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DEGLACIATION OF THE LAURENTIDE ICE SHEET FROM THE LAST GLACIAL MAXIMUM

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ABSTRACT. *The last deglaciation of the Laurentide Ice Sheet (LIS) was associated with major reorganisations in the ocean-climate system and its retreat also represents a valuable analogue for understanding the rates and mechanisms of ice sheet collapse. This paper reviews the characteristics of the LIS at its Last Glacial Maximum (LGM) and its subsequent deglaciation, with particular emphasis on the pattern and timing of ice margin recession and the driving mechanisms of retreat. The LIS initiated over the eastern Canadian Arctic ~116-110 ka (MIS 5d), but its growth towards the LGM was highly non-linear and punctuated by several episodes of expansion (~65 ka: MIS 4) and retreat (~50-40 ka: MIS 3). It attained its maximum position around 26-25 ka (MIS 2) and existed for several thousand years as an extensive ice sheet with major domes over Keewatin, Foxe Basin and northern Quebec/Labrador. It extended to the edge of the continental shelf at its marine margins and likely stored a sea-level equivalent of around 50 m and with a maximum ice surface ~3000 m above present sea-level. Retreat from its maximum was triggered by an increase in boreal summer insolation, but areal shrinkage was initially slow and the net surface mass balance was positive, indicating that ice streams likely played an important role in reducing the ice sheet volume, if not its extent, via calving at marine margins. Between ~16 and ~13 ka, the ice sheet margin retreated more rapidly, particularly in the south and west, whereas the north and east underwent only minimal recession. The overall rate of retreat decreased during the Younger Dryas (YD), when several localised readvances occurred. Following the YD, the ice sheet retreated two to five times faster than previously, and this was primarily driven by enhanced surface melting while ice streams reduced in effectiveness. Final deglaciation of the Keewatin and Foxe Domes, left a remnant Labrador Dome that disappeared ~6.7 ka.*

La deglaciación del inlandsis Lauréntide desde el Último Máximo Glaciar

RESUMEN. *La última deglaciación del inlandsis Lauréntide (LIS) estuvo relacionada con grandes reorganizaciones en el océano-clima y su retroceso representa un valioso ejemplo para comprender las tasas y mecanismos del colapso del inlandsis. Este artículo revisa las características del LIS en su Último Máximo Glaciar (LGM) y su consiguiente deglaciación, con especial énfasis en* *el patrón y temporalidad de la recesión del margen de hielo y de los factores de retroceso. El LIS se inició en el este del Ártico canadiense hace ~116-110 ka (MIS 5d), aunque su crecimiento hacia el LGM no fue lineal sino afectado por distintos episodios de expansión (~65 ka: MIS 4) y retroceso (~50-40: MIS 3). Alcanzó su máxima posición alrededor de 26-25 ka (MIS 2) y existió durante varios miles de años como un extenso inlandsis con importantes domos en Keewatin, Foxe Basin y norte de Quebec/Labrador. Se extendió hasta el límite de la plataforma continental en sus márgenes marinos y probablemente almacenó el equivalente de unos 50 m de nivel del mar, con una máxima acumulación de hielo de ~3000 m sobre el actual nivel del mar. El retroceso desde su máximo fue desencadenado por un aumento en la insolación del verano boreal, aunque la contracción fue inicialmente lenta y el balance neto de masa superficial fue positivo, indicando que los ríos de hielo jugaron un papel importante en la reducción del volumen del inlandsis, si no en su extensión, por medio de fusión en los márgenes marinos. Entre ~16 y ~13 ka, el borde del inlandsis retrocedió más rápidamente, sobre todo en el sur y oeste, mientras el norte y el este experimentaron una recesión mínima. La tasa general de retroceso disminuyó durante el Younger Dryas (YD), cuando ocurrieron algunos avances locales. Después del YD, el inlandsis retrocedió a un ritmo dos veces superior al anterior, y se debió especialmente a un incremento de la fusión superficial, mientras los ríos de hielo redujeron su eficacia. La deglaciación final de los domos de Keewatin y Foxe dejó un único domo en Labrador que desapareció hacia 6.7 ka.*

Key words: Laurentide Ice Sheet, Last Glacial Maximum, deglaciation, ice streams.

Palabras clave: Inlandsis Lauréntide, Último Máximo Glaciar, deglaciación, ríos de hielo.

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1. Introduction

The North American Laurentide Ice Sheet (LIS) was the largest ice sheet to grow and decay during the last glacial cycle, dominating Late Pleistocene fluctuations in global sea-level (Lambeck *et al*., 2014) and delivering the largest contribution to early Holocene sea level rise (Tarasov *et al*., 2012; Peltier, 2004). Accurate reconstructions of its extent, volume and dynamics are, therefore, critical to our understanding of glacial-interglacial cycles and the sensitivity of ice sheets to climate change (Clark *et al*., 2009; Carlson and Clark, 2012). Knowledge of its deglaciation is also required to understand the rates, magnitude and mechanisms of ice sheet decay and associated impacts on sea level (Carlson *et al.,* 2008; Carlson and Winsor, 2012; Kleman and Applegate, 2013; Stokes *et al*., 2016), which is relevant to assessments of the future stability of modern-day ice sheets in Greenland and Antarctica (IPCC, 2013; Nick *et al.*, 2013; Ritz *et al.,* 2015). It is also clear that, in addition to responding to climate forcing, the behaviour of the LIS was capable of driving abrupt climate change through the delivery of both meltwater and icebergs that perturbed the ocean-climate system (Barber *et al*., 1999; Clark *et al*., 2001). More broadly, the configuration and retreat history of the LIS was an important constraint on the migration and dispersal of flora and fauna (Shapiro *et al*., 2004), including early humans (Goebel *et al.,* 2008; Eriksson *et al*., 2012; Dixon, 2013; Pedersen *et al*., 2016).

Given its size and importance, the LIS is one of the most widely-studied palaeo-ice sheets and there are hundreds of papers that have attempted to reconstruct its extent and dynamics using a variety of both empirical and modelling approaches (see review in Stokes *et al.*, 2015). However, the majority of studies, especially those taking an empirical approach, have tended to focus on specific regions and time periods, and fewer studies have attempted to summarise both the timing and driving mechanisms of deglaciation since the global Last Glacial Maximum (gLGM). Building on several major syntheses over the last few decades (Denton and Hughes, 1981; Dyke and Prest, 1987; Fulton, 1989; Dyke, 2004), this paper aims to provide an up-to-date review of the LIS at the gLGM with an emphasis on the pattern and timing of its deglaciation and the mechanisms that led to its demise. Following an overview of the characteristics of the LIS at its Local LGM (LLGM) in Section 2, Section 3 focusses on the pattern and timing of deglaciation, followed by a discussion of the mechanisms that have been invoked to explain deglaciation in Section 4. Some of the associated impacts of deglaciation, such as the origin of Heinrich events (e.g. Andrews, 1998) and major meltwater pulses and routing (e.g. Tarasov *et al.,* 2012; Gregoire *et al.,* 2012) are beyond the scope of the present paper and will receive less attention (see comprehensive reviews by Hemming, 2004; Carlson and Clark, 2012).

For the purposes of this paper, I use 'Laurentide Ice Sheet' in its broadest sense and, except where indicated explicitly. I include the Innuitian Ice Sheet (IIS) (Dyke *et al.,* 2002) and other small ice caps (e.g. in Newfoundland and the Appalachians) with which it was contiguous for most of its history. This does not include the Cordilleran Ice Sheet (CIS), which was a separate ice sheet except for brief periods during glacial maxima (Prest, 1969; Dyke and Prest, 1987; Dyke *et al.,* 2002; Stokes *et al*., 2012). For consistency, all dates are quoted in thousands of calendar years (ka) before present. Where the original source used only radiocarbon ages (e.g. Dyke and Prest, 1987), they have been converted to calendar years using a mixed marine and Northern Hemisphere atmosphere calibration curve (Stuiver *et al*., 2017) and the original radiocarbon dates appear in parentheses (^{14}C ka).

2. The Laurentide Ice Sheet at its Last Glacial Maximum

2.1. Inception and build-up to its Last Glacial Maximum

Before describing the characteristics of the LIS at its LGM, it is useful to briefly outline its inception and growth since the last interglacial during Marine Isotope Stage 5 (MIS 5). Unfortunately, this aspect of the ice sheet's history is very poorly constrained compared to the post-LGM period, largely because of the fragmentary nature of the terrestrial evidence relating to ice sheet build-up, most of which was erased by the much larger Late Wisconsinan (MIS 2) ice sheet. This has perhaps led to an over-reliance on numerical ice sheet models of pre-LGM ice sheet configurations (e.g. Marshall *et al*., 2000; Kleman *et al*., 2002; Stokes *et al.,* 2012) which are themselves limited by the availability of constraint data. However, the ocean-sediment record has proved particularly useful for investigating pre-LGM iceberg fluxes and meltwater events (e.g. Andrews and MacLean, 2003; Hemming, 2004), and there are pockets of evidence in the glacial geomorphological and stratigraphic record (e.g. Kleman *et al*., 2010) that have survived modification and, in some places, been dated to periods prior to the LGM (e.g. Allard *et al*., 2012; Dalton *et al*., 2016).

During the penultimate glacial maximum around 140 ka (MIS 6), the LIS is known to have been smaller than its LGM (MIS 2) counterpart, and is thought to have been similar in size to its extent around 13 ka (Colleoni *et al*., 2016). This is consistent with global sea level records and empirical evidence that indicates that the Eurasian Ice Sheet was larger during MIS 6 than during MIS 2 (Svendsen *et al*., 2004). Indeed, the smaller size of the LIS is consistent with changes in large-scale atmospheric circulation that facilitated the development of a larger Eurasian Ice Sheet during MIS 6 (Colleoni e*t al*., 2016). Little is known about the deglaciation of the LIS at the end of MIS 6, but a major glacial lake outburst flood has been reported from a proximal marine core in the Labrador Sea around 124 ka (Nicholl *et al*., 2012), which may be analogous to the widely reported drainage of glacial Lake Agassiz during the final deglaciation of the LIS around 8.2 ka (e.g. Barber *et al.,* 1999). Following this event, the general consensus is that there was virtually no ice cover in North America during the peak of the MIS 5 (Sangamonian) interglacial (~125-122 ka), which is primarily based on ages obtained from organic-rich sediments in the Hudson Bay Lowlands (e.g. Allard *et al*., 2011; Dalton *et al.*, 2016) and a widespread acknowledgement that global sea levels were 6-9 m higher than present during MIS 5e (e.g. Dutton *et al.,* 2015).

The consensus from both empirically-based arguments and numerical modelling is that the LIS initiated over the Arctic/sub-Arctic plateaux along the eastern seaboard of Canada (e.g. Ives, 1957; Ives *et al.,* 1975; Marshall *et al.,* 2000; Marshall and Clark, 2002; Kleman *et al.,* 2002; Stokes *et al*., 2012; Abe-Ouchi *et al.,* 2013). These are locations where only a small decrease in temperature resulted in a large decrease in the equilibrium line altitude (ELA) – a process termed 'instantaneous glaciation' (Koerner, 1980). It is thought that an embryonic dome formed over Labrador during MIS 5d (cf. Andrews and Mahaffy, 1976; Boulton *et al*., 1985; Vincent and Prest, 1987; Clark *et al*., 1993; Marshall *et al.,* 2000; Kleman *et al*., 2010), possibly as early as 116-114 ka, and with some modelling (Stokes *et al*., 2012) indicating a large but thin ice sheet at 110 ka that covered 70-80% of the area occupied by the MIS 2 ice sheet (see also Vincent and Prest, 1987; Boulton and Clark, 1990a, b; Clark *et al.,* 1993). This "explosive ice sheet growth" (Marshall, 2002: p. 133) during MIS 5d is consistent with records of a rapid fall in global sea level around that time (Marshall *et al*., 2000; Cutler *et al.,* 2003), but some workers suggest more minimal ice volumes in North America (~2-3 m of sea level equivalent: Kleman *et al.,* 2002) and that the LIS did not grow substantially until MIS 4 (e.g. Kleman *et al*., 2002; Marshall and Clark, 2002; Kleman *et al*., 2010). If the ice sheet was relatively large during MIS 5d (e.g.

the ~20 m of sea level equivalent modelled by Stokes *et al.,* 2012), it had shrunk rapidly by 100 ka (MIS 5c) (cf. St-Onge, 1987), and likely existed only as a small, thin ice sheet over the original inception grounds in north-eastern Canada by ~80 ka (MIS 5a) (Marshall *et al*., 2000; Stokes *et al.*, 2012). Thereafter, the LIS is thought to have grown rapidly during MIS 4, reaching a maximum extent around 65 ka (Vincent and Prest, 1987; Marshall *et al.,* 2000; Kleman *et al*., 2002; Stokes *et al.,* 2012), which coincides with the oldest recognised Heinrich event (H6) and a marked increase in ice-rafted debris from that time (Kirby and Andrews, 1999; Hemming, 2004, Bassis *et al*., 2017).

Following an MIS 4 maximum that may have been almost as large as the MIS 2 (LGM) volume according to some models (Marshall *et al.,* 2000; Stokes *et al.,* 2012), the ice sheet retreated to a mid-Wisconsinan (early MIS 3) minimum at some point between 60 and 40 ka (Dredge and Thorleifson, 1987; Clark *et al*., 1993; Kleman *et al*., 2010; Stokes *et al.,* 2012). Indeed, the extent of the ice sheet during MIS 3 is very poorly constrained (e.g. see review in Dredge and Thorleifson, 1987), with numerical modelling indicating a relatively large ice sheet that stored up to 30 m of sea level equivalent at 55 ka (Marshall *et al.,* 2000; Stokes *et al*., 2012), but with a suite of new dates raising the possibility that the Hudson Bay Lowlands, close to the geographic centre of the ice sheet, were completely ice free between ~50 and ~40 ka (see Dalton *et al*., 2016). Following the MIS 3 minimum, the ice sheet underwent gradual expansion that was punctuated by episodes of successively less recession (e.g. at 30 ka) before a final rapid growth towards the maximum LGM position (Dyke *et al*., 2002; Stokes *et al.,* 2012).

