Reconciling records of ice streaming and ice margin retreat to produce a palaeogeographic reconstruction of the deglaciation of the Laurentide Ice Sheet

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Abstract

This paper reconstructs the deglaciation of the Laurentide Ice Sheet (LIS; including the Innuitian Ice Sheet) from the Last Glacial Maximum, with a particular focus on the spatial and temporal variations in ice streaming and the associated changes in flow patterns and ice divides. We build on a recent inventory of Laurentide ice streams and use an existing ice margin chronology to produce the first detailed transient reconstruction of the ice stream drainage network in the LIS, which we depict in a series of palaeogeographic maps. Results show that the drainage network at the LGM was similar to modern-day Antarctica. The majority of the ice streams were marine terminating and topographically-controlled and many of these continued to function late into the deglaciation, until the ice sheet lost its marine margin. Ice streams with a terrestrial ice margin in the west and south were more transient and ice flow directions changed with the build-up, peak-phase and collapse of the Cordilleran-Laurentide ice saddle. The south-eastern marine margin in Atlantic Canada started to retreat relatively early and some of the ice streams in this region switched off at or shortly after the LGM. In contrast, the ice streams draining towards the north-western and north-eastern marine margins in the Beaufort Sea and in Baffin Bay appear to have remained stable throughout most of the Late Glacial, and some of them continued to function until after the Younger Dryas (YD). The YD influenced the dynamics of the deglaciation, but there remains uncertainty about the response of the ice sheet in several sectors. We tentatively ascribe the switching-on of some major ice streams during this period (e.g. M'Clintock Channel Ice Stream at the north-west margin), but for other large ice streams whose timing partially overlaps with the YD, the drivers are less clear and ice-dynamical processes, rather than effects of climate and surface mass balance are viewed as more likely drivers. Retreat rates markedly increased after the YD and the ice sheet became limited to the Canadian Shield. This hard-bed substrate brought a change in the character of ice streaming, which became less frequent but generated much broader terrestrial ice streams. The final collapse of the ice sheet saw a series of small ephemeral ice streams that resulted from the rapidly changing ice sheet geometry in and around Hudson Bay. Our reconstruction indicates that the LIS underwent a transition from a topographically-controlled ice drainage network at the LGM to an ice drainage network characterised by less frequent, broad ice streams during the later stages of deglaciation. These deglacial ice streams are mostly interpreted as a reaction to localised ice-dynamical forcing (flotation and calving of the ice front in glacial lakes and transgressing sea; basal de-coupling due to large amount of meltwater reaching the bed, debuttressing due to rapid changes in ice sheet geometry) rather than as conveyors of excess mass from the accumulation area of the ice sheet. At an ice sheet scale, the ice stream drainage network became less widespread and less efficient with the decreasing size of the deglaciating ice sheet, the final elimination of which was mostly driven by surface melt.

Keywords: Pleistocene; Glaciation; North America; Geomorphology, glacial; Laurentide Ice Sheet; Last Glacial Maximum; Late Glacial; deglaciation; ice stream

Research highlights:

- Ice drainage network in the LIS reconstructed for the last deglaciation
- Ice stream activity linked to both margin retreat and migration of ice domes
- Transition from topographically-controlled to less frequent, broad ice streams
- Uncertainty remains about the response of the LIS to the YD in several sectors
- Deglacial ice streams mostly a reaction to specific localised ice-dynamical forcing

1. Introduction

Ice streams have long been recognised for the Pleistocene ice sheets of the Northern Hemisphere (Løken and Hodgson, 1971; Hughes et al., 1977; Denton and Hughes, 1981; Dyke and Prest, 1987a, b; Dyke and Morris, 1988; Mathews, 1991; Patterson, 1998; Stokes and Clark, 2001; Ottesen et al., 2005; Kleman and Glasser, 2007; Winsborrow et al., 2012). Most attention has been given to the largest of these ice sheets, the Laurentide Ice Sheet (LIS), where some of the first investigations of palaeo-ice streams were undertaken (Løken and Hodgson, 1971; Dyke and Morris, 1988) and where an ice-discharge pattern broadly similar to the pattern of ice flow in modern ice sheets has gradually emerged (Dyke and Prest, 1987a, b; Patterson, 1998; De Angelis and Kleman, 2005; Stokes et al., 2009; Margold et al., 2015a, b).

