that are used to constrain the extent, thickness and dynamics of ice sheets at the regional to continental scale. Numbers in brackets refer to sub-sections in the manuscript where recent developments in each of these areas are discussed.

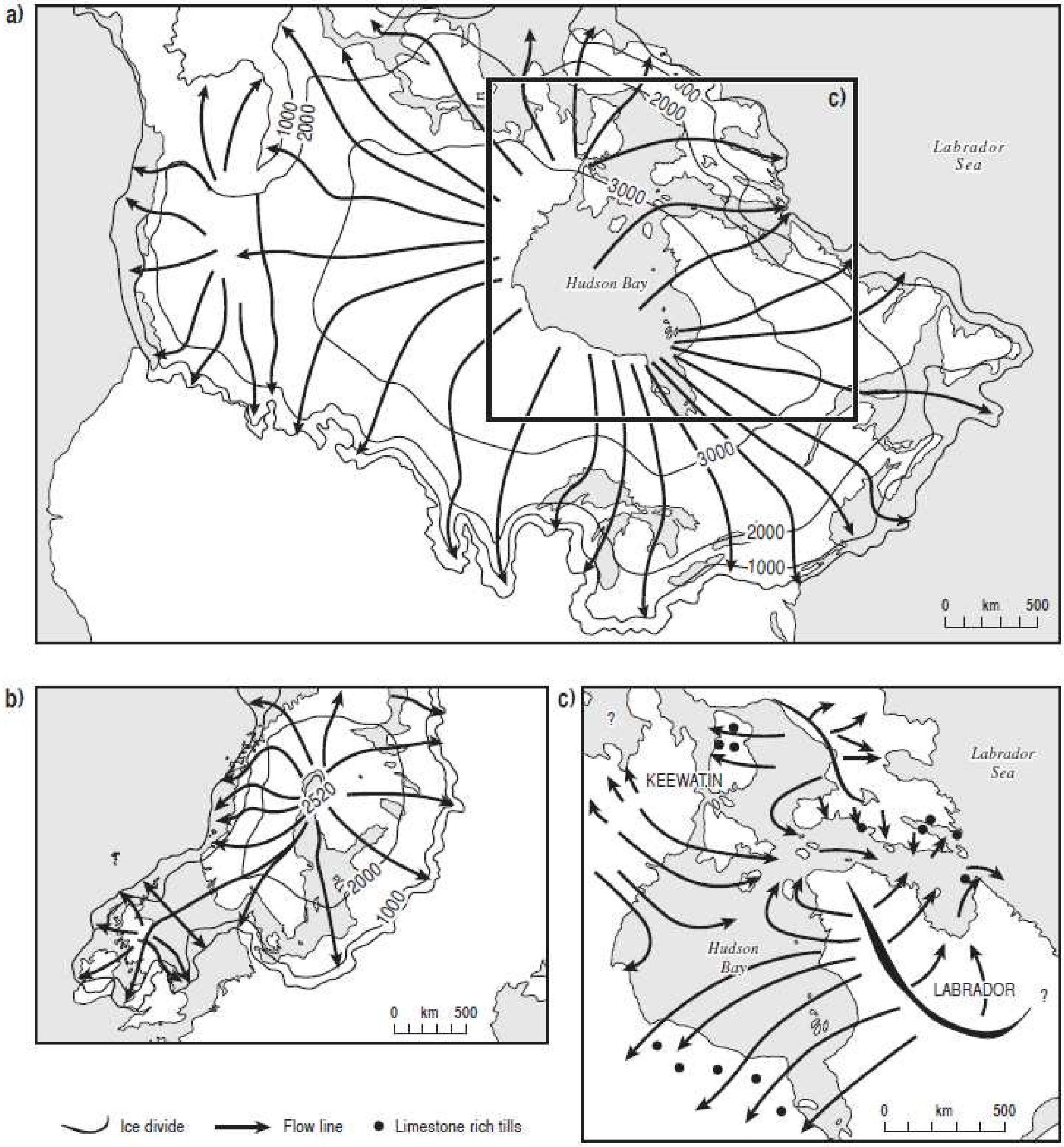
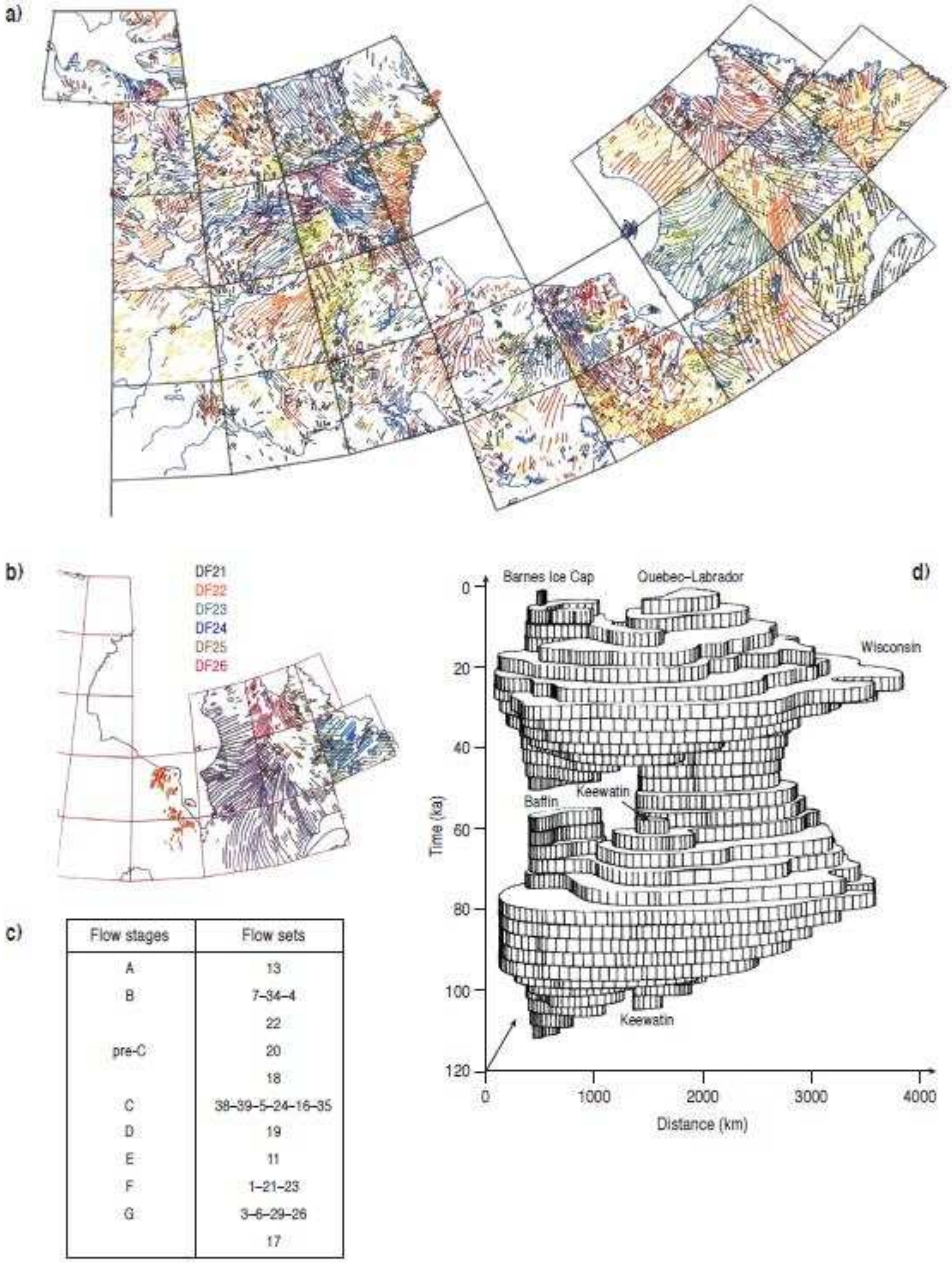


Figure 2: Map of modelled ice thickness (metres) and flow-lines of the North American (**a**) and Eurasian Ice Sheets (**b**) after Denton and Hughes (1981). Andrews (1982) noted the mismatch between the single-domed ice sheet configurations and flow patterns inferred from till stratigraphy and erratic dispersal data (e.g. Shilts, 1980), shown in (**c**). Figures redrawn from Andrews (1982).