2.2. The timing of the Local Last Glacial Maximum Laurentide Ice Sheet (Late Wisconsinan)

In its broadest sense, the global LGM (gLGM) is conventionally defined from sea-level records “as the most recent interval in Earth history when global ice sheets reached their maximum integrated volume” (Clark *et al.,* 2009: p. 710). It has been recognised for some time, however, that because global sea levels are an integrated signal of ice volume, this does not imply that all ice sheets, or even various sectors within the same ice sheet, reached their ‘Local’ Last Glacial Maximum (from hereon LLGM) extent simultaneously (Clark *et al*., 2009; Hughes *et al.,* 2013). In a recent synthesis, Clark *et al*. (2009) constrained the timing of the gLGM period, based on relative sea-level data, as occurring from 26.5 to 19.0 ka, and suggested that this broadly coincided with the duration of maximum extent of most global ice sheets, including the LIS. They noted, however, that the LLGM of the various sectors of the LIS were asynchronous (albeit with large uncertainties), with some margins (e.g. in the south) potentially reaching their maximum early, perhaps even prior to, the gLGM and others occurring much later (e.g. the Maritime provinces in the south-east). Indeed, Dyke *et al.* (2002) suggested that ice advanced to its Late Wisconsinan (MIS 2) limit in the northwest, northeast and south about 27-28 ka (23-24 ^{14}C ka), and in the southwest and far north about ~24-25 ka (20-21 ^{14}C ka). More recently, a number of studies have shown that ice sheet margin in the far north-west, in the vicinity of the Mackenzie River delta and along the Richardson Mountains, attained its maximum position relatively late and certainly less than 20 ka (e.g. Murton *et al*., 2007; Kennedy *et al*., 2010; Lacelle *et al.,* 2013), possibly as a short-lived advance between 17 and 15 ka (Murton *et al.,* 2015).

Thus, the consensus is that - overall - the LIS reached its local maximum extent early in the gLGM period (cf. Dyke *et al.,* 2002), but with some margins advancing much later, e.g. in the far north-west. Most recent modelling experiments converge on maximum volumes ~26-25 ka (e.g. Tarasov *et al*., 2012; Stokes *et al*., 2012; Abe-Ouchi *et al*., 2013), although some place it closer to 21-20 ka (e.g. Marshall *et al.,* 2000).

It is very likely that the LIS existed at its near-maximum extent for several thousand years (cf. Dyke *et al.,* 2002; Tarasov *et al.,* 2012). Given that it grew to this position from a relatively large ice sheet late in MIS 3 (Dyke *et al*., 2002; Stokes *et al*., 2012; Tarasov *et al.*, 2012), the prolonged duration of its maximal configuration suggests that, for the most part, it had a surface geometry and mass balance in equilibrium with the gLGM climate for a few thousand years (Dyke *et al*., 2002).

2.3. Extent and thickness of the LIS at its Local Last Glacial Maximum (Late Wisconsinan)

The maximum extent and thickness of the ice sheet during its Late Wisconsinan maximum has been the subject of debate for over 150 years (e.g. Bell, 1884) and, despite numerous studies on this subject, consensus has only recently emerged (Dyke *et al.,* 2002). A comprehensive review of the literature on this subject is beyond the scope of this paper (see Ives (1978) and Dyke *et al.* (2002) for authoritative reviews), but it is useful to summarise key areas of contention and consider how different ideas have evolved and, more often than not, been revisited.

Much of the early work on the extent and thickness of the LIS during its LLGM (e.g. Bell, 1884; Daly, 1902; Coleman, 1920) focussed on the mountains of the east coast of Canada and argued that many of the highest peaks (e.g. the Torngat of northern Labrador) either remained as nunataks or were only affected by local ice caps or glaciers. This 'minimum' model was based on the identification of erosional trimlines (e.g. at 650 m in the Torngat Mountains, see Daly, 1902) and the presence of frost shattered bedrock and blockfields above these limits (e.g. Coleman, 1920). These interpretations were first questioned by Odell (1933) who reported high-level erratics and poorly preserved striations at 1446 m in the Torngat Mountains, thus arguing that the last ice sheet had overtopped the mountains, and that block-fields formed after deglaciation. Similar observations informed similar interpretations by Flint *et al*. (1942) in the Shickshock (Chic-Choc) Mountains, and Flint's hypothesis for the inception and growth of the LIS (Flint, 1943) called for a highland origin and windward-growth that subsequently inundated the high coastal mountains. This 'maximum' model (Fig. 1) was adopted in a series of major publications (e.g. Flint, 1947; 1971) and on a new 'Glacial Map of North America' (Flint *et al*., 1945), which was one of the first attempts (see also Chamberlin, 1913) to synthesise the glacial features in detail and on a large scale. An important corollary of Flint's 'maximum' model was that the ice sheet was viewed as a monolithic single-domed ice sheet centred over Hudson Bay, although earlier workers had suggested alternative multi-domed configurations (Tyrell, 1898; Coleman, 1920).

As noted by Ives (1978), Flint's maximum model appeared to have been widely accepted and clearly influenced the boundary conditions for the first CLIMAP (Climate:

Figure 1. An outline of Flint's (1971) portrayal of the maximum extent of the Laurentide Ice Sheet east of the Cordillera during the Quaternary (simplified and redrawn from Ives, 1978). Note that the Cordilleran Ice Sheet was not depicted in the original version.

Long range Investigation, MApping, and Prediction) reconstruction of the 'Ice-Age' earth (CLIMAP, 1976; Denton and Hughes, 1981), in addition to becoming firmly entrenched in high school and University curricula. Indeed, when Prest (1969) produced one of the most detailed maps of the retreat of the ice sheet (discussed in Section 3; see Fig. 7), his Late Wisconsinan limit extended, for the most part, on to the continental shelf along the east coast from northern Baffin Island, all the way down to the Atlantic provinces, and covered most of the Canadian Arctic Archipelago (apart from Banks Island). He also depicted an extensive southern margin that transgressed well into northern USA and as far south at 40 degrees.

Despite the ascendency of the maximum model from the 1940s, a number of papers in the 1950s, 1960s and 1970s (Ives, 1957; Andrews and Miller, 1972; Miller and Dyke, 1974) had questioned it on the basis that several different 'weathering zones' could be distinguished by their relative maturity (see review in Ives, 1978), with the oldest weathering zones interpreted to pre-date the LLGM and indicating ice-free refugia. In some locations, this interpretation was further strengthened by a small number of radiocarbon dates that gave ages much older than the Late Wisconsinan (e.g. Løken, 1966). Those arguing for a return to a more minimal model interpreted the high-altitude erratics and evidence of glacial abrasion (e.g. Odell, 1933) to be from a much older (pre-Late Wisconsinan) glaciation, although it was becoming increasingly recognised that they might also have been preserved beneath more recent cold-based ice (e.g. Sugden, 1977; Sugden and Watts, 1977).

Similar debates were being played out along other parts of the ice sheet margin, with multiple weathering zones and, in some cases, correlative till sheets being used to infer reduced ice sheet extent on the Queen Elizabeth Islands (England, 1976a, 1976b) and Banks Island (Vincent, 1982); and even at the south-western margin of the ice sheet, where drift previously thought to have been of Late Wisconsinan age was subdivided on the basis of morphological degradation, with the fresher drift delimiting the last ice cover (Stalker, 1977). Thus, in a comprehensive review of a rapidly-growing body of literature, and on the basis of a large amount of fieldwork carried out since the 1950s, Ives (1978) called for a return to the minimum model that had prevailed prior to the 1940s and with Late Wisconsinan limits well behind those proposed by Prest (1969) at the north, eastern and south-eastern margins of the ice sheet, i.e. with localised refugia around much of the ice sheet's perimeter and on the continental shelf. Ives (1978) also emphasised that the minimum model implied a much reduced ice thickness and that it was unlikely to have been a simple, monolithic dome with maximum ice thicknesses over Hudson Bay, as originally envisaged by Flint.

By the early 1980s, therefore, a large body of work had argued for a retraction of the Late Wisconsinan limit and "adherents of minimum ice sheet models" (e.g. Boulton *et al*., 1985: p. 452) adopted a more restricted margin, but this was not universally accepted and the debate continued (e.g. Hughes *et al.,* 1977; Denton and Hughes, 1981). Indeed, Dyke and Prest, 1987) noted that Prest (1984) was unable to portray a single Late Wisconsinan limit that met with any consensus and he instead showed a minimum and maximum limit, with the maximum similar to his 1969 reconstruction.

It was in the context of this highly contentious body of literature, that Dyke and Prest (1987) produced one of the most influential reconstructions of the pattern and timing of the LIS that would act as a benchmark for several decades. Their maximum extent at 21.4 ka (18 ^{14}C ka BP) was clearly influenced by the growing body of evidence for a retracted ice margin along the northern and eastern coasts of Canada, with the Torngat Mountains (and parts of the Appalachians, including the Shickshock Mountains) protruding as nunataks, and with large areas of Baffin Island and the Queen Elizabeth Islands ice-free, together with Banks Island (see Fig. 2). Acknowledging much larger uncertainty, Dyke and Prest (1987) also depicted major ice shelves in association with the Appalachian ice complex and others extending off the coast of Labrador, together with ice shelves in the Gulf of Boothia/Lancaster Sounds and in M'Clure Strait. Elsewhere, the southern margin (e.g. the Lake Michigan Lobe) extended south of 40 degrees (based on work by Clayton and Moran, 1982) and they depicted fully coalescent Laurentide and Cordilleran ice sheets at this maximum extent that was, at that time, far more controversial than it is now (cf. Stalker, 1977). A further significant component of the Dyke and Prest (1987) reconstruction was that it clearly portrayed a multi-domed configuration at its maximum, with centres of ice mass (domes) located over Labrador, Keewatin and Foxe Basin, and with major ice divides emanating from them (Fig. 2). This geometry attempted to reconcile new evidence from erratic dispersal trains that clearly indicated a complex multi-domed configuration (Shilts *et al*., 1979; Shilts, 1980). Dyke and Prest (1987) also discussed the importance of ice streams and the availability of 'soft' deformable sediments (cf. Fisher *et al*., 1985) in influencing the ice surface topography (see also Section 2.4), noting that many of the ice lobes at the southern margin of the ice sheet had extremely low ice surface gradients (Mathews, 1974).

Figure 2. Reconstruction of the North American ice sheets at 21.8 ka (18 ^{14}C ka), simplified and redrawn from Dyke and Prest (1987). The three major domes of the LIS over Quebec/Labrador (Q-L), Keewatin (K) and Foxe Basin (F) are labelled. Major ice divides shown in red with lower-lying 'saddles' in the ice sheet surface labelled 'S'.

The pendulum swung again in the mid-1990s (cf. Miller *et al*., 2002) when new lines of evidence were uncovered to interpret a more extensive Late Wisconsinan limit than had been portrayed by Dyke and Prest (1987), particularly at is northern margin, but also along the eastern margin and in the Atlantic provinces. As noted by Dyke *et al.* (2002: p. 11): "after a century of debate, intensively for the last 25 years, about the existence of an Innuitian Ice Sheet during the LGM over the northern half of the Canadian Arctic Archipelago, a consensus has emerged that such an ice sheet did in fact cover most of that region". This was based on new glacial geological evidence of ice streams within several inter-island channels and large fjord systems (e.g. Blake, 1992, 1993; Dyke, 1999; Lamoureux and England, 2000; Ó Cofaigh *et al.,* 2000) and numerous sets of lateral meltwater channels that descended to marine limits of early Holocene age (Dyke, 1999; England, 1999; England *et al.,* 2000). This appeared to offer conclusive evidence that the Innuitian Ice Sheet (IIS) extended offshore at its maximum extent and that it was fully coalescent with the Greenland Ice Sheet along Nares Strait in the east, and with Laurentide ice along Parry Chanel in the south, which was originally proposed by Blake (1970) and illustrated by Prest (1969), see Fig. 3. The limits of the ice sheet in the north-west were much more uncertain, but Dyke *et al.* (2002) portrayed the whole of Prince Patrick Island as ice-free, and large parts of Melville Island and most of Banks Island as unglaciated (Fig. 3).

The other region that benefitted from increased scrutiny from the mid-1980s onwards, particularly on the continental shelf, was the south-eastern margin of the ice sheet (cf. Miller *et al*., 2002). Dyke and Prest (1987) had portrayed large ice-free areas at the ice sheet's maximum extent, but new cosmogenic dating of intensively weathered terrains suggested that they could have been covered by cold-based ice (Gosse *et al.,* 1995). New ^{14}C AMS radiocarbon ages from marine sediments on the continental shelf off Nova Scotia and southern Newfoundland also dated sediment above the youngest till to be of post-LGM ages (e.g. Amos and Knoll, 1987; Bonifay and Piper, 1988; Gipp and Piper, 1989; Mosher *et al*., 1989; Amos and Miller, 1990; Forbes *et al*., 1991; Gipp, 1994; King, 1996; Stea *et al*., 1998; Josenhans and Lehman, 1999). To the north, sedimentological studies and high-resolution AMS dating of marine sediment cores from the SE Baffin and Labrador shelves, and adjacent slopes, led to a reconsideration of LLGM ice extent in that region (Dyke *et al*., 2002). Jennings (1993), for example, concluded that Cumberland Sound was filled by an ice stream until ~11.5 ka (~10 ^{14}C ka) which Kaplan (1999) suggested may have extended onto the continental shelf (see also Miller *et al*., 2002). In northern Labrador, Clark and Josenhans (1990) combined marine and terrestrial evidence to suggest that LGM ice was more extensive than previously mapped, with the ice limit extending onto the continental shelf. Cosmogenic exposure dating (Marsella *et al.,* 2000) also confirmed extensive Late Wisconsinan outlet glaciers in the fiords of Cumberland Peninsula.

Thus, numerous lines of evidence had been uncovered to suggest that previous assigned 'old' (pre-LGM) moraines on Cumberland Peninsula and northern Baffin Island were of Late Wisconsinan age, leading to the most significant re-interpretations of the ice extent along the north-eastern Laurentide margin for several decades, which was summarised in Dyke *et al.* (2002). The Late Wisconsinan extent portrayed in Dyke *et al.* (2002) (Fig. 3) was subsequently incorporated into the updated deglaciation sequence for North America (Dyke *et al.,* 2003). This proposed LLGM extent has also been reproduced in more recent overviews and syntheses (e.g. Dyke, 2004) and the margin positions have been used as constraint data for numerical modelling of the LIS (e.g. Tarasov *et al.,* 2012; Peltier *et al*., 2015).