A large number of ice streams have been identified for the LIS and ice streams are inferred to have operated during the build-up to the Last Glacial Maximum (LGM), at the LGM, and most commonly during its deglaciation (Denton and Hughes, 1981; Dyke and Prest, 1987a, b; Patterson, 1998; Stokes and Clark, 2003a, b; Winsborrow et al., 2004; De Angelis and Kleman, 2005, 2007; Stokes et al., 2009; Stokes and Tarasov, 2010; Stokes et al., 2012;

Margold et al., 2015a, b). However, and perhaps surprisingly, ice streams have thus far not been fully included in any of the ice-sheet-wide reconstructions of the LIS evolution from the LGM to its disappearance in the Middle Holocene. We therefore have only a limited understanding of how the drainage network of ice streams and associated ice divides, domes and catchment areas interacted and evolved during deglaciation.

Denton and Hughes (1981) produced one of the first maps of putative ice stream locations and portrayed a topographically-controlled ice-stream network for the Canadian Arctic Archipelago (CAA) that, despite certain simplifications, largely resembled ice-drainage pattern shown in present-day reconstructions (De Angelis and Kleman, 2005; England et al., 2006; De Angelis and Kleman, 2007; Stokes et al., 2009; Margold et al., 2015 a, b). In contrast, the ice streams they depicted for the terrestrial portion of the ice sheet (terminating on land) were purely conceptual. Later reconstructions by Boulton et al. (1985) and Boulton and Clark (1990a, b) largely ignored ice streams, focussing instead on broader changes in flow geometry and ice divide configurations, and it was the reconstruction of Dyke and Prest (1987a, b) that first portrayed and discussed ice streams in more detail. Dyke and Prest (1987a, b) included some of the largest ice streams, most importantly the Hudson Strait Ice Stream, and they also recognised several of the smaller ice streams in the Canadian Arctic that are characterised by distinct sediment dispersal trains. However, their reconstruction lacked many of the ice streams on the continental shelf due to what is now known to be their overly restricted ice extent at the LGM (see review in Stokes, 2017). The 1990s saw a growing recognition that the southern lobes of the LIS represented terrestrial ice streams (Patterson, 1997, 1998). Subsequently, the development of objective criteria for palaeo-ice stream identification (Stokes and Clark, 1999; Stokes and Clark, 2001), their application to the research of the LIS (see e.g., Clark and Stokes, 2001; De Angelis and Kleman, 2005;

Kehew et al., 2005; Ross et al., 2006; Shaw et al., 2006), together with updated LGM ice extents on the continental shelf (England, 1999; Dyke et al., 2003; Dyke, 2004; England et al., 2006; Shaw et al., 2006), has resulted in a rapid increase in the number of ice streams that have been recognised (e.g., ~10 in Stokes and Clark, 2001; ~50 in Winsborrow et al., 2004; ~120 in Margold et al., 2015a, b).

Nevertheless, detailed reconstructions of ice streaming through time have thus far only been carried out for some specific sectors of the LIS, namely the south-western part of the CAA, Foxe Basin, the Hudson Strait region (De Angelis, 2007b; De Angelis and Kleman, 2007; Stokes et al., 2009) and the Atlantic seaboard south of Newfoundland (Shaw et al., 2006). Elsewhere, such as on the southern Interior Plains, ice streams have been studied but their evolution at the regional scale has not yet been fully constrained with the available chronological data (Evans et al., 1999; Evans et al., 2008; Ross et al., 2009; Ó Cofaigh et al., 2010; Evans et al., 2012, 2014). Furthermore, some regions of the LIS have largely escaped attention from an ice dynamical point of view; namely the central Interior Plains, the north-eastern coast of Labrador, and large parts of the LIS interior on the Canadian Shield (Margold et al., 2015a). The modelling of ice streams in the LIS has also seen some important advances (e.g. Sugden, 1977; MacAyeal, 1993; Marshall et al., 1996; Marshall and Clarke, 1997a, b; Kaplan et al., 2001; Calov et al., 2002; Stokes and Tarasov, 2010; Robel and Tziperman, 2016), but few studies have investigated the behaviour of ice streams throughout deglaciation. In addition to the complexity of the physics involved, a key limitation has been a lack of information on the location and timing of ice streams within the ice sheet that could either be compiled into an empirical reconstruction of ice streaming activity or used to test numerical modelling results (Stokes et al., 2015).