Flow stages	Flow sets
A	13
B	7–34–4
	22
pre-C	20
	18
C	38–39–5–24–16–35
D	19
E	11
F	1–21–23
G	3–6–29–26
	17

Figure 3: (**a**) Multiple cross-cutting flow-sets (colours used arbitrarily to show different flow-sets) from the central part of the Laurentide Ice Sheet mapped by Boulton and Clark (1990b), including close-up in (**b**). These flow-sets are numbered (e.g. DF21, DF22 in (**b**)) and stacked into relative age sequence (flow-stages) (**c**) using cross-cutting relationships, and reveal a highly mobile ice sheet reconstruction, which shows the areal extent of the ice sheet through time in (**d**). Figures redrawn from Boulton and Clark (1990b).

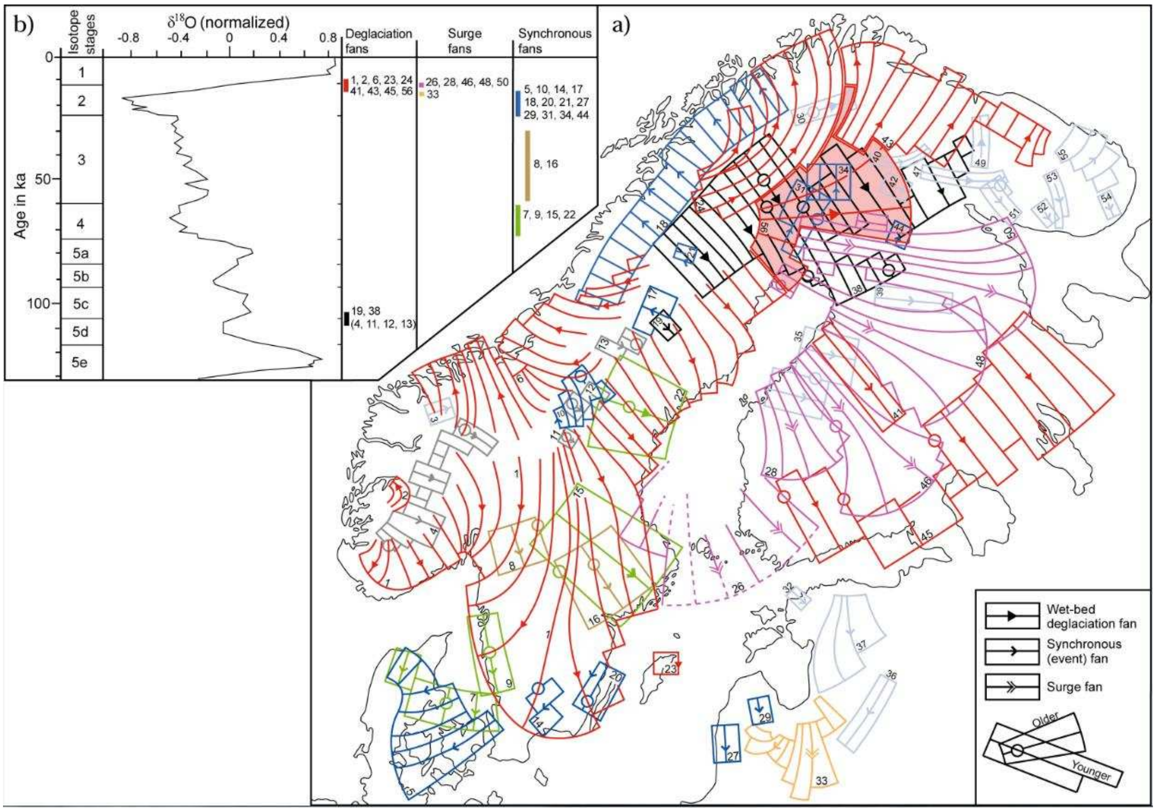


Figure 4: Application of Kleman and Borgström's (1996) glacial inversion of flow-sets (fans) to the Fennoscandian Ice Sheet, from Kleman et al. (2006). Different fan types are 'unfolded' to produce a time-slice sequence of ice sheet evolution that spans from 115 to 9 ka.