The most dramatic changes to the LLGM extent of the LIS in the last decade has seen the ice margin extended to cover the >70,000 km^2 Banks Island in the Western Canadian Arctic (England *et al*., 2009; Lakeman *et al.,* 2012, 2013) and the recognition that it likely extended to (or close to) the continental shelf edge in Baffin Bay (Briner *et al.,* 2006) and in Atlantic Canada (Shaw *et al*., 2006) (see Fig 4). Banks Island had long been regarded as an ice-free refugium (Prest, 1969; Vincent, 1982; Dyke and Prest, 1987) and was portrayed as such in the most recent Dyke *et al.* (2003) synthesis (see also Dyke, 2004). A suite of new radiocarbon dates and glacial geomorphological mapping, however, has clearly indicated that the LIS inundated Banks Island during the LLGM (England *et al*., 2009; Lakeman *et al*., 2012, 2013). Similar methods and new dates have also extended the ice margin over the entirety of Melville Island and onto Eglington Island (Nixon *et al*., 2014) and it is likely that the ice also overran Prince Patrick Island and extended onto the continental shelf along the entire IIS margin (see Stokes *et al.,* 2016). At the same time, a large body of work has used cosmogenic dating of high-elevation erratics close to

Figure 3. Revised reconstruction of the North American ice sheets at 21.8 ka (18 ^{14}C ka) simplified and redrawn from Dyke et al. *(2002). Note the increase in the areal extent of the ice sheet compared to Figure 2, especially over the Queen Elizabeth Islands and at the south-eastern margin (e.g. over Newfoundland), but with Banks Island and Prince Patrick Island remaining ice-free.*

fjord mouths (e.g. on Baffin Island) to demonstrate that a relatively thick LIS must have terminated on the continental shelf during the LLGM (e.g. Briner *et al*., 2006), and with most high-elevation areas covered by non-erosive cold-based ice that accounts for the preservation of highly weathered surfaces in those locations. In the Atlantic provinces, a large body of work undertaken offshore has identified moraines, flutings, till tongues, cross-shelf troughs and associated trough mouth fans, which attest to warm-based ice streams extending to the edge of the continental shelf and separated by more stagnant ice on the shallow banks (e.g. Mosher *et al*., 1989; Piper and Skene, 1998; Schnikter *et al.*, 2001; Shaw *et al*., 2006). It is also noteworthy that data-calibrated modelling of the North American Ice Sheet complex at the LLGM (Fig. 5) also generates a large multi-domed ice sheet that extends to the edge of the continental shelf, with some of these major ice streams (see Fig. 4) captured in the basal velocity pattern (Stokes and Tarasov, 2010).

In summary, after over 150 years, consensus appears to have been reached that at its Late Wisconsinan maximum (~25-24 ka), the LIS was a large multi-domed ice sheet with a southern margin that extended south of 40° in the Great Lakes region, with a western margin that was fully coalescent with the Cordilleran Ice Sheet, and with northern and eastern margins that extended to the edge of the continental shelf (Fig. 4).

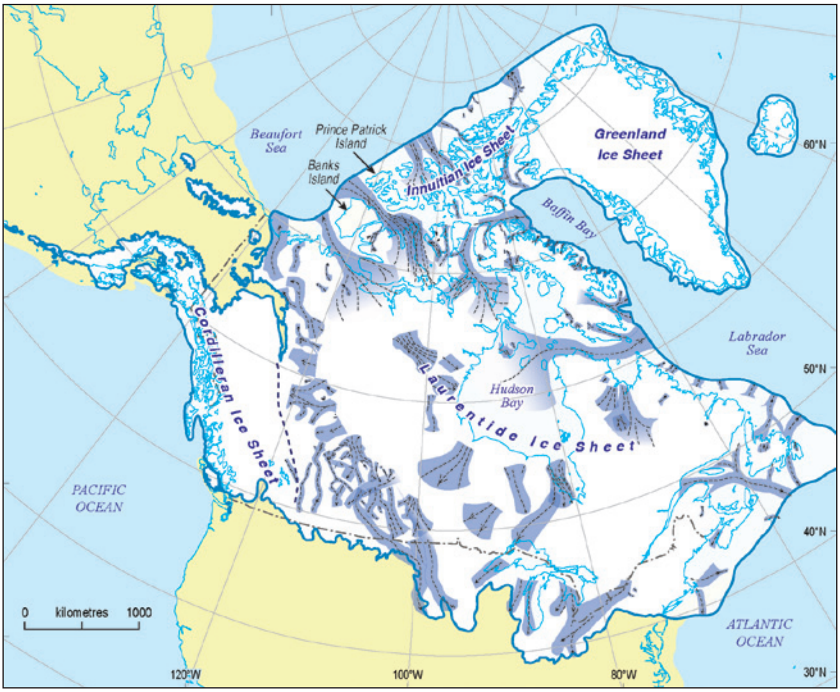

Figure 4. A recent reconstruction of the Laurentide Ice Sheet redrawn from Stokes et al. *(2016). Note the expansion of the ice sheet margin over Banks and Prince Patrick Island. This reconstruction also includes the location of 117 hypothesised ice streams (dark blue with flow-lines) in the Laurentide Ice Sheet based on published literature and new mapping in Margold* et al. *(2015a). Note that the ice streams did not all operate at the LLGM and that the inventory excluded the Cordilleran Ice Sheet.*

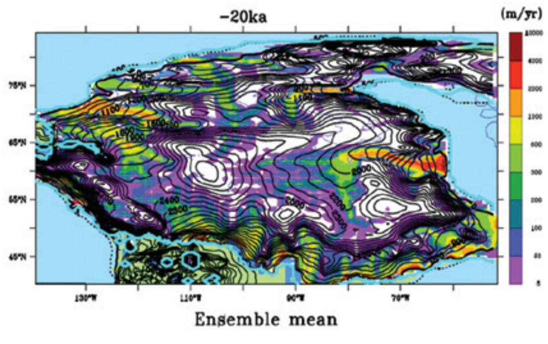

Figure 5. Weighted mean basal velocity and surface elevation of the North American Ice Sheet complex at 20 ka taken from Tarasov et al. *(2012). Note that this ensemble is not representative of a single glaciologically-self-consistent model run and that the weighted averaging also blurs ice stream locations and magnitudes, and smooths ice surface topography. It is simply the expectation value. Note, however, that the mean captures some of the major ice-streams (compare with Figure 4) and most of the key features of the geologically inferred reconstruction of Dyke and Prest (1987) (Fig. 3).*

2.4. Quantification of LIS volume at its Local Last Glacial Maximum (Late Wisconsinan)

Quantifying the volume of the LIS is important for reconciling records of global sea level (Carlson and Clark, 2012; Lambeck *et al.*, 2017), but this has often proved difficult due to the lack of nunataks and trimlines from interior regions (cf. Simon *et al.*, 2014). As noted by Dyke *et al.* (2002: p. 20): "all that can be concluded from direct mapping is that the vast interior region of the ice sheet, generally the part that was more than about 1000 km behind the margin, lay more than 2000 m above present sea level". Thus, despite numerous studies and debates regarding the extent and geometry of the ice sheet (see Section 2.3), few have attempted to quantify its volume.

Paterson (1972) was one of the first to consider the theory of ice flow and relate the area of the ice sheet to its thickness and volume. Using previous maps of ice sheet extent (from Prest, 1969), that were informed by Flint's (1943) large monolithic reconstruction, he argued that the LIS may have been up to 3.6 km thick at its maximum and comprised 26.5 x 10^6 km^3 of ice (with volume errors estimated at 16%). During the 1980s, however, the importance of 'deforming beds' was becoming increasingly recognised (cf. Alley *et al*,. 1986; Boulton and Hindmarsh, 1987) and this clearly influenced attempts to reconstruct the LIS, with important implications for its volume. In particular, numerical modelling experiments (e.g. Fisher *et al.*, 1985; Boulton *et al.*, 1985) clearly showed that the incorporation of deformable beds with low basal shear stress beneath the ice sheet generated thinner ice and removed much of the radial symmetry of some of the earlier reconstructions (e.g. Flint, 1943). Indeed, the difference between Fisher *et al.*'s (1985) 'soft bed' and 'hard bed' models of the LIS was a volumetric reduction of around 30% for the former. The incorporation of deforming beds also generated multi-domed configurations that were deemed to be more compatible with the emerging geological evidence at that time (e.g. Shilts *et al.*, 1979; Shilts, 1980; see Fig. 2). Later modelling work by Clark *et al.* (1996) and Licciardi *et al.* (1998) also demonstrated that a reduction in the effective viscosity of the till in regions underlain by 'soft' sediments generated a multi-domed ice sheet with a large bowl-shaped depression over Hudson Bay and thin ice (~1000 m above modern sea level) over the western and southern sectors of the ice sheet. A thinner, multi-domed ice sheet was also consistent with inverse modelling of crustal rebound and relative sea level data used in the early ICE-NG series (e.g. ICE-3G, Tushingham and Peltier, 1991; ICE-4G, Peltier, 1994).

More recently, Peltier's ICE-5G model of the ice load history (Peltier, 2004) indicated much larger ice sheet thicknesses over the Keewatin region (>4 km) and correspondingly larger volumes for the LIS. Tarasov *et al.* (2012) also noted that their modelled ice thicknesses over Hudson Bay were "possibly a kilometer too thick" (p. 37). Other work indicates thinner ice in this region (e.g. Lambert e*t al.*, 2006; Argus and Peltier, 2010; Mazzotti *et al.*, 2011), including the most recent ICE-6G modelling (Peltier *et al.,* 2015), which shows ice thickness over Keewatin that are 1.5 km thinner than ICE-5G (Vettoretti and Peltier, 2013). A suite of new radiocarbon dates that constrain the relative sea-level history of Arviat on the west coast of Hudson Bay are also consistent with a peak thickness of ~3.4 km at the LLGM (Simon *et al.*, 2014).

Thus, although estimates of the LIS have varied quite dramatically over the last ~100 years, and this aspect of the ice sheet's history remains poorly constrained (see Lambeck *et al.,* 2017), Table 1 indicates a convergence towards lower volumes in the more recent literature (cf. Abe-Ouchi *et al.,* 2015), which are consistent with the influence of deforming beds and a thinner, multi-domed configuration (Fisher *et al.,* 1985; Dyke and Prest, 1987; Clark *et al.,* 1996; Marshall *et al.,* 2000; Tarasov *et al.,* 2012; Lambeck *et al*., 2017). At its Late Wisconsinan maximum, the LIS (including the IIS and the Appalachian Ice Complex) most likely contained around 20 x 10^6 km^3, which is equivalent to ~50 m of global sea level (Clark *et al*., 1996) (Table 1). This is around 500-1000 m lower in elevation than the original CLIMAP reconstructions (Denton and Hughes, 1981), which contained 34.2 x 10^6 km^3 (85 m of global sea level), or around 75% more ice (Table 1).

Table 1. Chronological compilation of published estimates of the LIS extent, elevation and volume at its Local LGM (updated from Licciardi et al.*, 1998). Where indicated (*), note that some higher values are due to the inclusion of the Cordilleran Ice Sheet because separate values for the LIS were not quoted.*

Reference	**Extent (10^6 km^2)**	**Maximum elevation (km above present sea level)**	**Volume (x 10^6 km^3)**
Ramsay (1931)	15.75	2.9	45.45
Donn *et al.* (1962)	12.74	-	31.85 - 25.48
Andrews (1969)	11.82	-	26.0
Flint (1971)	13.39	-	29.46
Paterson (1972)[1]	11.6	2.7	26.5
Sugden (1977)[1]	-	3.5	37.0
Budd and Smith (1981)[1]	-	4-4.5*	<32*
Denton and Hughes (1981) (CLIMAP)[1]	-	3.8 (max. model) 3.5 (min. model)	34.2 (max. model) 30.5 (min. model)
Boulton *et al*. (1985)[1]	-	3-3.5 (hard bed model) >3 (soft bed model)	33-44 (hard bed model) (soft bed not reported)
Fisher *et al*. (1985)[1]	-	>3.2 (max. hard bed model) >3.2 (min. hard bed model) 2.8-3.2 (soft-bed model)	25.9 (max. hard bed model) 21.1 (min. hard bed model) 18.0 (soft-bed model)
Tushingham and Peltier (1991) (ICE-3G)[1]	-	>3	21.0
Peltier (1994) (ICE-4G)[1]	-	~3	19.0

Clark *et al*. (1996)	-	2-2.5	19.7
Marshall and Clark (1997a, b)[2]	~14	4.2	36.4
Licciardi *et al.* (1998)[1]	-	3.1 (min. model) 3.6 (max. model)	15.9 (min. model) 19.7 (max. model)
Tarasov and Peltier (1999)[2]	~13*	~3.8*	25*
Peltier (2004) (ICE-5G)	-	>4	-
Andrews (2006)	12	3-4	-
Tarasov *et al.* (2012)	-	-	28 (model nn9927)
Gregoire *et al.* (2012)	~16	>3	~35*
Lambeck *et al*. (2017)		≥3.5	

[1] Taken from Licciardi *et al.* (1998).

[2] Cited in Marshall *et al.* (2000).

3. Pattern and timing of deglaciation

It was not until the 1960s that researchers attempted to systematically reconstruct the pattern and timing of deglaciation at the scale of the entire ice sheet and produce maps of the ice margins at specific time-steps (isochrones). One of the first attempts to undertake this was by Bryson *et al.* (1969) who utilised existing radiocarbon dates (289 in total) and geological information to plot the ice sheet perimeter at 500 to 1000 yr intervals through time from about 13 ka (Fig. 6). Key conclusions from that pioneering study were that the northern limit of the ice sheet lay close to the Arctic mainland coast of Canada, now known to be incorrect (see Section 2.3), and that the most dramatic retreat took place along the western margin, creating an ice-free corridor from the Arctic Ocean to the Great Plains around 10.1 ka (9 ^{14}C ka). They also noted that the LIS "retained its identity as a distinct unit" (p. 1) until around 8.4 ka, which they termed the Cockburn Phase. At the time, this was thought to be the "only major glacial pulsation" (Bryson *et al*., 1969: p. 7) that had been recognized stratigraphically and geomorphologically (e.g. major moraine systems) over large areas of the eastern and central Canadian Arctic (e.g. Falconer *et al*., 1965). It was significant because it represented the final phase of the contiguous LIS, with remnant domes over Keewatin, Labrador, and Foxe Basin-Baffin Island. By around 8 ka, however, Bryson *et al.* (1969) argued that marine incursion into Hudson Bay heralded the rapid disintegration of the ice sheet, with only a small ice cap surviving to present day (Barnes Ice Cap).