Here we build on and extend recent work on LIS ice streams. Margold et al. (2015b) produced an updated inventory of Laurentide ice streams based on a review of the literature and new mapping from across the ice sheet bed (reviewed in Margold et al., 2015a; Fig. 1). Using this inventory and the ice margin chronology of Dyke et al. (2003), Stokes et al. (2016a) recently bracketed the duration of each ice stream and calculated their likely discharge during deglaciation, guided by empirical data from modern ice streams. A key conclusion was that ice streaming was strongly scaled to the ice sheet volume and likely reduced in effectiveness during ice sheet deglaciation. Here, we extend that work by reconciling ice stream activity and the associated changes in ice stream catchments (and ice divides and domes) with the ice margin chronology (Dyke et al., 2003) into a palaeogeographic reconstruction of the LIS. We then discuss the reconstructed ice sheet evolution during the Late Glacial and early Holocene in the context of the available information on climate forcing and other possible drivers of ice stream activity.

Fig. 1 here (full-page width)

Table 1 here

2. Methods

2.1. Data

To reconstruct ice stream activity in the LIS we adopt the dating of ice stream operation presented by Stokes et al. (2016a), who used the recently-compiled inventory of Laurentide ice streams (Margold et al., 2015b) in combination with the North American ice retreat chronology of Dyke et al. (2003). The ice retreat chronology of Dyke et al. (2003), the construction of which is briefly described in Dyke (2004) and in the metadata of the 2003 Open File, builds on decades of earlier research (Prest et al., 1968; Bryson et al., 1969; Prest

1969; 1970; Dyke and Prest 1987a, b; Fulton, 1989) and combines the interpretation of the geomorphological and geological record (moraine systems, esker networks, drumlin orientation, regionally recognised tills, glaciolacustrine sediments) with a large set of ^{14}C ages, most of which are minimum deglaciation ages. This ice retreat chronology is the most up-to-date source of information for the entire ice sheet, but recent studies have shown that it significantly underestimates the ice extent on the continental shelf (e.g., England et al., 2006; Shaw et al., 2006; Rashid and Piper, 2007; England et al., 2009; Li et al., 2011; Batchelor et al., 2013a, b, 2014; Jakobsson et al., 2014; Brouard and Lajeunesse, 2017). Whilst there is now a consensus that grounded ice occupied large stretches of the continental shelf during the LGM (see review in Stokes, 2017), an exact chronology has not yet been established in most of the shelf areas. Thus, in some regions such as the northern CAA or Atlantic Canada, around Newfoundland, we use regional deglaciation models (England et al., 2006, resp. Shaw et al., 2006). Other regions, such as the shelf off the northeast coast of Baffin Island or the Labrador shelf, remain largely undescribed with respect to ice retreat chronology; here, the dating of ice streams is an approximation building on the better studied areas adjacent to the region in question. Additional minor changes to the ice margin chronology have been implemented based on the ongoing community effort to update the chronology of Dyke et al. (2003) that is carried within the framework of the MOCA (Meltwater routing and Ocean-Cryosphere-Atmosphere response) group of the International Union for Quaternary Research (INQUA). Where the ice margin chronology diverges from that of Dyke et al. (2003), we provide the necessary information in the Supplementary Data.

The ice margin chronology by Dyke et al. (2003) starts at 18 ^{14}C ka, and thus forms our starting point from which we reconstruct ice stream activity, although it is likely that many ice streams initiated prior to the LGM, especially those controlled by underlying topography (see Stokes et al., 2012). In this manuscript, we refer to 18 ^{14}C ka ≈ 21.8 cal ka as the LGM, even though the maximum extent of the ice sheet margin in different regions was reached at different times and the LGM for the whole LIS lasted a few thousand years (Dyke et al., 2002; Clark et al., 2009; Stokes, 2017).