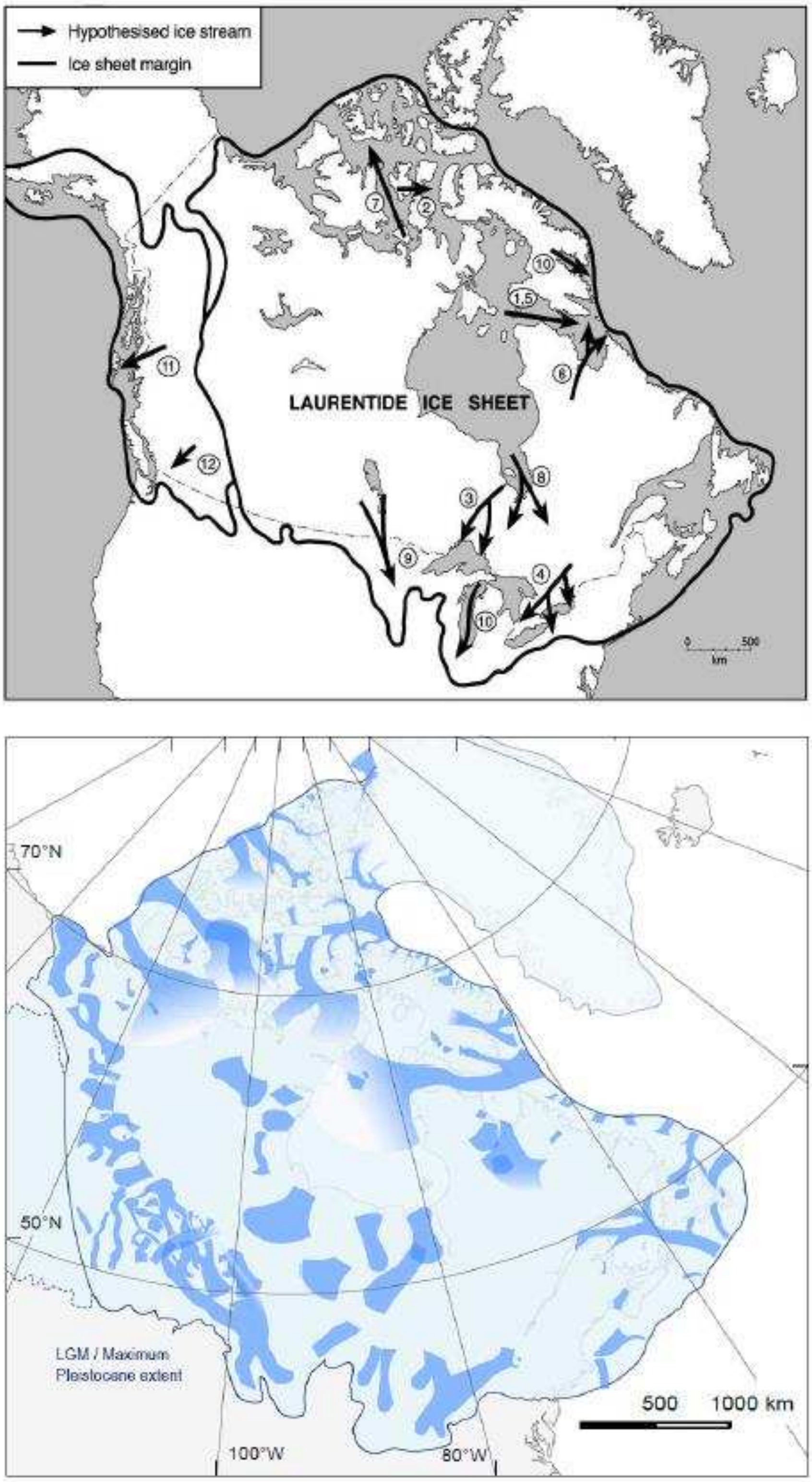


Figure 5: The number of hypothesised ice streams has grown rapidly over recent years, with those in the Laurentide Ice Sheet increasing from 10 (black arrows) that were reviewed by Stokes and Clark (2001) in (**a**), to over 100 (blue shading) in the latest inventory compiled by Margold et al. (2015) in (**b**).

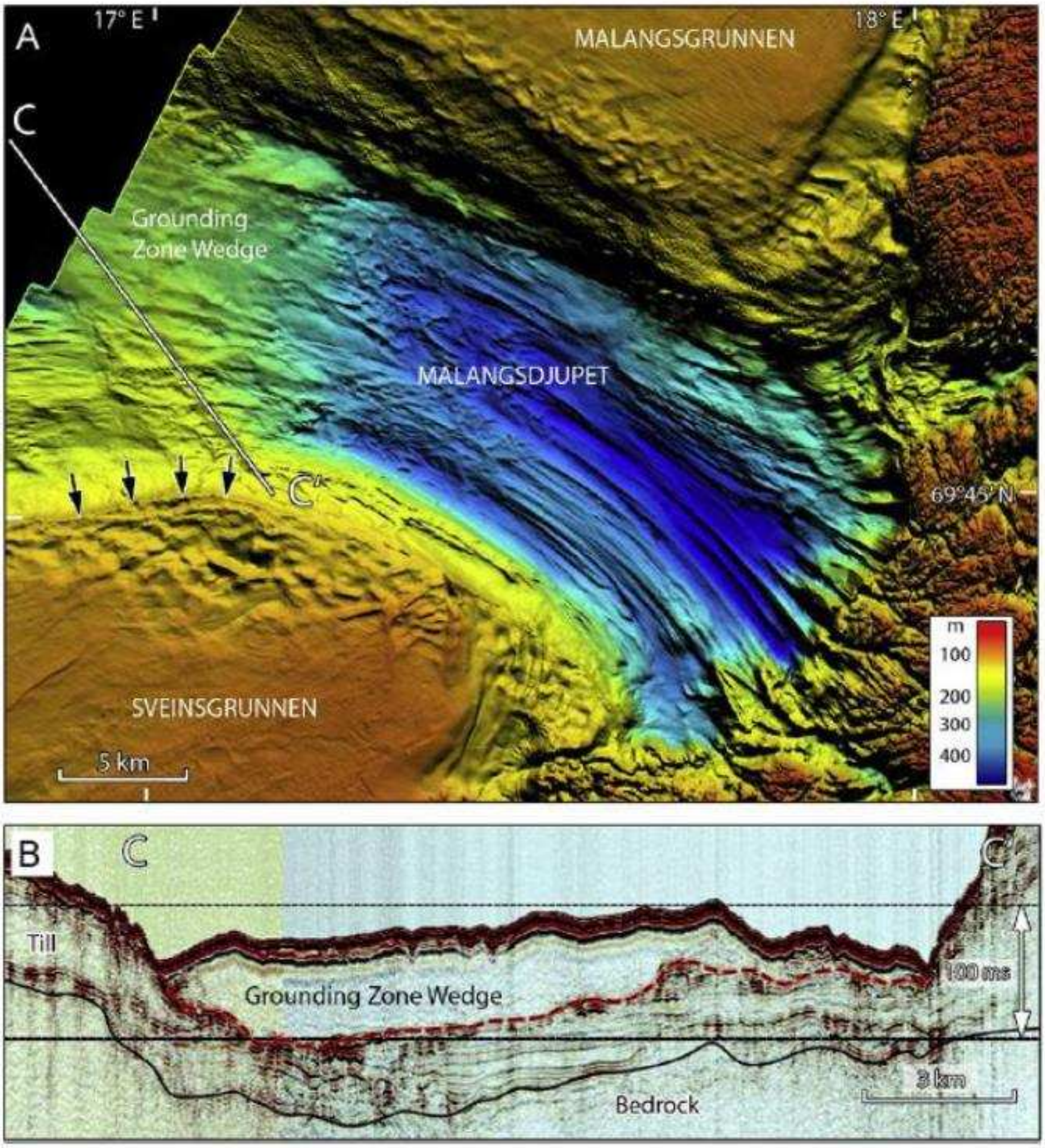


Figure 6: (**a**) Multi-beam bathymetric data showing a submarine palaeo-ice stream bed (ice flow bottom right to top left) in Malangsdjupet, northern Norway (from Ottesen et al., 2008). This cross-shelf trough exhibits many of the geomorphological criteria for identifying palaeo-ice streams (Stokes and Clark, 1999), including a convergent onset zone feeding into a main trunk characterised by mega-scale glacial lineations with abrupt lateral margins and ice stream shear margin moraines (black arrows); (**b**) Seismic data across the grounding zone wedge (100 m s = ~90 m on the vertical axis).