Around the same time, Prest *et al.* (1968) produced their Glacial Map of Canada, which summarised a vast amount of information and literature on the glacial geomorphology of the ice sheet. This was soon followed by Prest's impressively detailed map of the 'Retreat of Wisconsin and Recent Ice in North America' (Prest, 1969), which resembled Bryson *et al*.'s (1969) synthesis in many respects, but was far more detailed and with isochrones portrayed at a much higher temporal resolution (Fig. 7). So impressive was this compilation that an anonymous author in the journal *Nature* wrote that it "deserves a place on every class room wall where earth sciences

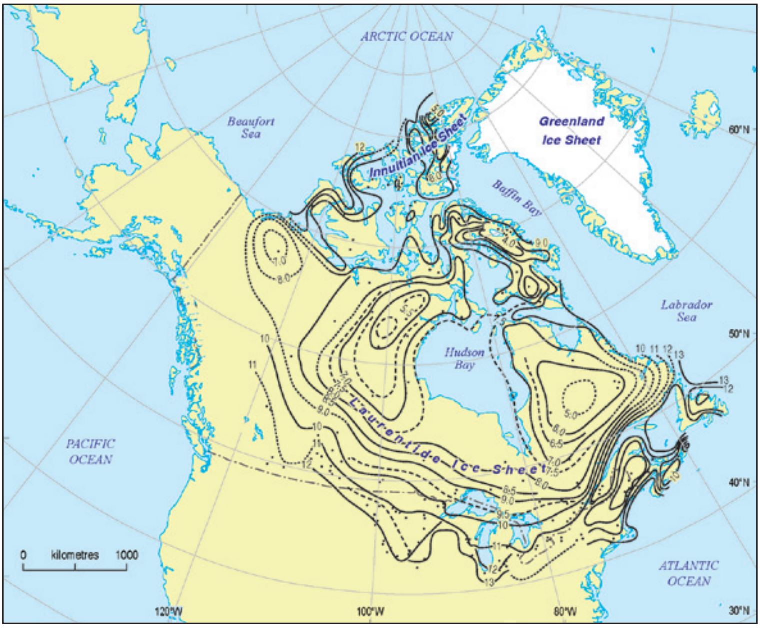

Figure 6. Radiocarbon-constrained isochrones of the retreat of the Laurentide Ice Sheet redrawn from Bryson et al. *(1969). Note that ages are in* ^{14}C *years BP. Dots indicate location of radiocarbon dates and dashed lines indicate larger uncertainty.*

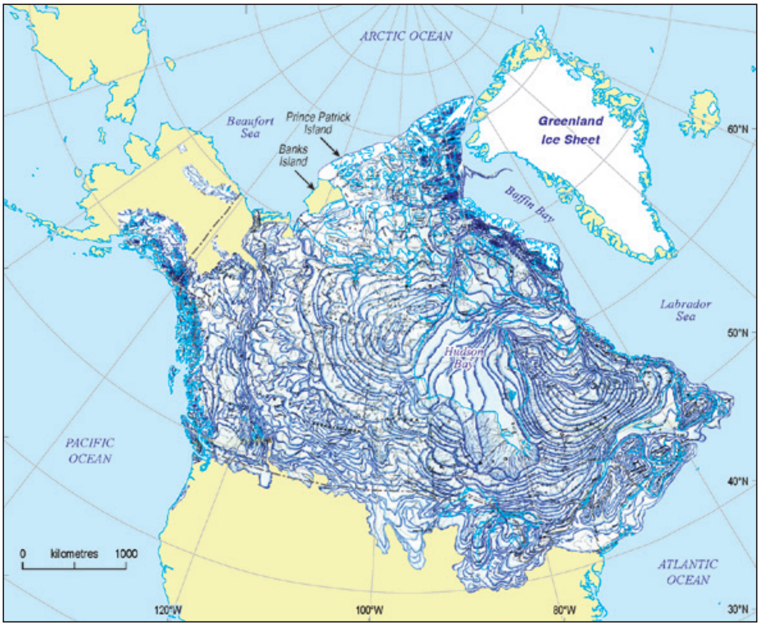

Figure 7. Reconstruction of the retreat pattern of the Laurentide and Cordilleran ice sheets redrawn from Prest (1969).

and American archaeology are taught…" (Anonymous, 1970, p. 224). Denton and Hughes' (1981) impressive compendium on 'The Last Great Ice Sheets' also presented a continental-scale synthesis of the LIS, but with a particular focus on its configuration during the Late Wisconsinan and at 4-5 key time-steps during deglaciation (see Chapter 2: Mayewski *et al*., 1981). The next detailed synthesis of the ice retreat pattern was by Boulton *et al.* (1985). A key conclusion from their reconstruction was the rapid retreat of the southern margins and a very slow retreat at the northern margins of the LIS, perhaps reflecting a strong N-S climate gradient during deglaciation. They also noted that overall ice margin recession must have paused and that the margin maybe re-advanced at various locations during overall deglaciation.

Building on a large number of reports and maps on the glacial geology of Canada, much of it undertaken with impressive detail by the Geological Survey of Canada, Dyke and Prest (1987) produced their influential reconstruction of the pattern and timing of the LIS retreat (see Section 2.3). This comprised a series of palaeogeographic reconstructions at 11 time-steps (4 simplified examples are shown in Fig. 8) which included information on the ice sheet outline, geometry and associated changes in proglacial lake drainage and relative sea-level oscillations; and an accompanying map showing much higher resolution isochrones (Map 1702A: Dyke and Prest, 1987). In part, the paper was motivated by several debates that had emerged since Prest's (1969) map (Fig. 7), namely: (i) the location (extent) of the maximum Late Wisconsinan limit (see Section 2.3), (ii) the surface geometry of the ice sheet (i.e. the location of major ice domes and divides and their evolution through time), and (iii) the synchronicity of ice marginal fluctuations in the north versus the south (Dyke and Prest, 1987).

In many respects, Dyke and Prest's (1987) reconstruction has remained the benchmark for the last three decades, with the only major changes being the revisions to a more extensive Late Wisconsinan maximum at the LGM (see Section 2.3) and a refined ice margin chronology that has benefited from improvements in radiocarbon dating (mainly the advent of AMS dating methods) and the 'retirement' of hundreds of earlier conventional radiocarbon dates (see Dyke, 2004). The updated ice margin chronology is described in Dyke (2004) and is available in digital format in Dyke *et al*. (2003), which includes 36 time steps, starting 21.8 ka (18 ^{14}C ka) and ending at 0.9 ka (1 ^{14}C ka). This new chronology is based on >4000 dates that are spread across the entire ice sheet bed and consist 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. bulk samples with probable blended ages) were excluded in the Dyke *et al.* (2003) and Dyke (2004) chronologies and marine-shell dates, a major component, were also adjusted for regionally variable marine-reservoir effects on the basis of a new set of radiocarbon ages. The net effect is that deglaciation is delayed in most places by 500-2000 years with respect to the Dyke and Prest (1987) reconstructions (cf. Dyke, 2004). However, the spatial pattern of ice recession resembles earlier reconstructions and the pattern of deglaciation is described in Dyke and Prest (1987) and Dyke (2004), which form the basis of the following discussion and to which the reader is referred for more detail (see also the compendium in Fulton (1989), which covers several regions in impressive detail). The following sections present a broad overview of the pattern and timing of deglaciation at the continental scale, with potential driving mechanisms of these broad patterns discussed in Section 4.

3.1. Local LGM to early Late Glacial: ~25-17.6 ka (18-14.5 ^{14}C ka)

Dyke and Prest's (1987) reconstruction at the LLGM included ice flow patterns that were informed by the distribution or glacial landforms (e.g. moraines, eskers and glacial lineations) on Prest *et al.*'s (1968) Glacial Map of Canada. Ice flowlines intersected the margin at right angles (unless they were in highly lobate areas with assumed divergent flow) and were followed back toward the centre of the ice sheet until features orientated in a different direction were encountered. A key feature of the LLGM ice flow pattern was the major ice stream in Hudson Strait that issued from a catchment area centred over Hudson Bay and with major domes over Quebec-Labrador to the south-east, Keewatin to the west and the Foxe-Baffin dome to the north (see Fig. 2). A major 'Trans-Laurentide Ice Divide' extended from near Victoria Island in the Canadian Arctic Archipelago south towards the Keewatin Dome and then westwards to connect with the Labrador dome, with secondary ice divides emanating from regional ice dispersal centres over the Queen Elizabeth Islands, Baffin Island, Newfoundland, and the Appalachians (Fig. 2). Dyke and Prest (1987) also depicted several major ice streams in regions where flow-lines exhibited strong convergence and they noted that some of these coincided with distinctive erratic dispersal plumes (e.g. Dyke *et al*., 1982; Dyke, 1984). Other ice streams were invoked at the southern margin of the ice sheet, associated with the major ice lobes that were known to have possessed very low ice surface gradients (cf. Mathews, 1974) as a result of deformable bed conditions beneath the ice sheet (e.g. Boulton *et al*., 1985; Fisher *et al*., 1985; see Section 2.4).

It is widely recognised that initial deglaciation from the LLGM configuration outlined above was generally slow during the first part of the period known as the Late Glacial (Dyke and Prest, 1987; Dyke *et al.,* 2002) and that some margins may even have been advancing to their local maximum, e.g. at the far north-west (Kennedy *et al.,* 2010; Lacelle *et al*., 2013; Murton *et al.,* 2007, 2015). Dyke *et al.* (2002) pointed out that there is little evidence of regional recession prior to 17 ka (14 ^{14}C ka), with the exception of the Atlantic Provinces, the Lake Michigan basin, the Mackenzie Lobe in the far north-west (cf. Harrington, 1989), and possibly in Hudson Strait, following Heinrich Event 2 (H2) (Andrews *et al.,* 1998) (Fig. 8a). The Atlantic provinces underwent the most dramatic retreat during this period, where deglaciation was associated with a margin in deep water (Mosher *et al.,* 1989; Piper *et al*., 1990; King, 1996; Scnitker *et al*., 2001; Shaw *et al.,* 2006), and was perhaps triggered by eustatic sea level rise (Dyke, 2004).

Dyke (2004) also noted that the lobes of the southern margin were probably oscillating during slow net recession, but had retreated more substantially by the culmination of the Erie Interstadial, which is poorly dated, but which Dyke (2004) placed at around 18.8 ka (15.5 ^{14}C ka, see Barnett, 1992). It has then been suggested that several of the Great Lakes ice lobes underwent a major readvance (several hundred kilometres) during the succeeding Port Bruce Stadial (e.g. Erie/Huron lobe, Des Moines lobe: Clayton and

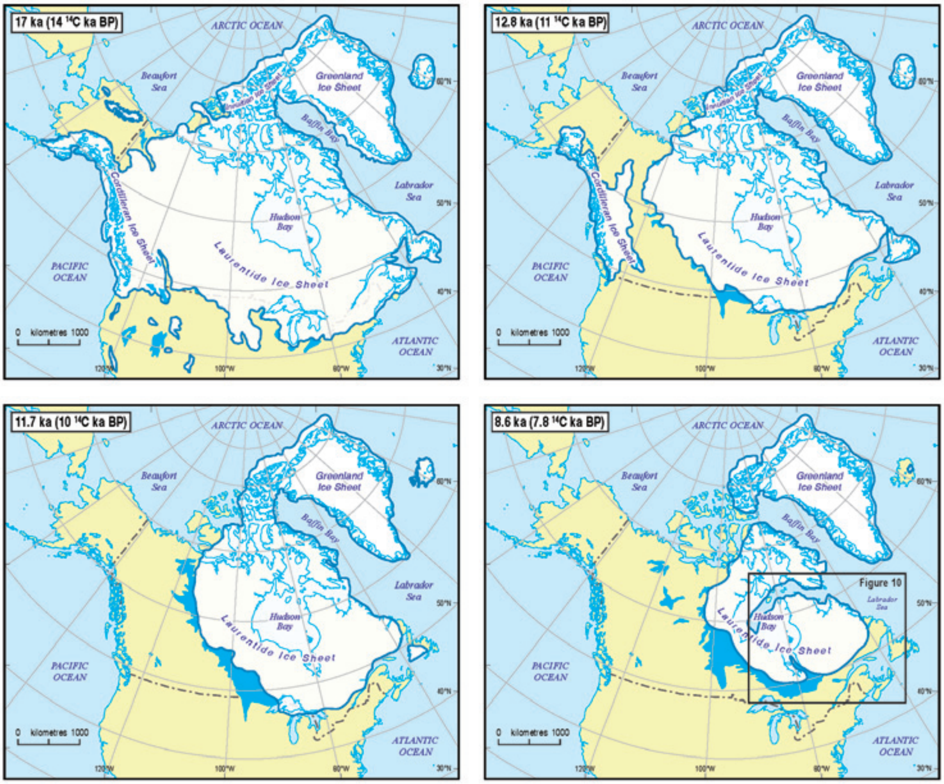

Figure 8. Examples of the reconstruction of the North American ice sheet during deglaciation redrawn from Dyke (2004): (**a**) 17 ka, (**b**) 12.8 ka, (**c**) 11.7 ka and (**d**) 8.6 ka.

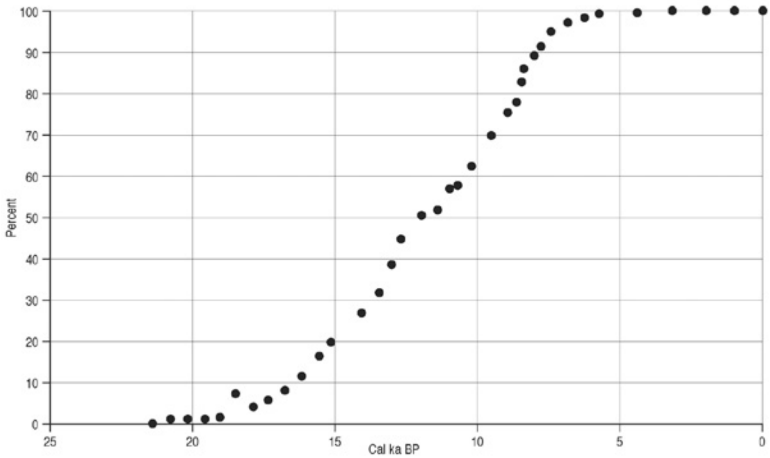

Figure 9. Percentage area deglaciated in North America compared to its Local Last Glacial Maximum, redrawn from Dyke (2004).

Moran, 1982; Clayton *et al.*, 1985) and produced the only net increase in ice extent during overall deglaciation (Dyke, 2004) (see Fig. 9). It is difficult to date correlative advances elsewhere in the ice sheet (see discussion in Dyke, 2004), but it is thought that they may also have taken place in Hudson Strait (Andrews *et al.*, 2001) and perhaps in the Atlantic provinces (Miller *et al.*, 2001; Shaw, 2003). Elsewhere, Dyke (2004) noted that an AMS date on wood from basal lake sediments in south-western Alberta (Beierle and Smith, 1998) indicates that initial decoupling of Laurentide and Cordilleran ice had begun by around 19 ka (15.7 ^{14}C ka BP).