The accuracy of our reconstruction of the ice drainage network evolution is dependent on the accuracy of the ice margin chronology (additional uncertainty then stems from the method to determine the time of ice stream activity – see next section). The database of ^{14}C ages, on which the deglaciation chronology of Dyke et al. (2003) is based, contains ~4000 individual dates (Fig. 2). The spatial distribution of these is highly uneven; more easily accessible field locations such as the Great Lakes or New England have much denser coverage than remote regions of the Canadian North (Fig. 2). For New England, in particular, the uncertainty has further been narrowed by the existence of an independent varve chronology to which the radiocarbon chronology has been linked (Ridge and Larsen, 1990; Ridge et al., 1999, 2001; the latest version, postdating Dyke et al. [2003]: Ridge et al., 2012). In contrast, the region that has the sparsest coverage of ^{14}C dates is Keewatin, where one of the major domes was located (Figs. 1, 2). It is especially the south-eastward and eastward retreat of the western ice margin, where some of the highest retreat rates were reconstructed, that has extremely loose chronological control (Dyke et., 2003; Dyke, 2004; Fig. 2). In addition, dates on the continental shelf, besides sparse coverage in some regions, suffer from the marine reservoir effect, that is still not well quantified in most of the concerned areas (e.g., Stern and Lisiecki, 2013; Jennings et al., 2015). The dating issues are further aggravated in the Beaufort Sea where no datable material was deposited until about 11.5 ^{14}C ka (Kaufman et al., 2004; England and Furze, 2008; Lakeman and England, 2013).

Fig. 2 here (full-page width)

An additional uncertainty, which cannot be quantified, is introduced by the conversion of the radiocarbon time, in which the ice margin chronology of Dyke et al. (2003) has been compiled, to calendar years (see Section 2.2.). It needs to be noted that such a conversion is not a radiocarbon date calibration that would assign a single date a range of ages, which would have to be undertaken for ~4,000 dates. Instead we simply convert the ages in ^{14}C time to corresponding median probability calendar years based on the mixed Northern Hemisphere calibration function (combining Marine and INTCAL calibration curves) in Calib 7.0 (Stuiver et al., 2017).

2.2. Bracketing the age of ice stream activity

To determine when an ice stream was in operation, Stokes et al. (2016a) considered it to be active when the 'known' ice margin was either a short distance (distally) from the known ice stream track (bedform imprint) or cut across the track (Fig. 3). Their dating of ice stream activity was based largely on the Dyke et al. (2003) isochrones, with less consideration given to the individual radiocarbon dates on which the isochrones are based (Fig. 3). The reasoning was that the isochrones, while being a local approximation of the actual ice margin position at the time, represent a regionally consistent model of ice retreat.

To account for and quantify the above uncertainty, Stokes et al. (2016a) identified the best estimate time for both the start and the end of operation of each individual ice stream. In addition, they identified the earliest and the latest possible start and end of operation for each ice stream based on the available chronology (Fig. 3; see Supplementary Data for more information related to individual ice streams). That allowed them to calculate the longest and the shortest possible time of operation as well as the best estimate duration of operation for each ice stream (Fig. 4). In cases when the shortest possible time of operation was negative (the latest possible start of operation falling beyond the earliest possible stop of operation, which happens for smaller ice stream tracks in regions with high reconstructed ice margin retreat rate – see Supplementary Data), they nevertheless assumed that the ice stream must have operated for at least 100 years.

The method used to bracket the time of operation of Laurentide ice streams carries with it additional uncertainty. In most cases this uncertainty is captured within the approach that assigns each ice stream a minimum, a best estimate, and a maximum time of operation. Examples of such types of ice streams where determining the time of operation carries significant uncertainty might be the fan-like lobes at the south-western margin (nos.179, 180 in Figs. 1, 4). The fan-shaped ice stream track likely indicates a one-off fast ice flow event (in the literature on the regional glacial history often called a surge, even though a surging glacier *sensu stricto* should undergo repeated periods of advance and quiescence; Raymond, 1987). The reasoning for this is that a prolonged activity of these large lobes would probably not be sustainable within an ice sheet that at the time had only a limited accumulation zone. However, when the time of operation is determined based on the existing ice margin isochrones, assuming that fast ice flow continued as long as the ice margin kept a lobate form, the length of operation of these lobes reaches over one thousand years. Another type of ice stream that carries additional uncertainty are those for which it is unclear whether they experienced fast ice flow only as tributaries of other ice streams, and thus were located far from the ice margin, or whether they also operated as ice streams in their own right at the time when they were proximal to the ice front (e.g., nos. 26, 174 in Figs. 1, 4).