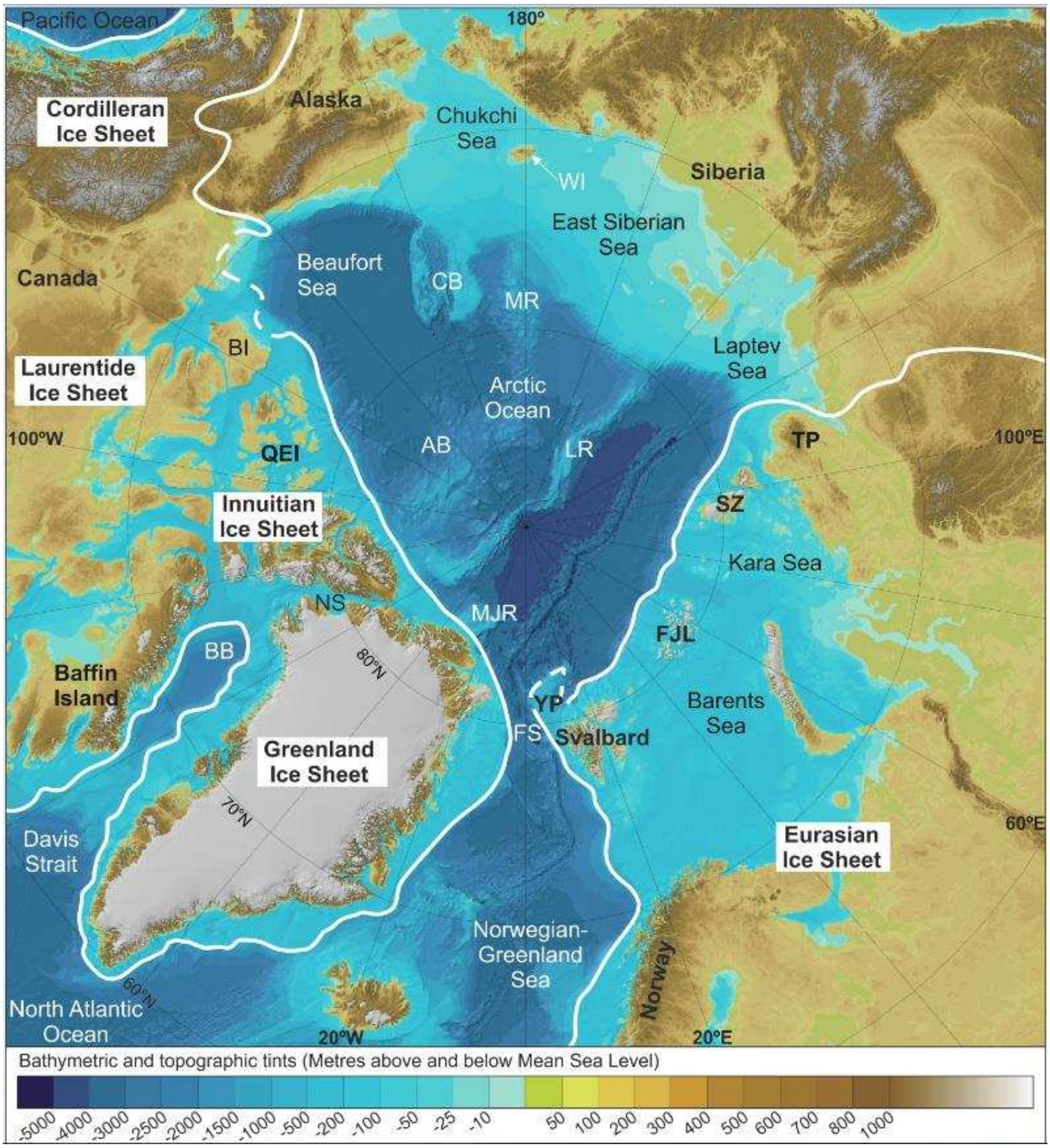


Figure 7: Maximum Quaternary extent (white line; dashed in areas of uncertainty) of the Eurasian (MIS 6: Svendsen et al., 2004), Greenland (MIS 2: Ehlers and Gibbard, 2004), Innuitian and Laurentide (MIS 2: Dyke et al., 2002), and Cordilleran ice sheets (Pleistocene maximum from Manley and Kaufman, 2002) in the high Arctic from Batchelor and Dowdeswell (2014) displayed on IBCAO bathymetric data (Jakobsson et al., 2012), with place names (WI = Wrangel Island; MR = Mendeleev Ridge; SZ = Severnaya Zemlya; FJL = Franz Josef Land; YP = Yermak Plateau; FS = Fram Strait; MJR = Morris Jesup Rise; NS = Nares Strait; QEI = Queen Elizabeth Islands; BB = Baffin Bay; CB = Chukchi Borderland; LR = Lomonosov Ridge). Figure modified from Batchelor and Dowdeswell (2014).

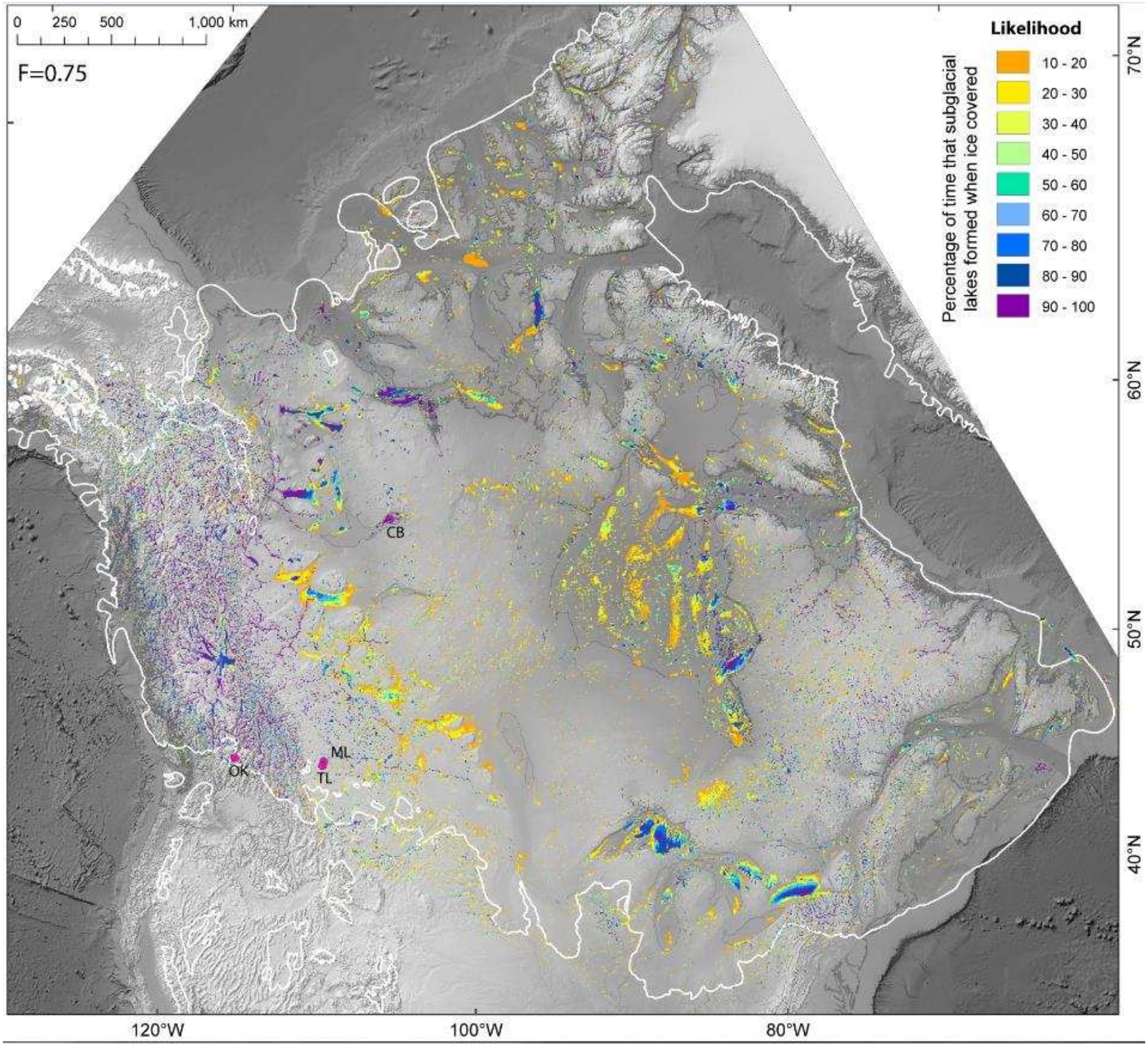


Figure 8: Modelled subglacial lake likelihood map for the North American Ice Sheet Complex from 32 to 6 ka from Livingstone et al. (2013). F is the flotation criteria which is the ratio of non-local, subglacial water pressure to ice overburden pressure (i.e., if F=1, the water pressure is at the ice-overburden pressure). White line shows ice extent from Dyke et al. (2002). Pink circles refer to published palaeo-subglacial lake records: CB = Christie Bay, Great Slave Lake; ML: McGregor Lake; TL: Travers Lake; OK: Okanagon.