The fact that large-scale retreat of the LIS did not begin until around 16.8 ka (14 ^{14}C ka: Dyke *et al.*, 2002) is noteworthy because far-field sea-level records indicate that global sea levels had begun to increase a few thousand years prior to that time (Clark *et al.*, 2009). Thus, it has been argued that if the LIS was contributing to sea-level rise in the early Late Glacial, then it must have been largely through thinning, rather than areal recession, and that this thinning and drawdown may have been associated with a transition from a thick, cold-based LGM ice sheet to thinner, warm-based ice sheet during early deglaciation (Marshall *et al.,* 2000; Marshall and Clark*,* 2002; Robel and Tziperman, 2016). This transition may also have been manifest in a major internal flow re-organisation that may have been correlative with Heinrich event 1 (H1) around 17.7 ka (14.5 ^{14}C ka) (see Veillette *et al.*, 1999). Indeed, Dyke *et al.* (2002) noted that if this reorganisation occurred, it is likely that the drawdown of central ice surface would have promoted subsequent deglaciation by increasing the ELA. However, Dyke and Prest (1987) noted that changes in the ice marginal configuration between the LLGM and 17 ka (14 ^{14}C ka) were insufficient to effect any permanent or substantial changes in the position of the primary ice domes and divides.

3.2. Late Glacial Interstadial: ~17.6-12.8 ka (14.5-11 ^{14}C ka)

This period includes the Bølling-Allerød warm interval, punctuated by the brief Older Dryas cold event (Lowe *et al.*, 1994; Dyke, 2004). As noted by Dyke (2004), this period was associated with a clear pattern of net retreat of the LIS, particularly along the southern and western margins. It has also been pinpointed as a time of marked volume loss, particularly between ~15 and 14.5 ka (Lambeck *et al.,* 2017). Indeed, the LIS had likely become fully separated from the Cordilleran Ice Sheet (CIS) by the end of the Late Glacial Interstadial (Bølling-Allerød), whilst the northern and eastern margins of the ice sheet underwent only minimal recession (Fig. 8b) (Dyke and Prest, 1987; Dyke, 2004). Despite this asymmetric retreat, the overall configuration of the ice sheet geometry is thought to have changed little, with many of the major ice domes and divides remaining stable (Dyke and Prest, 1987). However, the marked retreat of the western margin of the ice sheet is likely to have driven an eastward migration of the main north-south M'Clintock Ice Divide and there were some marked changes in the regional ice flow directions over the interior plains (Dyke and Prest, 1987; Ó Cofaigh *et al.*, 2009; Ross *et al.*, 2009). It is also noticeable that this broad time interval was associated with the development of numerous glacial lakes along the western and southern margin of the ice sheet (Dyke and Prest, 1987; Fig. 8b).

The timing of the opening of the ice-free corridor between the LIS and CIS has been the subject of much debate, largely due to its importance as a potential route for the peopling of North America (e.g. Dixon, 1999; Goebel *et al.,* 2008; Eriksson *et al.,* 2012; Pedersen *et al.,* 2016), but also in relation to the routing of meltwater from Glacial Lake Agassiz (Smith and Fisher, 1993; Fisher and Smith, 1994; Fisher *et al*., 2002; Tarasov and Peltier, 2005; Murton *et al*., 2010; Fisher and Lowell, 2012; Teller, 2013; see Section 3.3). The rapid collapse of the saddle between the LIS and CIS has also been hypothesised as a potential source of meltwater pulse 1A (Gregoire *et al*., 2012, 2015a), although the precise contributions from North America are subject to ongoing debate (Clark *et al.,* 2002; Carlson and Clark, 2012; Deschamps *et al.*, 2012).

Dyke and Prest's (1987) reconstruction showed the ice free corridor opening up ~15.6 ka (13 ^{14}C ka) and with the western margin of the LIS some 200-600 km east of the Cordilleran Mountains by ~12.9 ka (11 ^{14}C ka) (Fig. 8b). This conflicts with some arguments that were later put forward (e.g. Dixon, 1999) that suggested that the ice sheets must have been fully-coalescent until quite late in the interstadial (~12.9 ka). Recently, Lambeck *et al.* (2017) have also argued that any ice-free corridor is unlikely to have existed prior to 13 ka, using a new model of glacial rebound based on relative sea level data and the tilting of glacial lake shorelines. A similar conclusion was reached by Pedersen *et al.* (2016) who obtained radiocarbon dates, pollen, macrofossils and metagenomic DNA from lake sediment cores along the central portion of the corridor and found that it was not likely to be viable as a migration route before 12.6 ka. Unfortunately, this 'unzipping' of the two ice sheet remains very poorly dated. Dyke (2004) argued that the southern margin probably began opening around 18.2 ka (~15 ^{14}C ka) and that it is possible that it may have opened completely by ~16.3 ka (13.5 ^{14}C ka), based on radiocarbon dating of basal sediment in glacial Lake Peace (Catto *et al.,* 1996), situated mid-way along the corridor from south to north. However, Dyke (2004) concluded that this scenario is unlikely and that the northern part of the corridor probably deglaciated later and around 14.7-13.9 ka (12.5-12 ^{14}C ka). A new analysis of available dates has attempted to constrain the minimum timing of the opening of the ice-fee corridor and suggests that it must have been completed by 11 ka (Gowan, 2013).

Elsewhere, the Late Glacial Interstadial is characterised by oscillations of the ice lobes at the southern margin of the ice sheet, superimposed on net recession, especially in the vicinity of the Great Lakes (Dyke, 2004). Ice margin retreat has been tracked at a remarkably high resolution (<100 years) in numerous glacial lake sequences that were deposited as various basins became isolated (e.g. Karrow and Calkin, 1985). Many of these sequences record the initiation of retreat at ~17 ka (14 ^{14}C ka), but it was clearly punctuated by readvances of the ice margin that blocked drainage routes of glacial lakes. For example, the large readvance of the Lake Michigan Lobe around 13.6 to 12.3 ka (11.8 to 11.5 ^{14}C ka) is thought to have diverted water away from a westward route into the Labrador Sea and back towards the Mississippi drainage basin and the Gulf of Mexico (Dyke, 2004).

3.3. The Younger Dryas (YD): 12.9-11.7 ka (11-10 ^{14}C ka)

Ice recession during the YD was generally slow, particularly at the northern and eastern margins of the ice sheet, where deglaciation mostly occurred after this period

(Fig. 8) (Andrews, 1973; Dyke, 2004). However, the clear asymmetry of the retreat pattern (cf. Andrews, 1973) continued to drive the main M'Clintock Ice Divide (running north from the Keewatin dome: Fig. 2) and the ancestral Keewatin Ice Divide eastward (Dyke and Prest, 1987). Indeed, Dyke and Prest (1987) also noted that the period beginning around 11.5 ka (10 ^{14}C ka) marked the beginning of the demise of the main Trans Laurentide Ice Divide and increased the autonomy of the regional ice dispersal centres.

Although it was barely mentioned in comprehensive treatments of North American deglaciation prior to the late 1980s (Fulton, 1989), the Younger Dryas (YD) cold event is now known (cf. Dyke, 2004) to have been characterised by a period of moraine construction and, in several places, major readvances of the ice margin and ice marginal lobes (e.g. Dyke and Savelle, 2000). Such readvances have been well-documented along several parts of the ice sheet margin and include the large Gold Cove readvance from Labrador across Hudson Strait (Miller and Kaufman, 1990) and major readvances at the north-western margin of the ice sheet in the Canadian Arctic Archipelago (Dyke and Savelle, 2000). In many places, such readvances were associated with major ice lobes/ice streams, such as the Cumberland Sound Ice Stream on Baffin Island (Jennings *et al.,* 1996; Andrews *et al*., 1998) and the M'Clintock Channel ice stream on Victoria Island (Hodgson, 1994; Stokes *et al.,* 2009). As Dyke (2004) noted, although some moraines are clearly distinguishable as of YD age, others are likely to be correlative, but have not been precisely dated.

Any discussion of the LIS retreat during the YD warrants a mention of the drainage routes of glacial Lake Agassiz, which has been implicated as causative mechanism of this abrupt climatic reversal (Broecker *et al*., 1989). Initially, glacial Lake Agassiz drained to the south and into the Gulf of Mexico via the Mississippi River. The traditional model (cf. Dyke, 2004; Carlson and Clark, 2012) is that retreat of the Lake Superior Lobe after ~12.9 ka (Fig. 8b) allowed glacial Lake Agassiz to drain rapidly towards the east via the St Lawrence River and into the North Atlantic Ocean (Broecker *et al.,* 1989; Dyke, 2004). This may have released up to 9,500 km^3 of water, which is thought to have been capable of disrupting the North Atlantic's Meridional Overturning Circulation (AMOC) and instigating the YD cooling (Broecker *et al.,* 1989; Dyke, 2004). Following the initial outburst, the eastward drainage is thought to have continued until it was blocked by ice during the Marqueette readvance, which culminated around 11.5 ka (10 ^{14}C ka) (Dyke, 2004). It is then thought that the drainage route may have switched northwards via the Clearwater spillway and towards the Arctic Ocean, via glacial Lake McConnell (Smith and Fisher, 1993; Fisher and Smith, 1994; Fisher *et al.,* 2002). The drainage is then thought to have switched back to its original southward route until the recession of ice north of Lake Superior once again opened the westward route (Teller and Thorleifson, 1983). More recently, however, an alternative model has been suggested which indicates that Lake Agassiz may have drained to the northwest and into the Arctic Ocean much earlier than originally thought, and at the onset of the Younger Dryas (Murton *et al*., 2010). This is based on the dating of sands associated with the Mackenzie delta and upstream gravels and erosional channels (Murton *et al*., 2010), but numerical modelling has also indicated increased runoff via this outlet around this time, even in the absence of any lake drainage (Tarasov and Peltier, 2005). Furthermore, high-resolution ocean modelling indicates that freshwater input to the Arctic Ocean is much more effective at perturbing the AMOC compared to an input from the eastern drainage route (Condron and Winsor, 2012). Field evidence for the opening of the Clearwater spillway at the onset of the YD is, however, far from equivocal (Fisher and Lowell, 2012) and the debate continues.

3.4. Final deglaciation: 11.5-6 ka (10-5.2 ^{14}C ka)

Final deglaciation of the LIS occurred during the early to middle Holocene (11.5-6.0 ka) in response to increased summer insolation and increasing levels of carbon dioxide (CO_2) (Carlson *et al*., 2007; 2008; Marcott *et al.,* 2013). This warming led to the disappearance of most Northern Hemisphere ice sheets, but Ullman *et al.* (2016) noted that, despite this strong radiative and temperature forcing, global mean sea level (GMSL) was still around 60 m below present at the start of the Holocene (Lambeck *et al.,* 2014), indicating a lag (of as much as 4 ka) between deglaciation of the LIS and peak insolation and CO_2 forcings (see also Section 4.1).

Retreat of the LIS was most dramatic along the northern and western margins of the ice sheet. Recession of the northern margin of the ice sheet accelerated dramatically soon after 11.5 ka (10 ^{14}C ka) and it is thought that the LIS and IIS had separated by 10.1 ka (9 ^{14}C ka), but that the IIS remained confluent with the Greenland Ice Sheet until 8.6 ka (7.8 ^{14}C ka) (Fig. 8c, d) (England, 1999; Dyke, 2004). Dyke (2004) suggests that the IIS had fragmented by 9.5 ka (8.5 ^{14}C ka) and had retreated close to modern ice margins by 8.6 ka (7.88 ^{14}C ka). Dyke (2004) also noted that the Keewatin Sector of the ice sheet had cleared the Canadian Arctic Archipelago by 8.6 ka (7.8 ^{14}C ka) (Dyke, 2004). Retreat across the mainland was also rapid, but the ice sheet constructed a series of major moraine systems (e.g. the MacAlpine and Chantrey moraines) that Dyke (2004) assigned to around 9.1 to 8.6 (8.2 to 7.8 ^{14}C ka). These moraines lie well inland of the marine limit and likely represent some form of readvance of the ice margin, perhaps associated with ice streaming (e.g. Stokes and Clark, 2003), rather than what Dyke (2004) referred to as 'end-of-calving' stabilisations in the Labrador Sector (see below).

Recession of the southern margin of the ice sheet was also rapid and has been reconstructed with impressive detail from tracing glacial lake shorelines to end moraines (e.g. Barnett, 1992). As the ice margin retreated into an isostatically-depressed interior, numerous lakes were decanted and their drainage re-routed, including a major eastward discharge of Lake Agassiz at around 10.1 ka (9 ^{14}C ka) (Dyke, 2004). These lakes may also have facilitated localised readvances of the ice margin (often termed 'surges' or ice streaming), such as into glacial Lake Ojibway (Thorleifson *et al.,* 1993) and help explain the contrasting dynamics of neighbouring lobes during overall ice margin retreat (Cutler *et al*., 2011).

The final evolution of lakes Agassiz and Ojibway after 8.9 ka (8 ^{14}C ka) is more speculative, but their northwards drainage into the Tyrrell Sea (ancestral Hudson Bay) is evidenced by glacial bedforms, subglacial drainage channels and numerous iceberg scour-marks (Josenhans and Zevenhuizen, 1990). Josenhans and Zevenhuizen (1990) argued that a large calving bay opened up in western Hudson Bay and that glacial lakes Agassiz-Ojibway initially drained into that region, rather than eastern Hudson Bay. The final catastrophic drainage of the Agassiz-Ojibway and the full incursion of the Tyrrell Sea has been dated to around 8.4 to 8.2 ka (Dyke, 2004), which is correlative with the '8.2 ka cold event' seen in Greenland ice cores (Alley *et al*., 1997; Barber *et al*., 1999; Rasmussen *et al.,* 2006). Indeed, Barber *et al.* (1999) have argued that the sudden release of freshwater that accompanied this event is likely to have disrupted the AMOC and lead to the abrupt but short-lived cooling seen in numerous circum-North Atlantic records. Elsewhere, Dyke (2004) noted that the western margin of the Québec-Labrador ice cap had stabilised at the 800-km long Sakami moraine (Fig. 10), most likely as a result of the sudden and large reduction of water depth that was accompanied by drainage of Lake Ojibway (Hardy, 1982).