Fig. 3 here (column width)

Fig. 4 here (column width)

In addition, the dating approach does not acknowledge the uncertainty in the time of activity of long-lived ice streams (typically operating from the LGM and throughout the early stages of deglaciation) whose ice fronts stayed stable for a considerable time. Such ice streams are treated as active for the whole time their track was proximal to the ice margin. Nevertheless, they could have either switched on later, being inactive early on, for example because of their basal thermal regime, or there could have been periods of quiescence when their ice discharge was considerably decreased. This caveat is most relevant for the Hudson Strait Ice Stream, for which periods of activity and quiescence have been suggested in connection with the periodicity of Heinrich events (see reviews of Andrews and MacLean, 2003; Hemming, 2004).

To summarise, we take a conservative approach to try and capture and quantify all sources of error (both methodological and chronological). However, these uncertainties are generally very small (i.e. of the order of a few hundred years) in the context of a pan-ice sheet reconstruction spanning ~15 ka where our aim is to broadly capture the major changes in the drainage network of ice streams.

2.3. Reconstructing ice sheet configuration

The evolving ice stream drainage network was interconnected with the overall ice sheet geometry and we attempt to determine the influence that these changes had on the position of ice divides in the ice sheet. For the LGM, we use the general ice-sheet configuration (the positions of ice domes, ice saddles and ice divides) from the earlier ice-sheet-scale reconstruction of the LIS by Dyke and Prest (1987a, b) that has yet to be superceded and the salient aspects of which are generally reproduced in both numerical simulations and glacioisostasy-based reconstructions of the ice sheet (Peltier, 2004; Tarasov et al., 2004; Peltier et al., 2015; Lambeck et al., 2017). While Dyke and Prest (1987a, b) did not describe in detail the principles by which they reconstructed the ice sheet geometry, it can be assumed that they employed their expertise to combine information contained in the available geomorphological and geological record (ice flow patterns, dispersal trains, the tilt of proglacial lake shorelines, etc.) to derive a glaciologically-plausible reconstruction of the ice sheet's major domes and ice divides. We complement Dyke and Prest (1987a, b) with more recent studies reconstructing the deglacial ice dynamics at a regional scale (Clark et al., 2000; England et al., 2006; Shaw et al., 2006; De Angelis, 2007b; De Angelis and Kleman, 2007; Stokes et al., 2009) for additional information on ice sheet geometry. Some of these studies used a more formalised approach that inverts the glacial geomorphological record into palaeo-ice sheet dynamics (these methods are described in Clark, 1993; Kleman and Borgström, 1996; Clark, 1997; Kleman et al., 2006).

We combine the information on the ice sheet configuration derived from literature with the reconstructed ice stream network (Margold et al., 2015a, b; Stokes et al., 2016a) to reconstruct the evolution of the ice sheet throughout its deglaciation. In doing so, we follow the principles described in Section 2.5 of Greenwood and Clark (2009). These include ice divides being fitted upstream of flow imprints and in an overall scheme that attempts to follow the symmetry (central positions for divides) and structure (divide branching) of modern ice sheets and a rule for adopting minimum complexity. Locally, the recently reconstructed ice stream network requires modifications in the ice sheet geometry suggested by earlier studies; which is discussed in Section 4.

3. Ice streaming activity at the LGM and throughout deglaciation

In this section, we describe the evolution of the ice stream network at key time-steps throughout deglaciation. This is illustrated in Fig. 5 and the names and numbers of individual ice streams are cross-referenced in Table 1 (more detailed information on the individual ice streams is available in the Supplementary Data).