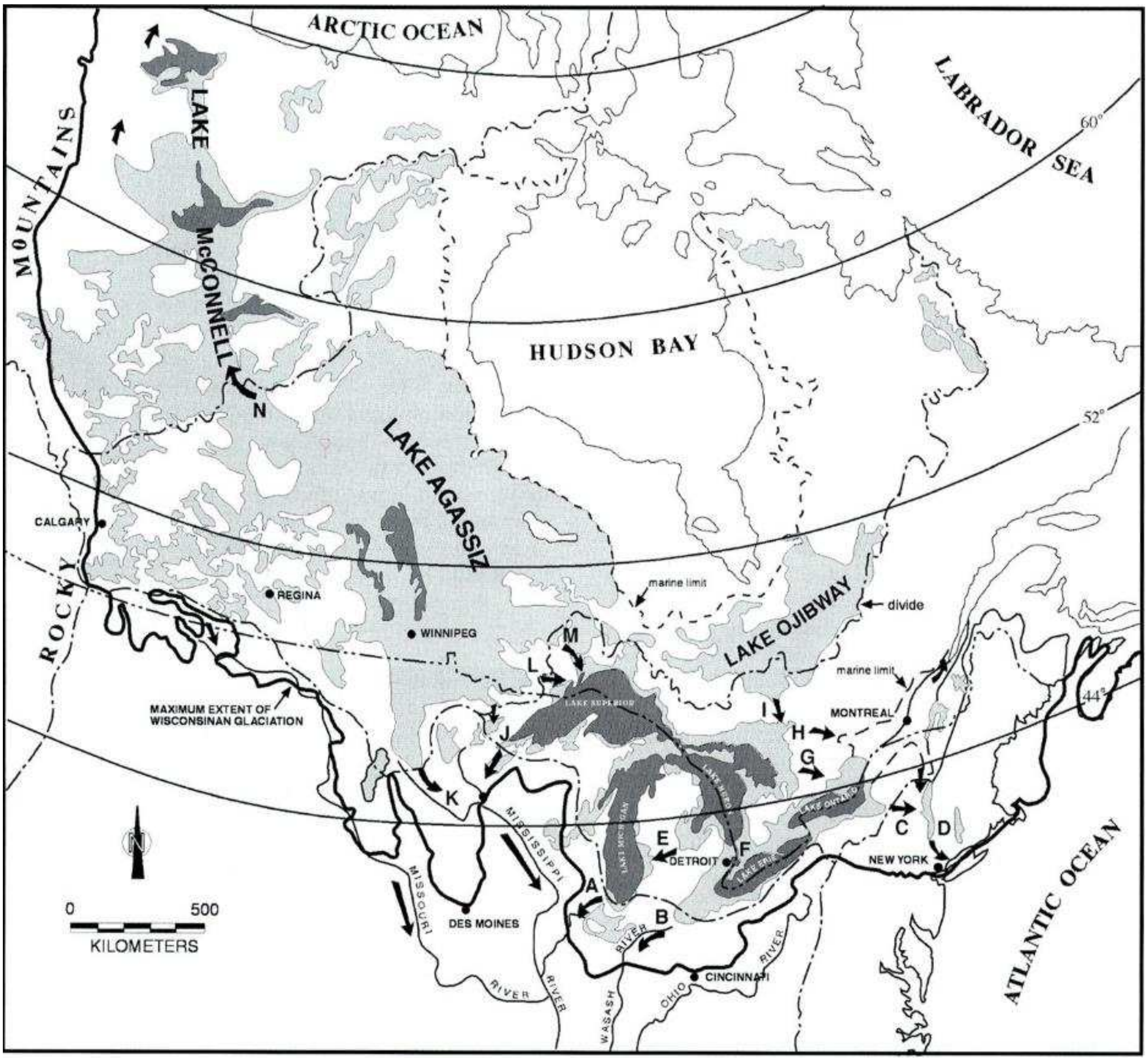


Figure 9: Total area covered by large proglacial lakes in North America (grey) during the last retreat of the LIS (from Teller, 2004, Fig. 1, after Teller, 1987). Major continental divides shown by dash-dot lines. Names of the major lakes are shown in the region where they formed. Major overflow routes from lakes are shown by arrows; letters identify their names as follows: A = Chicago outlet, B = Wabash River Valley, C = Mohawk Valley, D = Hudson Valley, E = Grand River Valley, F = Port Huron outlet, G = Fenelon Falls outlet, H = North Bay outlet, I = Temiskaming outlet, J = Duluth outlet, K = Minnesota River Valley, L = Kaminiskwia outlet, M = eastern Agassiz outlets, N = Clearwater outlet. Extent of proglacial lakes in Hudson Bay Lowland and St. Lawrence Lowland are not shown where lacustrine sediments are now buried by marine sediment.