In contrast to the rapid retreat seen elsewhere, Dyke (2004) suggests that the Baffin Sector was still close to its maximum configuration at 11.5 ka (10 ^{14}C ka) and that retreat of many of its outlet glaciers proceeded slowly between 11.5 ka (10 ^{14}C ka) and 9.5 ka (8.5 ^{14}C ka). A series of extensive moraine systems were also constructed around much of the Baffin Sector between 9.5 ka (8.5 ^{14}C ka) and 7.8 ka (7 ^{14}C ka), which Dyke (2004) suggests may reflect a mass balance that fluctuated from positive to slightly negative. That said, it is clear that some major outlet glaciers retreated rapidly during this period, especially through deep bathymetric troughs (Briner *et al*., 2009), and Dyke and Prest (1987) portray the first major recession of the terminus of the Hudson Strait Ice Stream at 10.1 ka (9 ^{14}C ka).

Dyke (2004) argued that the final break-up of the Foxe-Baffin Sector likely involved the northward progression of a calving bay from Hudson Bay between 7.8 and 6.9 ka (Fig. 8d) (7 to 6 ^{14}C ka), leaving residual ice caps on Baffin Island, Southampton Island, and Melville Peninsula that remain the last major remnants of the LIS. Deglaciation of the remnant Keewatin and Foxe Domes (Dyke, 2004; Ross *et al.,* 2012; Simon *et al*., 2014) left a remnant Labrador Dome that has been estimated to contain a sea level equivalent of 3.6 ± 0.4 m at ~8.2 ka (Ullman *et al.,* 2016) (Fig. 10). Recently, Ullman *et al*. (2016) constrained the final retreat of this ice mass using ^{10}Be surface exposure dating and demonstrated that the ice margin may have been highly sensitive to several abrupt climate events. Superimposed on overall retreat, they demonstrated that the ice sheet deposited a series of moraine systems at ~10.3 ka (Paradise Moraine), 9.3 ka (North Shore Moraine) and 8.2 ka (Sakami Moraine) (Fig. 10), which coincided with North Atlantic cold events (Bond *et al*. 1997; Rasmussen *et al*., 2006), and which may have helped to stabilise the ice sheet. Following the widely-documented 8.2 ka event (see also Alley *et al.,* 1997; Barber *et al*., 1999) and the opening of Hudson Strait, they suggest that Hudson Bay became seasonally ice-free and that the majority of the ice sheet melted abruptly and within a few centuries, with deglaciation of the LIS completed by 6.7 ± 0.4 ka (Fig. 10) (cf. Carlson *et al*., 2007; 2008). Indeed, using a Regional Climate Model, Ullman *et al.* (2016) argued that the loss of ice over Hudson Bay would have been important in driving negative mass balances in the surrounding ice masses, largely due to the increased thermal capacity and reduced albedo of seasonally open water.

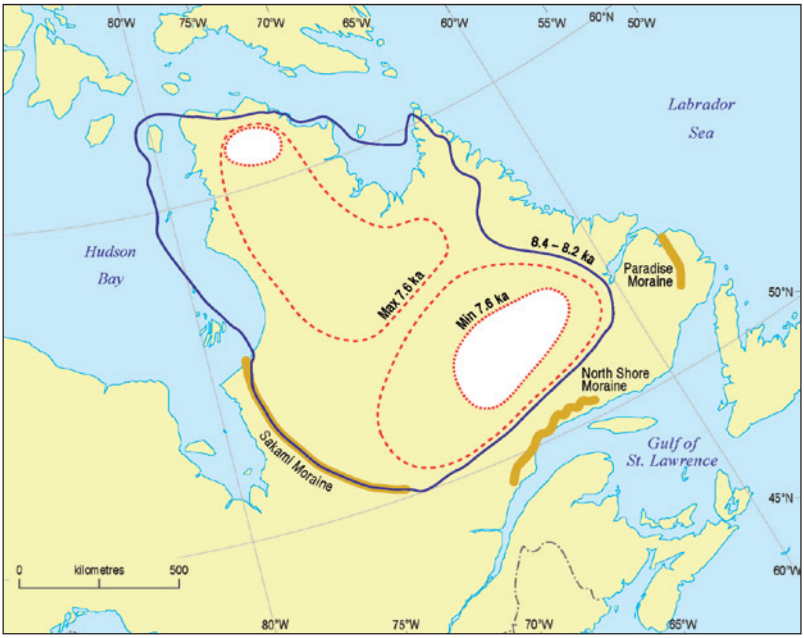

Figure 10. Reconstructed limits of the final deglaciation of the LIS in Quebec/Labrador, redrawn from Ullman et al. *(2016). Blue lines show Dyke's (2004) reconstruction at 8.4-8.2 ka and red lines show maximum (dashed) and minimum (dotted) 7.6 ka ice areas from Ullman* et al. *(2016) in relation to the Sakami, North Shore, and Paradise moraines (see Occhietti* et al.*, 2011).*

3.5. Summary

The pattern and timing of the LIS deglaciation is now reasonably well-known and is characterised by a clear asymmetry whereby the western and southern margins retreated back towards major dispersal centres over Foxe Basin and Labrador. In terms of the timing, deglaciation is characterised by a period of very slow recession prior to ~17 ka, when it lost <10% of its area, followed by a near-linear retreat until ~7.8 ka, when only 10% of the area remained more glaciated than present (Fig. 8, 9) (Dyke, 2004). Although there were numerous local-scale ice marginal fluctuations marked by rapid advance or retreat and internal flow reorganisation, the two most important events that interrupted the overall linear recession were: (i) the reduced rate of recession during the YD (including several well-documented readvances), and (ii) the final increased rate of recession during marine incursion into Hudson Bay (Dyke, 2004).

4. Climatic forcing and mechanisms of LIS deglaciation

As noted in the Introduction, palaeo-ice sheets represent a valuable analogue for understanding the rates and mechanisms of ice sheet deglaciation. Mass loss from ice sheets is complex, but can be broadly partitioned (cf. van den Broeke *et al.*, 2009) between melting (mostly at the surface, but also under the ice sheet and where it meets the ocean), and a 'dynamic' component whereby rapidly-flowing outlet glaciers transfer ice from the interior to the oceans. In the Greenland Ice Sheet (GrIS), these processes are thought to be contributing approximately equally to its recent negative mass balance (van den Broeke *et al.,* 2009). In Antarctica, however, there is much less surface melt and dynamic changes, namely the acceleration, thinning and retreat of outlet glaciers (Pritchard *et al.*, 2009), are a more important influence on its mass balance, which is negative in West Antarctica but negligible or even slightly positive in East Antarctica (Shepherd *et al.,* 2012). The extent to which surface mass balance and 'dynamic' discharge (ice streaming) influenced the deglaciation of the Laurentide Ice Sheet will now be discussed.

4.1. Surface mass balance during deglaciation of the LIS

In their discussion of the timing of the gLGM, Clark *et al.* (2009) argued that the primary mechanism for triggering the onset of deglaciation in the Northern Hemisphere between 20 and 19 ka was increased insolation from orbital forcing. The efficacy of this external forcing is via increased surface melt in marginal areas, particularly at the southern margin of the LIS. Clark *et al.* (2009) also noted that once deglaciation had been initiated, it is likely that several feedback mechanisms would have amplified the initial response (e.g. involving albedo, CO_2 and oceanic feedbacks), including a delayed crustal rebound, which keeps the ice sheet elevation relatively low and increases ablation (Abe-Ouchi *et al.,* 2013). Recent modelling by Gregoire *et al.* (2015b) has attempted to partition the influence of increasing greenhouse gases (GHGs, e.g. CO_2) and orbital forcing on North American deglaciation. They found that orbital forcing explains around 50% of the reduction in ice volume during deglaciation, while GHGs explain around 30%, but that the impact of GHGs lags orbital forcing. Orbital forcing begins around 23 ka and starts to impact the ice sheet from around 19 ka, but there is a delay of 3 ka before CO_2 forcing has a noticeable influence from around 16 ka.

Recently, Ullman *et al.* (2015a) pinpointed the initial retreat of the southern margin of the ice sheet in Wisconsin using a suite of ^{10}Be surface exposure ages from boulder surfaces in terminal moraines. These ages dated the initial retreat of the ice margin from the LLGM moraines to 23.0 ± 0.6 ka, which they noted was synchronous with several other locations along the southern margin and coincided with the initial increase in summer insolation around 24-23 ka. They also pointed out that an acceleration in retreat after around 20.5 ka was likely driven by an acceleration in boreal summer insolation and that this occurred before any increase in atmospheric CO_2, supporting an orbital forcing as the trigger for initial deglaciation (Clark *et al.*, 2009; Gregoire *et al.*, 2015b). This response of the ice sheet to atmospheric forcing also implies a higher sensitivity of land-terminating margins to small changes in climate forcing than had hitherto been recognised, although it should be noted that overall recession of the ice sheet was minimal (see Section 3.1). It is also interesting that whilst the southern margin was beginning to retreat, there is strong evidence that the margin in the far north-west was still advancing and likely attained its maximum position after 18.5 ka (e.g. Murton *et al.*, 2007; Kennedy *et al.*, 2010; Lacelle *et al.*, 2013). It is not clear why the northwest margin advanced to its maximum position a few millennia after the global LGM (*sensu* Clark *et al*., 2009), but Lacelle *et al*. (2013) suggested that sea-level rise and the opening of the Arctic Ocean along the Beaufort Sea coastline may have provided a local source of moisture and increased precipitation that enabled an advance of the LIS in this region. They also pointed out that the abundance of deformable sediments in the region may also have facilitated a rapid advance of the Mackenzie Lobe (ice stream) (cf. Beget, 1987).

Although increased atmospheric warming is thought to have triggered the initial retreat of the LIS, surface energy balance modelling suggests that the ice sheet's overall net surface mass balance remained positive for much of the early part of deglaciation (Ullman *et al*., 2015b). Ullman *et al*. (2015b) used a surface energy balance model forced by climate data from simulations with a fully coupled atmosphere-ocean General Climate Model (GCM) for key time slices during the last deglaciation (24, 21, 19, 16.5, 15.5, 14, 13, 11.5 and 9 ka), see Fig. 11. They found that the net surface mass balance was positive until after 11.5 ka, which implied that mass loss was primarily driven by dynamic discharge via calving at marine-terminating ice streams (see Section 4.2). Only when summer temperatures increased by 6-7 °C (relative to the gLGM) did the ice sheet's surface mass balance become increasingly negative in the early Holocene. This occurred between 11.5 and 9 ka and was accompanied by an expansion of the ablation area that was previously restricted to the low-gradient lobe of the southern margin, but which expanded to most of the southern and western marginal areas by 9 ka (Fig. 11). Ullman *et al.* (2015b) noted that this time period also saw the LIS lose most of its marine margin and would have coincided with a large reduction in dynamic discharge via calving losses (cf. Stokes *et al.,* 2016).

It is worth noting, however, that the LIS had only lost around 40% of area in >10,000 years of deglaciation from the LLGM to around 9 ka, despite increasing boreal insolation and a ~80 ppm increase in CO_2 (Ullman *et al.,* 2015b). Thus, Ullman *et al.* (2015b) noted that the transition to a negative surface mass balance that occurred between ~11.5 and 9 ka, and the very rapid retreat of the ice sheet after ~9 ka (two to five times faster than before ~11.5 ka), suggests that some kind of instability threshold was crossed and that the final deglaciation of 60% of the ice sheet's area was driven by surface melt, rather than dynamic discharge. This is supported by numerous studies that have shown that the rapid decay of the LIS in the early Holocene was driven by enhanced boreal summer insolation and increased ablation (Carlson *et al*., 2007; 2008; 2009a). For example, Carlson *et al.* (2009a) used a surface energy balance model, driven by atmosphere-ocean general circulation model at 9 ka, and calculated a net surface mass balance of -0.67 ± 0.13 m a^{-1}. Given volume estimates of the LIS at this time, it indicates that surface ablation accounted for 74 ± 22% of mass loss at that time, with the remainder attributable to dynamic calving (Carlson *et al*., 2009a).

The evolving pattern of the surface mass balance of the ice sheet (e.g. Fig. 11) partly explains the clear asymmetry of retreat whereby the southern and western margins retreated much more rapidly than those in the north and especially the east. Following the initial retreat of the margins in the south, and the unzipping of the LIS and CIS in the east, surface-albedo feedbacks are likely to have helped drive further retreat (e.g. the emergence of darker land-surfaces). The formation of glacial lakes along the southern and western margins may also have enhanced localised calving and ice sheet draw-down

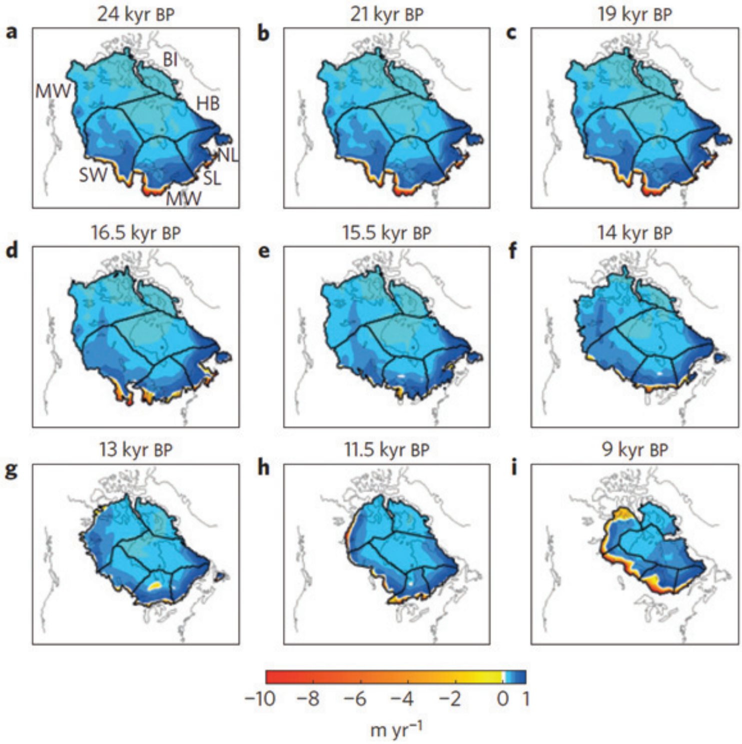

Figure 11. Modelled surface mass balance of the Laurentide Ice Sheet at various time-slices, with black lines demarcating various sectors of the ice sheet: St Lawrence (SL), Midwestern (MW), Southwestern (SW), Northwestern (NW), Hudson Bay (HB), Banks Island (BI) and Newfoundland (NL). Reprinted by permission from Macmillan Publishers Ltd: Nature Geoscience (Ullman et al., *2015b), copyright (2015).*

(e.g. Andrews, 1973; Cutler *et al.,* 2001; Stokes and Clark, 2004). Indeed, Andrews (1973) was one of the first to point out that the retreat of the LIS could not be explained solely by surface mass balance forcing and that an important process during deglaciation was mass loss associated with calving in both marine and lacustrine settings. However, Dyke and Prest (1987) noted that whilst calving was undoubtedly an important means of ice sheet ablation, its role should not be overemphasized. They pointed out that glacial lakes on the Prairies were small and that, until the formation of Lake Agassiz, calving cannot account for deglaciation for most of that region prior to its development. They also noted that the southern and eastern margins of the Labrador Sector had some of the longest marine margins, but that these margins retreated more slowly than the contemporaneous western margin of the Keewatin Sector, even though it occurred largely on dry land. It is also noteworthy that the marine-based part of the LIS over Hudson Bay and Foxe Basin were among the last to deglaciate, despite their susceptibility to calving and the fact that their central areas were isostatically-depressed hundreds of metres below sea level (Dyke and Prest, 1987). Indeed, even using an extreme ‘calving instability’, Marshall *et al*. (2000) were unable to evacuate a significant volume of ice from Hudson Strait in their modelling experiments, although more recent modelling (Bassis *et al.,* 2017) suggests that this ice stream may have been particularly vulnerable to calving triggered by subsurface ocean warming (see Section 4.2).