3.1. LGM ice extent and dispersal centres (domes)

The LIS occupied large portions of the continental shelf during the LGM and it likely reached to the shelf edge in most areas of Atlantic Canada, on the Labrador and Baffin shelves and around the CAA (Briner et al., 2006; England et al., 2006; Shaw et al., 2006; Li et al., 2011; Lakeman and England, 2012; Jakobsson et al., 2014; Brouard and Lajeunesse, 2017), although this requires confirmation in some regions. The LGM extent, as we depict it in this study, is based on the following: (1) major cross-shelf troughs have been shown to be occupied all the way to the shelf break at or around the LGM (Piper and Macdonald, 2001; Andrews and MacLean, 2003; Rashid and Piper, 2007; Li et al., 2011; Batchelor et al., 2013a, b, 2014; Brouard and Lajeunesse, 2017), (2) the LIS attained its furthermost Pleistocene extent along its western and north-western terrestrial margin during Oxygen Isotope Stage 2 (OIS; Duk-Rodkin and Hughes, 1991; Young et al., 1994; Zazula et al., 2004; Jackson et al., 2011), and (3) the North American Ice Sheet Complex has been modelled to reach its maximum Pleistocene ice volume, or values close to it, during the Late Wisconsinan (Marshall et al., 2000; Bintanja and van de Wal, 2008; Tarasov et al., 2012), while there are indications that it was considerably smaller during the penultimate glacial maximum (OIS 6; Naafs et al., 2013; Colleoni et al., 2016). Given that the LIS attained an extent and volume close to its postulated Quaternary maximum during the LGM (the other candidates being OIS

12 and, in particular, OIS 16 [Bintanja and van der Wal, 2008; Naafs et al. 2013]), we assume that all cross shelf troughs were filled with ice to the shelf edge at the LGM (Batchelor et al., 2013b). The logic here is that recent work clearly points to a more extensive LIS at the LGM than what the older reconstructions depicted (Dyke and Prest, 1987a,b; Dyke et al., 2002) and even where troughs have not been studied in detail, the simplest assumption is that they were occupied by ice at this time of maximum ice extent. However, we acknowledge that this “big ice” model (see Miller et al., 2002) needs to be tested in the field for the less-researched areas, such as the shelves off the north-western coast of the Ellesmere and the north-eastern coast of Baffin islands and the Labrador Shelf.

The commonly accepted LGM ice sheet configuration consists of three main ice domes within the LIS: the Keewatin Dome in the centre of the continental Canadian North, the Québec-Labrador Dome east of James Bay in Québec, and the Foxe-Baffin Dome centred approximately on the Prince Charles Island in Foxe Basin (Figs. 1, 5a; Dyke and Prest, 1987a, b; Dyke et al., 2002). An independent ice mass of the Innuitian Ice Sheet covered Queen Elisabeth Islands and its main ice divide stretched over Devon Island and the eastern portions of Ellesmere Island (Figs. 1, 5a; England et al., 2006). A semi-independent sector of the LIS, the Appalachian Ice Complex, covered Atlantic Canada. Its main divides stretched in a roughly NW-SE direction over Newfoundland and over New Brunswick and Nova Scotia, respectively (Figs. 1, 5a; Shaw et al., 2006). An ice saddle over the Interior Plains east of the Canadian Rocky Mountains connected the LIS with the Cordilleran Ice Sheet (CIS) in the west (Figs. 1, 5a; Dyke and Prest, 1987a, b; Dyke et al., 2002). This ice sheet geometry is broadly supported by glacial geological evidence (Fulton, 1989) as well as by numerical modelling studies (Tarasov and Peltier, 2004) and the pattern of glacial isostatic rebound rates (Peltier, 2004; Lambeck et al., 2017).

3.2. 21.8 cal ka (18 ^{14}C ka) – LGM ice stream drainage pattern

At the LGM, ice streams in the LIS drained ice towards the ice sheet margin in the north, east and south, while in the west the LIS coalesced with the CIS (Fig. 5a; Margold et al., 2015a; Stokes et al., 2016a). Large ice streams existed in the marine channels of the CAA and in the cross-shelf troughs that were fed by multiple fjords along the high relief coasts of north-eastern Baffin and north-western Ellesmere islands and Labrador. Ice streams in these sectors were largely topographically controlled and the ice drainage pattern in these sectors resembled that of the Antarctic ice sheets today. The terrestrial ice margin in the south was drained by several large ice streams that fed extensive ice lobes protruding from the ice margin and which were at least partially topographically controlled, despite the modest relief of the regional landscape (Mathews, 1974). No present-day analogue exists to this environment.