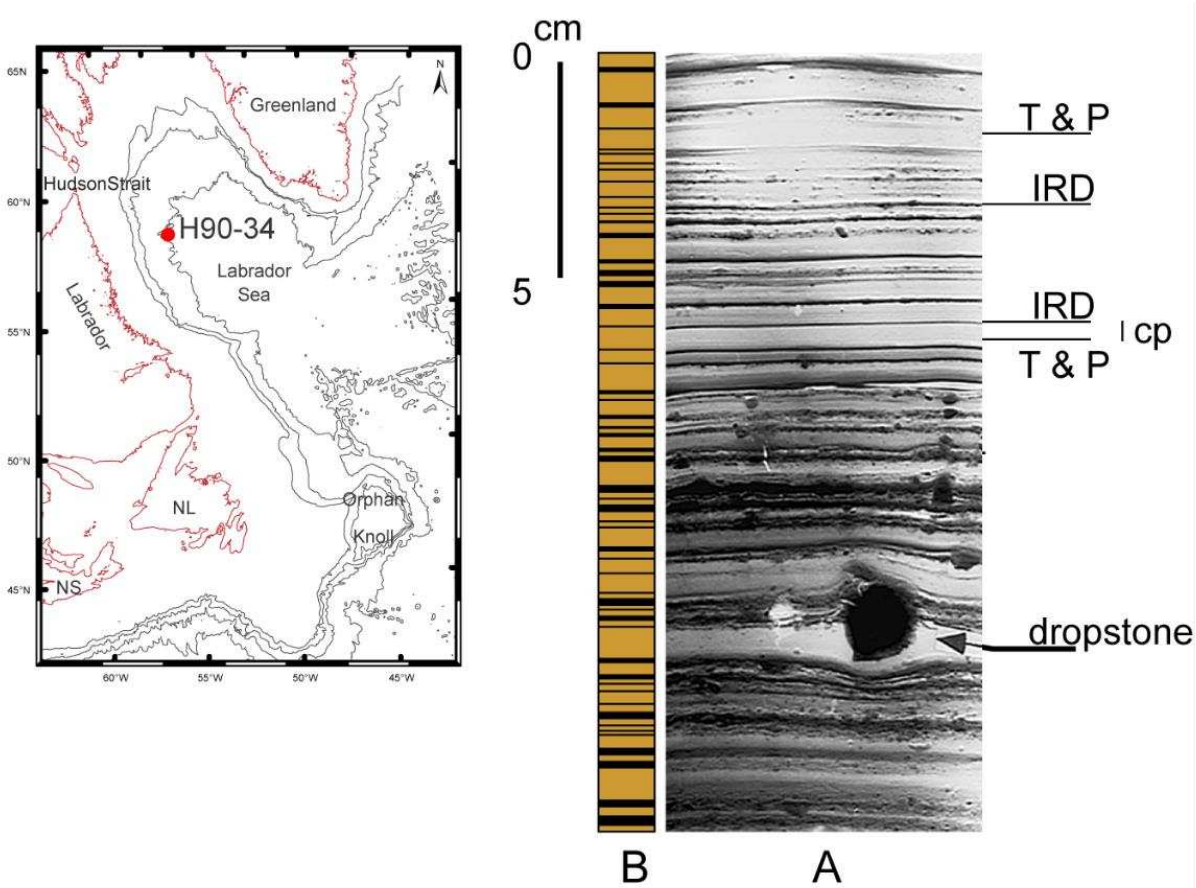


Figure 10: Example of fine-scale structure within a Heinrich Layer from the Labrador Sea that can be several metres thick (modified from Hesse and Khodabakhsh, 1998). Left panel shows the location map. Right panel shows: (**a**) X-radiograph showing the rhythmic succession of individual IRD layers (dark) and turbidite and/or meltwater plume deposit layers (light). The existence of individual IRD laminae is possible only if there is enough time between the occurrences of the turbidity currents, otherwise the IRD would be incorporated in the turbidites as drop-stones or reworked by the turbidity currents. (**b**) a large number of light-dark couplets are identified over a 17 cm interval. Additional work is required to determine the average duration of a couplet, but it is likely that they are annual (T and P = turbidites and meltwater plume deposits; IRD = ice rafted debris; cp = couplet).

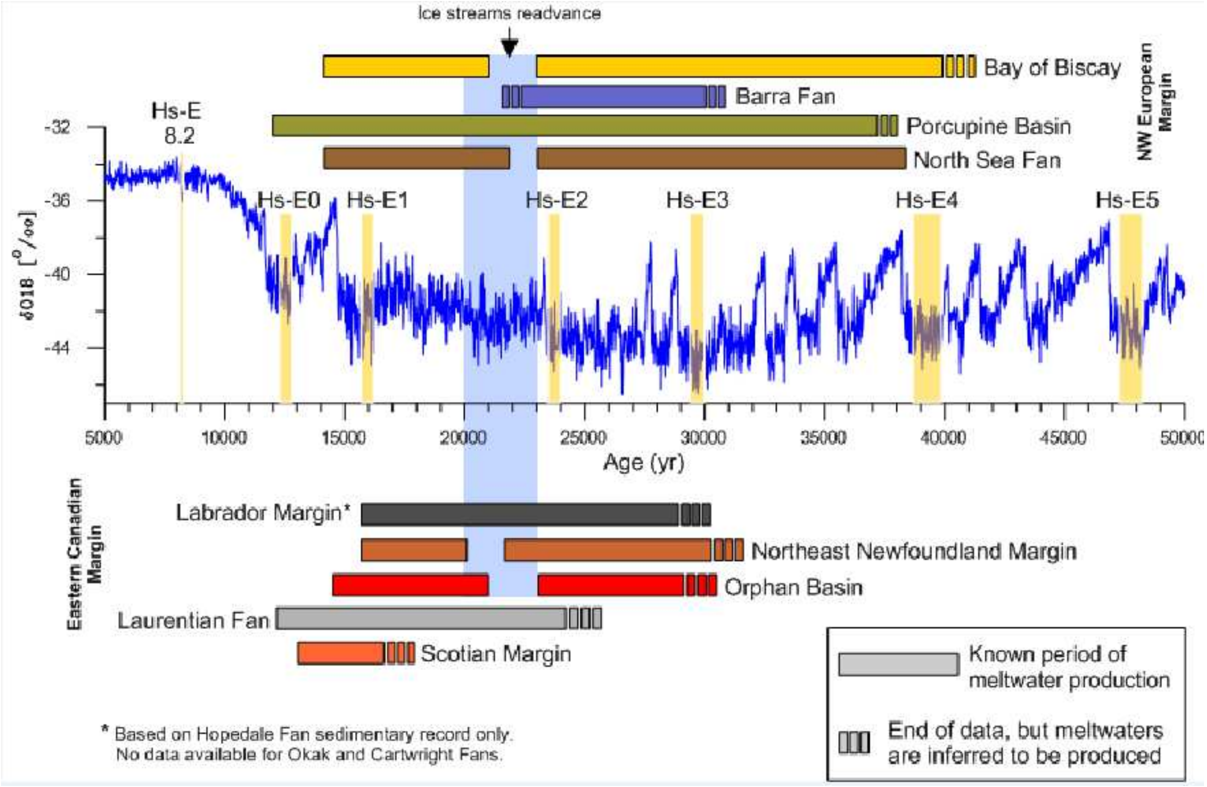


Figure 11: Major meltwater and iceberg (IRD) events from the Laurentide and Eurasian Ice Sheets from 50 to 5 ka (Hs-E = Hudson Strait events). Compilation based on sedimentation rates from Skene and Piper (2003), Lekens et al. (2005; 2006), Piper et al. (2007), Toucanne et al. (2009), Tripsanas and Piper (2008a, b), Scourse et al. (2009), Haapaniemi et al. (2010), Saint-Ange and Piper (2011) and Roger et al. (2013). Different coloured bars for meltwater production emphasise that each source has a distinctive detrital petrology.