A further explanation for the asymmetric retreat may also relate to the release and routing of meltwater during deglaciation and its subsequent impact on ocean circulation (Carlson *et al*., 2009b; Hoffman *et al*., 2012; Jennings *et al.,* 2015; Gregoire *et al*., 2015b). Climate modelling indicates that meltwater discharge events routed into the Labrador Sea (e.g. from meltwater runoff and glacial lakes) could cause a cooling of up to 1.5°C over the Labrador Dome (Morrill *et al.,* 2014), but with minimal cooling (<0.5°C) along the western margin of the ice sheet, west of Hudson Bay. This negative feedback mechanism has thus been invoked to explain the relatively stability of the Labrador Dome, whilst the western margin of the ice sheet continued to retreat (Ullman *et al.,* 2016).

4.2. The role of ice streaming during deglaciation of the LIS

Ice streams are the key drainage routes of an ice sheet (Bamber *et al*., 2000) and are known to exert a considerable influence on ice sheet configuration and mass balance (Nick *et al*., 2013; Ritz *et al*., 2015). It has been recognised for some time, therefore, that accurate reconstructions of the LIS require a detailed knowledge of the location of ice streams. The first use of the term 'ice stream' in relation to the LIS was by Bell (1895: p. 352-353), who inferred the presence of a "great ice stream" passing through Hudson Strait (see Brookes, 2007). However, it was not until 1981 that Denton and Hughes (1981) attempted to incorporate ice streams into a reconstruction of the entire LIS. They were the first to portray an extensive network of ice streams in the Northern Hemisphere ice sheets (Fig. 12a), which was clearly influenced by Hughes' knowledge of West Antarctic ice streams (e.g. Hughes, 1977). It is not clear what evidence was used to locate the ice streams, but several major ice streams were depicted in large topographic troughs, such as Hudson Strait, and others were depicted on low relief terrains where they appeared as a regularly-spaced network, perhaps hinting at some notion of spatial organisation. Dyke and Prest (1987) also incorporated ice stream flow-lines in their reconstruction of the LIS, some of which had earlier been recognised from erratic dispersal trains (Dyke *et al.,* 1982; Dyke, 1984). Around that time, Dyke and Morris (1988) published a classic paper that would be one of the first to describe, in detail, the geomorphological 'footprint' of an ice stream on Prince of Wales Island in the Canadian Arctic Archipelago. They reported evidence for a convergent flow pattern of highly elongate drumlins associated with an erratic dispersal train with abrupt lateral margins.

Many of the lobes of the southern margin were also attracting attention as possible zones of ice streaming (Dredge and Cowan, 1989; Alley, 1991), largely because of their low surface slopes (Mathews, 1974) and earlier suppositions about the possibility of rapid flow in the form of surging (Wright, 1973; Clayton *et al.,* 1985). Hicock and co-workers reported dispersal trains and tills associated with drumlins around the Great Lakes region that were interpreted to reflect ice streaming (Hicock, 1988, 1992; Hicock and Dreimanis, 1992). Later work by Patterson (1997, 1998) recognised entire landform assemblages that were

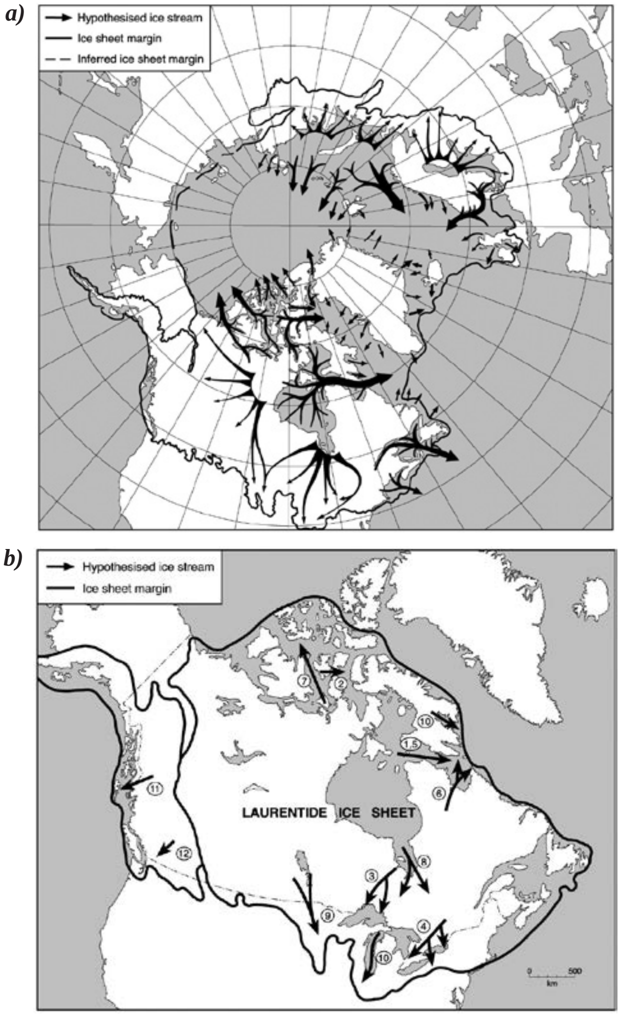

Figure 12. (a) Denton and Hughes' (1981) reconstruction of the Northern Hemisphere ice sheets at the global LGM with their hypothesised ice streams as black arrows. (b) Hypothesised ice streams based on a review of the literature from Stokes and Clark (2001). Compare with the latest inventory of 117 Laurentide ice streams shown in Figure 4.

interpreted to result from ice streaming and which included suites of level-to-streamlined fine-grained till, sometimes associated with highly elongate drumlins (Bluemle *et al*., 1993) that typically terminated towards the lobate margins, where thrusting of glacial sediment was evident in association with hummocky topography and major moraine systems. Similar, albeit much larger, patterns of streamlining were also reported by Clark (1993). Using Landsat satellite imagery, he identified a "hitherto undocumented and much larger form of ice moulded landscape" (p. 1) which comprised streamlined glacial lineations with typical lengths of between 8 and 70 km, widths between 200 and 1300 m, and spacings between 300 and 5 km. Clark (1993) termed these features 'mega-scale glacial lineations' (MSGLs) and discussed a variety of possible origins, concluding that were likely to form under conditions of extremely rapid flow such as ice streams or surges.

Thus, by the late 1990s, significant progress had been made in terms of identifying the glacial geological evidence of ice streaming on the bed of the LIS. These studies suggested that ice streaming should leave behind sedimentological evidence of fast ice flow in the form of heavily deformed tills and distinctive erratic dispersal trains that often depicted convergent flow-patterns (e.g. Dyke and Morris, 1988; Hicock, 1988; Alley, 1991; Patterson, 1997, 1998). Many of these flow-patterns, or fans (cf. Kleman and Borgström, 1996), also contained highly elongate glacial lineations, which were postulated to reflect rapid ice velocities (e.g. Clark, 1993); and some were characterised by abrupt lateral margins (e.g. Hodgson, 1994) and lateral shear margin moraines (Dyke and Morris, 1988). Taken together, these were argued to represent the key 'geomorphological criteria' for identifying palaeo-ice streams, which Stokes and Clark (1999) formalised in a series of landsystems models depending on whether the ice stream terminated in water or on land, and whether the glacial lineations were formed rapidly and synchronously or slowly and time-transgressively.

Despite much progress, however, the first systematic literature review of ice streams in the LIS (Stokes and Clark, 2001) found only 10 hypothesised ice streams that had been identified based on unambiguous glacial geological evidence (Fig. 12b). However, a large number of ice streams were uncovered in the early 2000s (e.g. Clark and Stokes, 2001; Shaw, 2003), such that by 2004, a new map of ice streams in the LIS depicted a total of 49 ice streams, 34 of which had good evidence, with the remainder more uncertain (Winsborrow *et al*., 2004). Much of the evidence was based on the terrestrial glacial geological record, but the recognition of discrete layers of ice rafted debris (IRD) in North Atlantic sediment cores (Heinrich, 1988) had also begun to implicate episodic ice streaming, particularly in Hudson Strait, as being primarily responsible for their deposition (Bond *et al.,* 1992; Andrews and Tedesco, 1992; MacAyeal, 1993; Marshall and Clark, 1997a, b; Andrews, 1998). In addition, the burgeoning growth of marine geophysical techniques saw a large number of ice stream footprints identified in offshore settings, particularly in Atlantic Canada (Shaw, 2003; Shaw *et al*., 2006; Todd *et al.,* 2007; Shaw *et al.,* 2009), but also in the Canadian Arctic Archipelago (MacLean *et al.,* 2010) and in Hudson Bay (Ross *et al*., 2011). These techniques, allied with the growth of remote sensing studies across large regions of the ice sheet bed enabled a large number of ice streams to be identified (e.g. De Angelis and Kleman, 2005, 2007; Evans *et al*., 2008; Ross *et al*., 2009; Stokes *et al*., 2009; Ó Cofaigh *et al.,* 2010).

Most recently, Margold *et al.* (2015a) compiled a new inventory of ice streams in the LIS based on an up-to-date review of the literature and systematic mapping from across the entire ice sheet bed using both terrestrial and offshore datasets. Their map (see Fig. 4) includes 117 ice streams and each ice stream is categorised according to the type of evidence it left behind, with an acknowledgement that some locations are more speculative than others. Indeed, identifying ice streams on more resistant 'hard-bed' terrain, such as the Canadian Shield, is more difficult, but recent work (e.g. Eyles

2012; Eyles and Putkinen, 2014; Krabbendam *et al.*, 2016) has described rock drumlins, megaflutes and mega-lineated terrain, which likely represent a hard-bedded landform assemblage cut by ice streams. Thus, it is unlikely that any major ice streams are missing (Margold *et al.,* 2015a, b). Indeed, most of the major ice streams are also captured in numerical modelling of the ice sheet, although the dynamics of land-terminating ice streams are much harder to reproduce (Stokes and Tarasov, 2010).

In a review of the spatial distribution of Laurentide ice streams, Margold *et al.* (2015b) noted that the pattern of ice streams (Fig. 4) during the LLGM resembled the present day velocity patterns in modern ice sheets (Fig. 13a). They estimated that around a third of the LIS margin perimeter was drained (intersected) by ice streams at the LLGM, which is a very similar value for the present-day Antarctic ice sheets. Large ice streams had extensive onset zones fed by multiple tributaries and, where ice drained through regions of high relief, the spacing of ice streams appears to show a degree of spatial self-organisation which was hinted at in the earlier work by Denton and Hughes (1981), but which has perhaps not been fully appreciated and explored. It is also clear that whilst topography exerted a primary control on fixing the location of ice many streams in the LIS, there were large areas along the western and southern margin of the ice sheet where networks of ice streams operated over soft sediments and switched direction repeatedly and probably over short (centennial) time scales (cf. Ó Cofaigh *et al*., 2000; Ross *et al*., 2009). As the ice sheet retreated on to its low relief interior, however, Margold *et al*. (2015a, b) noted that several ice streams showed no correspondence with topography or underlying geology, and were perhaps facilitated by localised build-up of pressurised subglacial meltwater (e.g. Stokes and Clark, 2003). Margold *et al.* (2015b) also highlighted that there have been very few attempts to date the initiation and cessation of the vast majority of ice streams, but that it is clear that they must have switched on and off during deglaciation, rather than maintaining the same trajectory as the ice margin retreated.

The extent to which changes in the ice stream drainage network were a cause or effect of ice sheet deglaciation is a key question. Put another way, does the drainage network of ice streams (Fig. 4) arise as a result of climatically-driven changes in ice sheet mass balance or could ice streams evolve to drive changes beyond that which might be expected from climate forcing alone? This question has rarely been addressed with respect to the LIS, but it is clear that ice streaming led to major reorganisations in the flow pattern of the ice sheet (Mooers *et al*., 1997; Veillette *et al.,* 1999; Ó Cofaigh *et al.,* 2000; Ross *et al.,* 2009; Stokes *et al.,* 2009). It is also clear that ice streaming is capable of rapidly lowering the ice sheet surface profile to lower elevations, where ablation will be increased (Robel and Tziperman, 2016). However, even where major reorganisations took place as a result of ice streaming, there is little evidence that deglaciation proceeded more rapidly. For example, if the Hudson Strait Ice Stream was responsible for major ice discharge events into the North Atlantic (e.g. Andrews and Tedesco, 1992; Andrews, 1998; Andrews and MacLean, 2003), there is tentative evidence for a reorganisation of the internal geometry and flow patterns of the ice sheet (Dyke and Prest, 1987; Mooers *et al.,* 1997; Veillette *et al.,* 1999; Dyke, 2004), but little evidence that these events resulted in a more rapid deglaciation of the area of the ice sheet, which was mostly linear through time (Fig. 9). The only exception to this is where individual outlet glaciers retreated through

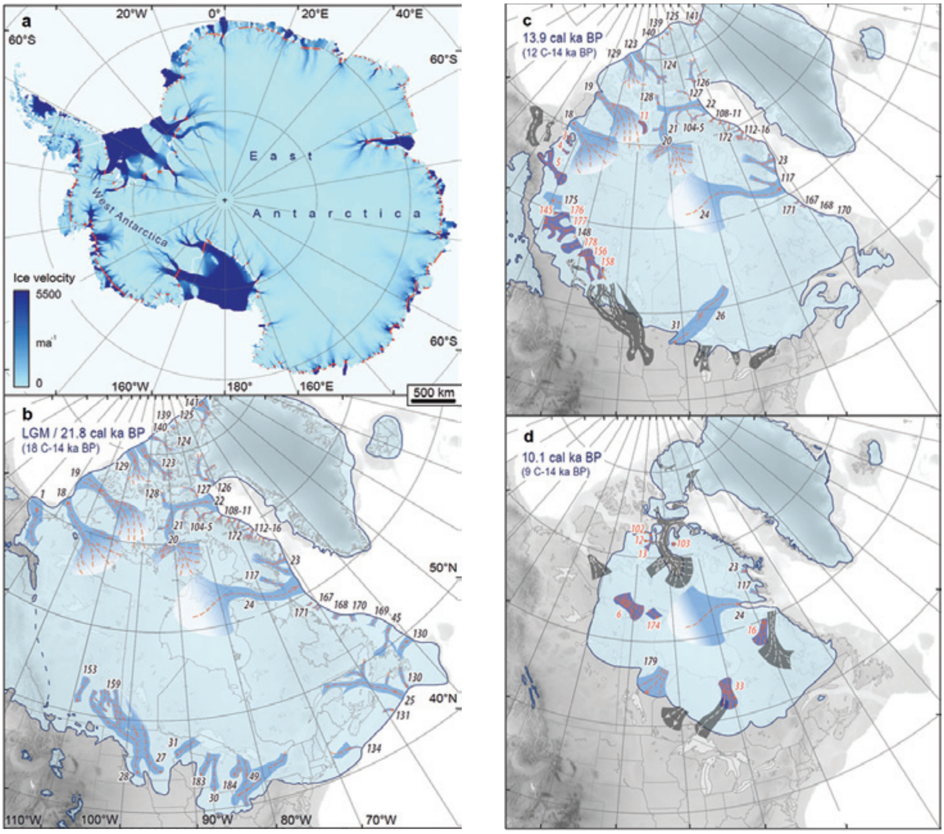

Figure 13. Ice flow velocity of the Antarctic ice sheet compared (at the same spatial scale) with reconstructions of ice stream activity in the LIS at selected time steps, from Stokes et al. *(2016). (a) Present-day Antarctic ice sheet velocity, with red lines indicating where ice streams intersect the grounding line, which is a similar cumulative proportion (~30% of the perimeter) to the LIS in (b) (Margold* et al.*, 2015b). (b, c, d) Ice streams reconstructed for the LIS at its LLGM (approximately 21.8 kyr ago), 13.9 ka, and 10.1 ka. The locations of ice streams that were active at the given time are shown in blue and numbered in black. Those that switched off within the preceding 1 ka are shown in grey and those that switched on during the subsequent 1 ka are shown in dark blue with numbers in red. For cross-referencing, the numbers refer to the inventory numbers in Margold* et al. *(2015a, b). Reprinted by permission from MacMillan Publishers Ltd: Nature (Stokes* et al.*, 2016), copyright (2016).*

deep troughs that resulted in a localised acceleration in retreat until the margin was able to re-stabilise on higher ground (Briner *et al*., 2009). It is also possible that eustatic sea-level rise early in deglaciation may have increased water depths close to marine-terminating ice streams, enhancing their discharge and leading to rapid draw-down of ice. As noted above, this might partly explain the relatively early deglaciation of the Atlantic provinces, where deglaciation was associated with a margin in deep water (Andrews, 1973; Mosher *et al.*, 1989; Piper *et al*., 1990; King, 1996; Scnitker *et al*., 2001; Shaw *et al*., 2006), but it appears to have had less effect at the marine margin in the Canadian Arctic Archipelago.

In a recent study, Stokes *et al.* (2016) examined whether the cumulative impact of ice streams was able to increase and sustain rates of mass loss during deglaciation of the LIS beyond those that might be expected from climate forcing alone. They used the Dyke *et al.* (2003) ice margin chronology to bracket the duration of the 117 ice streams in the inventory from Margold *et al.* (2015a: see Fig. 4). They found that as the ice sheet retreated, ice streams activated and deactivated in different locations (see Fig. 13) and, unsurprisingly, their overall number decreased. Perhaps more surprising was that ice streams occupied a progressively smaller percentage of the ice sheet perimeter during deglaciation. At its maximum, approximately 27% of the LIS margin was streaming, but this value decreased to between 25% and 20% from 16 ka to 13 ka, and then rapidly dropped to ~5% at 11 ka (Stokes *et al.,* 2016). This implies that the final 4 to 5 ka of deglaciation was largely driven by surface melt, which is corroborated surface mass balance modelling (see Section 4.1) and inferences based on the density of subglacial meltwater conduits (eskers) (Storrar *et al*., 2014). Stokes *et al.* (2016) also used a simple scaling relationship based on the width and discharge of modern ice streams to estimate the potential cumulative discharge from Laurentide ice streams through time, and found that this decreased and was strongly scaled to the ice sheet's volume. This scaling is also found in numerically modelled estimates of ice stream discharge (Stokes *et al*., 2012). They concluded that whilst the underlying geology and topography clearly influenced ice stream activity (cf. Marshall *et al.,* 1996), the drainage network of ice streams – at the ice sheet scale – appears to have adjusted in response to ice sheet volume. Thus, contrary to the view that sees ice streams as unstable entities that can accelerate ice-sheet deglaciation, Stokes *et al.* (2016) found that ice streams exerted progressively less influence on deglaciation of the LIS.

This is not to say, however, that ice streams did not play any role in deglaciation. They were likely to be very important in reducing the volume (if not the area) of the ice sheet in early deglaciation, when large parts of the LIS had a marine margin (Andrews, 1973), which is known to be a key control on ice streaming (Winsborrow *et al*., 2010). Tanner (1965) was one of the first to recognise the importance of glacial isostatic depression in generating relatively higher sea levels at the marine margin when the ice sheet was at near-maximum configurations. This would have increased calving losses and encouraged ice stream draw-down. More recently, numerical modelling has shown that a simple feedback between ocean forcing and isostatic adjustment can explain the observed magnitude and timing of Heinrich events from the Hudson Strait Ice Stream (Bassis *et al.,* 2017). Bassis *et al.* (2017) showed that when the LIS is at its near-maximum extent, the terminus of the ice stream remains grounded on bed topography depressed about 300 m below sea level, rendering it particularly vulnerable to sub-surface ocean warming. They argued that a small warming in the subsurface ocean is enough to trigger rapid retreat of the ice sheets into the over-deepened bed (generating a Heinrich event) and that retreat continues until isostatic adjustment allows the bed to uplift, isolating the terminus from ocean forcing. At this point, they noted that retreat ceases and, with the ice sheet at its minimum extent, bed uplift facilitates regrowth on a slower timescale than collapse.

Other numerical modelling experiments suggest that ice streaming may render large ice sheet more sensitive to Milankovitch forcing (Marshall and Clark, 2002; Robel and

Tziperman, 2016). Large ice sheets are likely to have a larger proportion of their bed above the pressure melting point and this is a first order control on the likelihood of generating ice streams (MacAyeal, 1993; Marshall and Clarke, 1997a, b; Stokes *et al*., 2012; Marshall and Clark, 200; Tarasov *et al.,* 2012; Robel and Tziperman, 2016). In their numerical modelling experiments, Robel and Tziperman (2016) found that when ice streams are sufficiently developed, an upward shift in the ELA caused by external climatic (Milankovitch) forcing results in rapid deglaciation. However, when the same shift in the ELA was applied to an ice sheet without fully-formed ice streams, it led to continued ice sheet growth or slower deglaciation. These idealised experiments were also repeated using key aspects of the climatic and geographic complexity of the LIS and generated similar results: enhanced discharge caused by ice stream acceleration is the primary source of mass loss during the early part of deglaciation in response to orbital forcing. An interesting corollary is that these processes also explain why Milankovitch forcing late in a 100 ka glacial cycle leads to full deglaciation, when the ice sheet is large, isostatically-depressed, and has developed numerous large ice streams; while the same forcing does not produce deglaciation early in the glacial cycle when the ice sheet is small and without ice streams (Marshall and Clark, 2002; Abe-Ouchi *et al.,* 2013; Robel and Tziperman, 2016).

5. Conclusions and Outlook

The LIS is thought to have initiated from an ice-free state over North America during MIS 5d, around 116-114 ka (e.g. Marshall *et al*., 2000; Stokes *et al*., 2012). It grew rapidly from its initial inception over the Arctic/sub-Arctic plateaux along the eastern seaboard of Canada and likely attained an MIS 4 maximum around 65-60 ka (Marshall *et al*., 2000; Stokes *et al*., 2012). Its extent during MIS 3 is uncertain (Dredge and Thorleifson, 1987; Stokes *et al*., 2012; Dalton *et al*., 2016), but it grew rapidly to its Local Last Glacial Maximum, which was attained around 26-25 ka (e.g. Dyke, 2002; Clark *et al*., 2009), although with some regions advancing much later, such as in the far north-west (Lacelle *et al*., 2013). After over a century of debate, a consensus has emerged that it existed as an extensive, multi-domed ice sheet that extended to the edge of the continental shelf at its marine margins, but that it was thinner (~3000 m) than some earlier work had suggested and consumed a sea-level equivalent of around 50 m (Clark *et al*., 1996). It is thought that it existed at or close to this maximal configuration for several thousand years (Dyke *et al.,* 2002).

Our understanding of the pattern and timing of deglaciation is due in no small part to several major syntheses that have remained benchmark reconstructions for several decades (e.g. Bryson *et al.,* 1969; Denton and Hughes, 1981; Boulton *et al*., 1995; Dyke and Prest, 1987; Dyke, 2004), augmented by numerical modelling, which has seen rapid developments over the last two decades (Marshall *et al*., 1996; Marshall *et al.,* 2000; Tarasov *et al*., 2012; Peltier *et al.,* 2015, Gregoire *et al*., 2015b). This body of work shows a clear asymmetry in retreat whereby the western and southern margins retreated back towards the major dispersal centres over Foxe Basin-Baffin Island and Quebec-Labrador. Ice margin retreat was relatively slow prior to ~17 ka, but it is clear that the ice sheet volume was decreasing. Between around 16 and 13 ka, however, the margin retreated rapidly, particularly along the southern and western margins, which led to the separation of the Laurentide from the Cordilleran Ice Sheet. In contrast, the northern and eastern margins of the ice sheet underwent only minimal recession (e.g. Dyke and Prest, 1987; Dyke, 2004). During the Younger Dryas, the overall net recession was reduced and several notable readvances are known to have taken place (e.g. Dyke and Savelle, 2000; Jennings *et al*., 1996; Andrews *et al*., 1998). Following the Younger Dryas, the ice sheet retreated two to five times faster than previous rates (Ullman *et al*., 2015b). Recession of the northern and eastern margins accompanied the continued rapid recession of the southern and western margins, although a series of moraine systems were built in some locations, which likely indicate temporary stabilisations (Dyke, 2004) or surges/ice streaming (Stokes and Clark, 2003). Final deglaciation of the remnant Keewatin and Foxe Domes (Dyke, 2004; Ross *et al.,* 2012; Simon *et al*., 2014), left a remnant Labrador Dome that is thought to have deglaciated by around 6.7 ka (e.g. Ullman *et al*., 2016).

The pattern and timing of deglaciation of the LIS represents a valuable analogue for understanding the rates and mechanisms of ice sheet deglaciation (Denton and Hughes, 1981; Kleman and Applegate, 2014; Margold *et al.,* 2015b; Stokes *et al*., 2016), which may be relevant to assessments of the future stability of modern-day ice sheets. In this context, it is generally accepted that the initial trigger for deglaciation was an increase in boreal summer insolation (Clark *et al.,* 2009; Ullman *et al.,* 2015a). However, modelling of the ice sheet's net surface mass balance indicates that it remained positive until around 11 ka (Ullman *et al*., 2015b). This suggests that the predominant source of mass loss was initially via rapidly-flowing ice streams, particularly at the ice sheet's marine margins (Andrews, 1973; Shaw *et al.,* 2006; De Angelis and Kleman, 2007; Stokes *et al.,* 2016). Indeed, our understanding of palaeo-ice streams has grown from almost completely ignorance in the early 1980s to the latest inventory of 117 ice streams that operated at various times during deglaciation (Margold *et al*., 2015a, b). Only when summer temperatures increased by 6-7°C relative to the LLGM did the ice sheet's surface mass balance become increasingly negative in the early Holocene (around 11.5 to 9 ka). Thereafter, deglaciation of the remaining 60% of the ice sheet's initial area was accomplished mostly by surface melt (Dyke, 2004; Carlson *et al.,* 2007, 2008, 2009a). This implies that 'dynamic discharge' via ice streams exerted progressively less influence on the deglaciation of the LIS (Stokes *et al*., 2016).

In his most recent synthesis, Dyke (2004) noted that the improved age control and more detailed mapping of deglacial patterns over the last few decades have enabled improved reconstructions of the LIS and a closer correlation between the deglaciation sequence and major climatic events recognised in the North Atlantic region and in the Greenland ice core record. That said, there remain areas of the ice sheet where deglaciation is relatively poorly constrained. The western margin of the LIS (and western interior) has emerged as a key area for a number of important debates, including the concept of an ice-free corridor and the peopling of North America (Dyke, 2004; Dixon, 2013; Pedersen *et al.,* 2016). The possibility of a rapid saddle-collapse and a large meltwater pulse (and sea-level jump) have also implicated this region (Gregoire *et al.,* 2012, 2015a), which has also been identified as a drainage route for a putative outburst flood from Glacial Lake Agassiz (Tarasov and Peltier, 2005; Murton *et al*., 2010; Fisher and Lowell, 2012). It is interesting to note that, with admirable foresight, this region was specifically highlighted almost 50 years ago by Bryson *et al.* (1969: p. 5) who stated that "it is unfortunate that there are so few available dates from the western interior; the corridor between the Cordillera and the retreating ice front is of great interest to the anthropologist, biologist, and climatologist". It is perhaps more unfortunate that, almost five decades on, the scarcity of dates from this region persists, and this is a key area for future research to address. Improvements in the ice margin chronology and the quantification of its uncertainties will also provide tighter constraints for numerical modelling of the ice sheet, which will also require improvements in terms of model resolution and in the representation of the key physics, such as the simulation of ice streams, subglacial processes and changes in meltwater drainage routes (Marshall *et al*., 2000; Hindmarsh, 2009; Stokes *et al*., 2015; Kirchner *et al*., 2016; Wickert, 2016). This should help resolve important debates about the role of the LIS during major reorganisations of the ocean-climate system (Broecker *et al.,* 1989; Bond *et al*., 1992; Barber *et al.,* 1999; Carlson and Clark, 2012), and enable improved predictions of the response of modern-day ice sheets to future climate change.

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