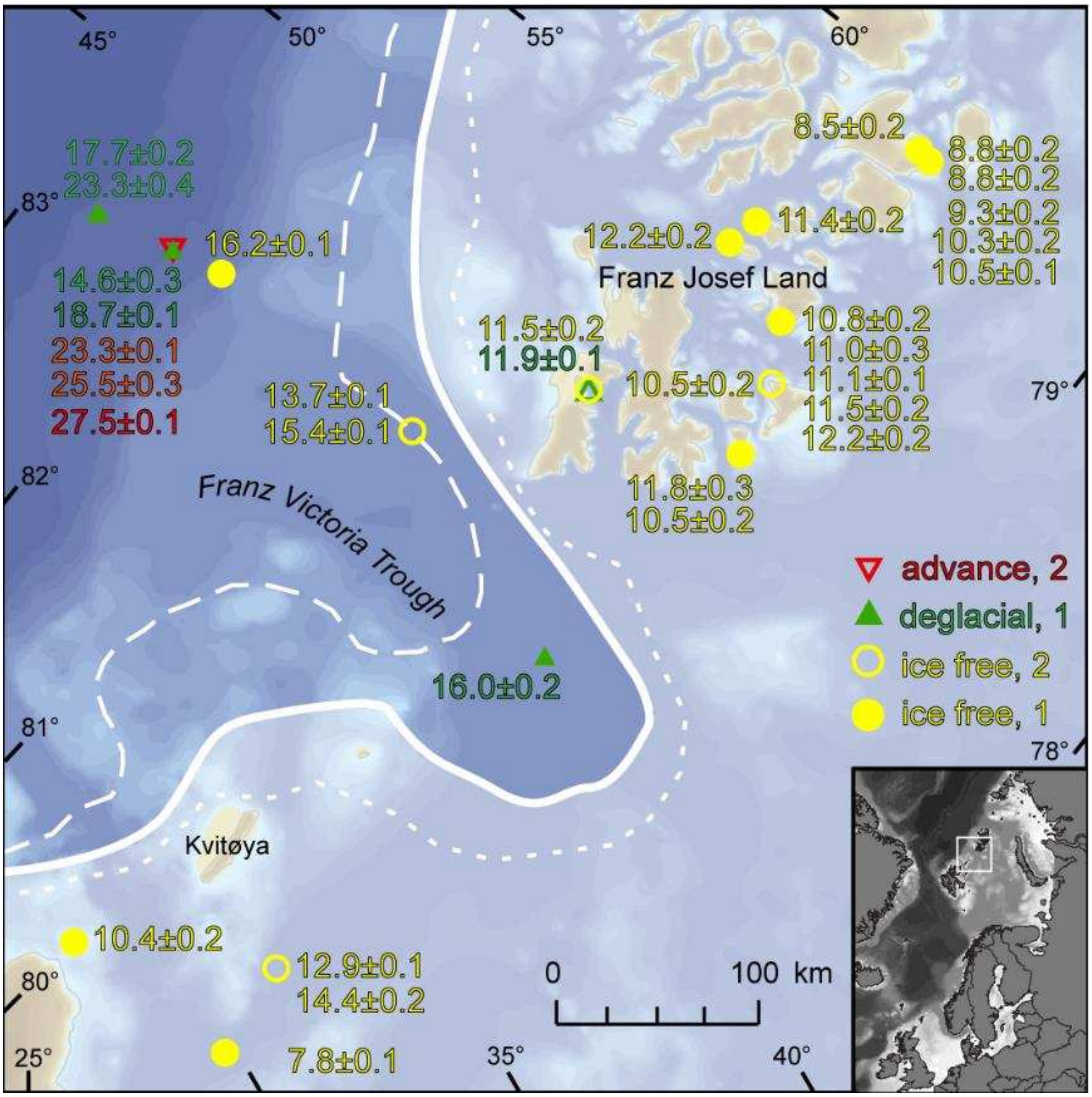


Figure 12: Excerpt from the DATED-1 ice margin reconstruction (Hughes et al., accepted) for 16 ka, east of Svalbard, showing ice margin positions interpolated between available dates. Maximum (dashed white line), most credible (solid white line) and minimum (dotted white line) ice margin positions are shown. Date sites and age labels (ka) are coloured by glacial classification (after Hughes et al., 2011): advance (e.g. below till; red); deglacial (e.g. above till; green); ice free (e.g. no stratigraphic information available; light green). Dates are considered reliable are shown by solid symbols (e.g. 1), dates that may be reliable are shown in outline (e.g. 2).

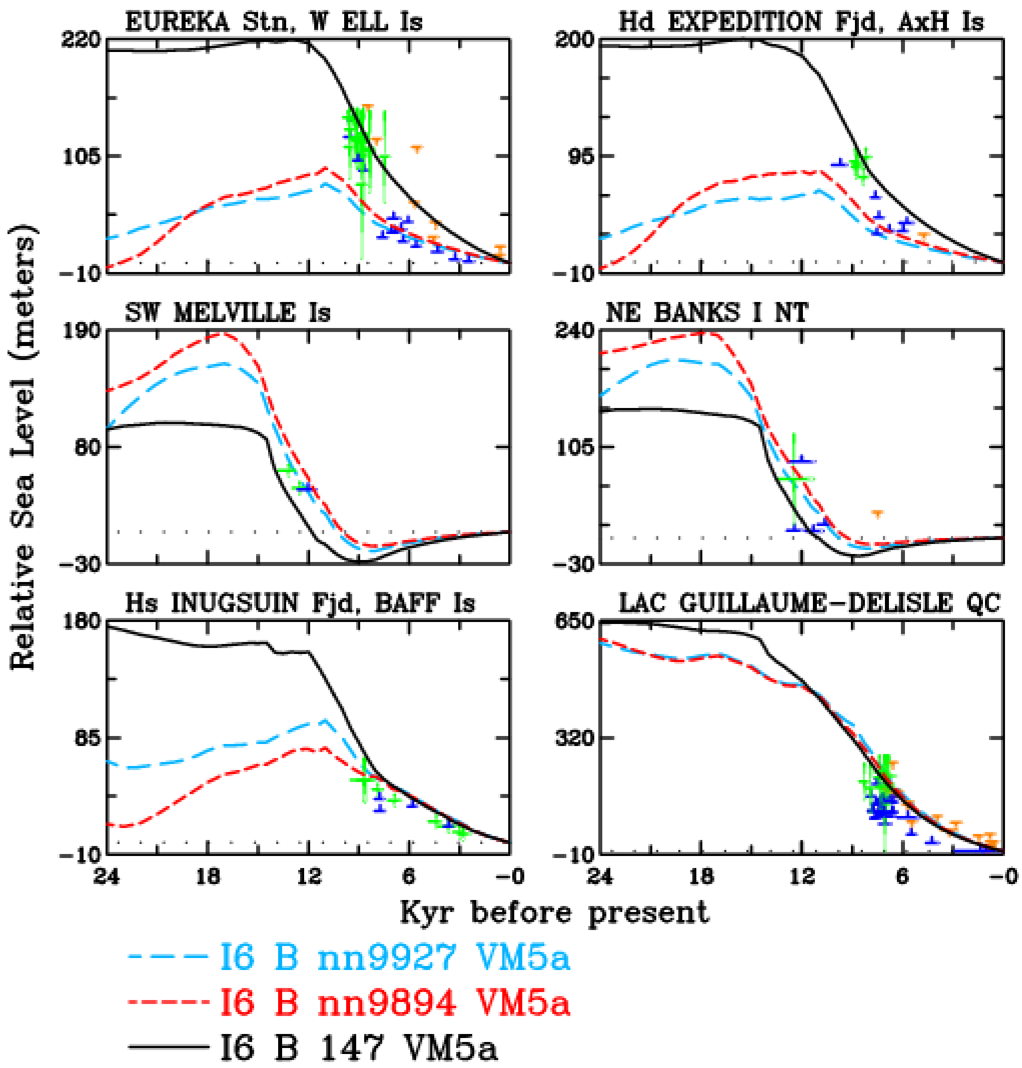


Figure 13: Comparisons of Glacial Isostatic Adjustment-based (GIA-based) and data-calibrated glaciological ice sheet modeling-based predictions of relative sea level (RSL) histories at six example locations on the North American continent that were once covered by the Laurentide and Innuitian ice-sheets. The black lines are the predictions of the GIA-based model. The blue and red dashed curves are for glaciological ice sheet models nn9927 and nn9894, respectively (Tarasov et al., 2012). These six comparisons are not intended to be representative of the differences between the quality of these different predictions at the totality of available locations from this geographical region. The site codes corresponding to the six site names shown here are 1160 (Eureka Stn.), 1186 (Hd Expedition Fjd), 1254 (SW Melville Is), 1356 (NE Banks Is), 1434 (Hs Inugsuin Fjd) and 1632 (Lac Guillaume-Delisle). These locations are shown as red circles on Supplementary Figure 1. A complete set of such comparisons is available in Supplementary Figure 2, which are shown in blue circles on Supplementary Figure 1.

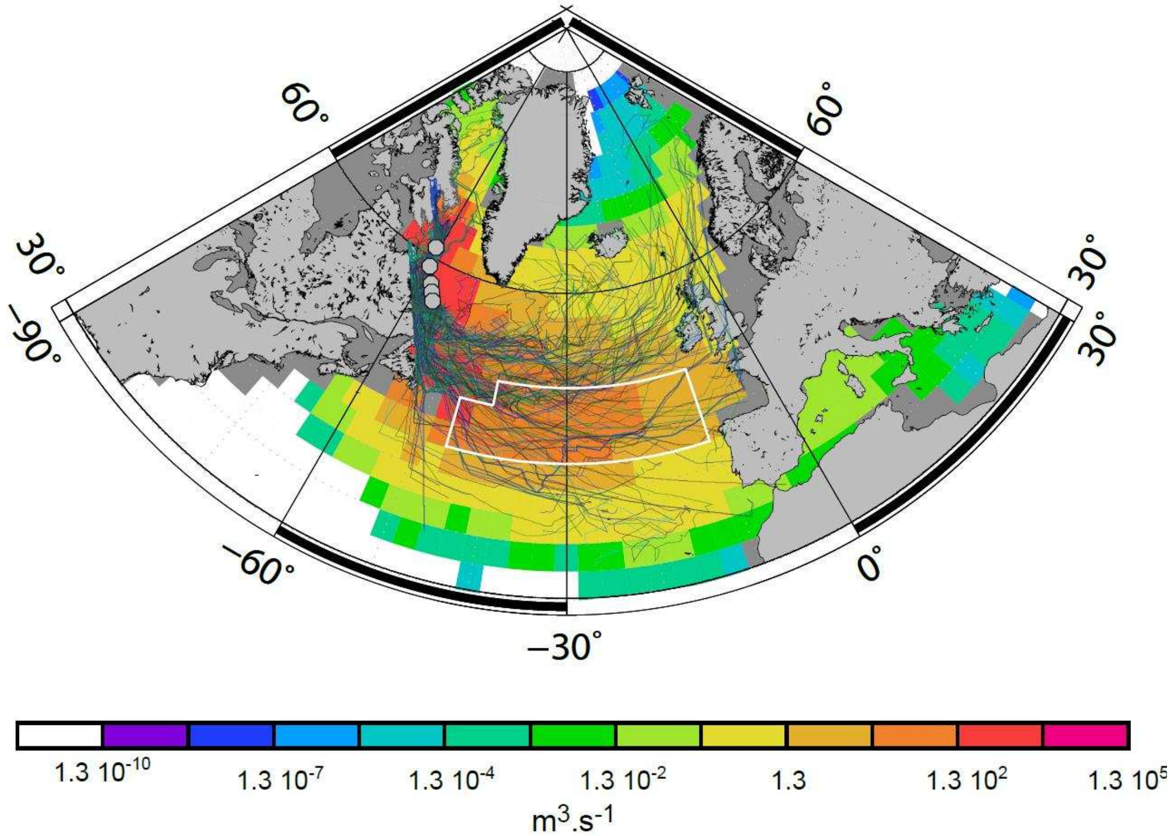


Figure 14: Simulated freshwater input and iceberg tracks in a Last Glacial Maximum climate simulated Heinrich event with interactive icebergs (from Jongma et al., 2013). The color scale indicates the freshwater flux input to the ocean in m^3 s^{-1} due to melting icebergs (note logarithmic scale). The five grey circles show the chosen fixed input locations for the iceberg delivery. The colored lines are a sample of iceberg tracks generated by the iceberg model. The white line in the center shows the area classically chosen for 'Ruddiman belt' freshwater hosing experiments for comparison with the colored area at the background.

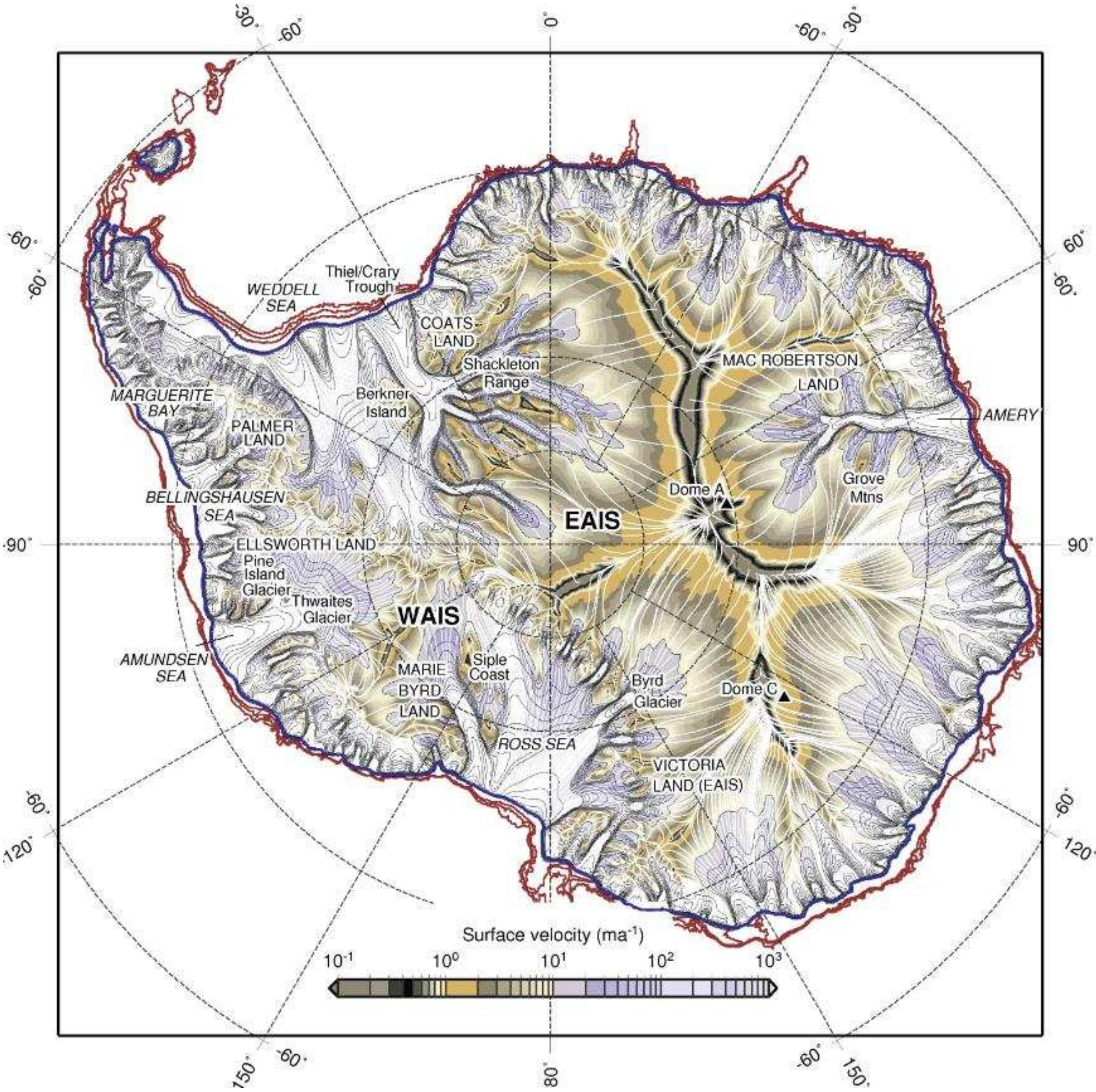


Figure 15: Surface velocity pattern of the modelled glacial maximum configuration ice sheet, with flow-lines illustrating steady-state surface transport routes from Golledge et al. (2013).