The north-western and north-eastern marine margins (to the Beaufort Sea and the Arctic Ocean, and the Baffin Bay, the Labrador Sea and the North Atlantic, respectively) were characterised by a stable ice drainage network due to the strong topographic control (Margold et al., 2015a). According to our reconstruction, ice streams operated at the LGM and throughout the early stages of deglaciation in these regions (Fig. 5a-f). The westernmost ice stream draining to the Beaufort Sea was the Mackenzie Trough Ice Stream (no. 1 in Figs. 1, 5; Kleman and Glasser, 2007; Brown, 2012; Batchelor et al., 2013a, b; Margold et al., 2015a, b). It was likely fed mainly by Keewatin ice from the south-east (see Fig. 13 in Kleman and Glasser, 2007 and Fig. 5-4 in Brown, 2012), but possibly also by ice from the Cordillera through the CIS-LIS saddle. While we draw the maximum extent of the north-western LIS at 21.8 cal ka (18 ^{14}C ka) and maintaining this position for the next few thousand years (see Dyke et al., 2003; Dyke, 2004), there is some indication that the maximum extent in this part of the ice sheet might have been reached later, at about 19–18 cal ka (Kennedy et al., 2010; Lacelle et al., 2013), which would have implications for the switching on of the Mackenzie Trough Ice Stream. Indeed, the furthermost ice extent in the Mackenzie delta might have only occurred for a brief period between 17.5 and 15 ka, or even only 16.6 and 15.9 ka, according to recent optically stimulated luminescence ages from the area (Murton et al., 2015).

The two main arteries draining Keewatin ice to the Arctic Ocean were the Amundsen Gulf and M'Clure Strait ice streams, occupying large marine channels in the western part of the CAA (nos. 18 and 19 in Figs. 1, 5; Sharpe, 1988; Hodgson, 1994; Clark and Stokes, 2001; Stokes, 2002; De Angelis and Kleman, 2005; Stokes et al., 2005; Stokes et al., 2006; Kleman and Glasser, 2007; Stokes et al., 2009; Batchelor et al., 2013b, 2014). The Innuitian Ice Sheet was drained to the northwest by a series of ice streams: two larger ones in the Prince Gustaf Adolf Sea (no. 129 in Figs. 1, 5; Jakobsson et al., 2014; Margold et al., 2015a, b) and Massey Sound (no. 123 in Figs. 1, 5; Lamoureux and England, 2000; Atkinson, 2003; England et al., 2006; Jakobsson et al., 2014; Margold et al., 2015a, b), an ice stream in Nansen Sound, draining the central parts of Ellesmere Island (no. 124 in Figs. 1, 5; Sugden, 1977; Bednarski, 1998; England et al., 2006; Jakobsson et al., 2014; Margold et al., 2015a, b), and several smaller ice streams draining the mountainous north-western coast of Ellesmere Island (nos. 140, 139 and 125 in Figs. 1, 5; Margold et al., 2015a, b). The ice saddle connecting the Innuitian and Greenland ice sheets was drained by Kennedy-Robeson Channel Ice Stream to the north (no. 141 in Figs. 1, 5; Jakobsson et al., 2014; Margold et al., 2014a, b) and by Smith Sound / Nares Strait Ice Stream to the south (no. 126 in Figs. 1, 5; Blake et al., 1996;

England, 1999; England et al., 2004; England et al., 2006; Q. Simon et al., 2014; Margold et al., 2015a, b).

Another major drainage route for Keewatin ice was the ice stream system of Gulf of Boothia and Lancaster Sound ice streams (nos. 20 and 22 in Figs. 1, 5; Sugden, 1977; Dyke et al., 1982; Dyke, 1984; Dyke and Dredge, 1989; Dredge, 2000, 2001; Hulbe et al., 2004; De Angelis and Kleman, 2005; Briner et al., 2006; De Angelis, 2007a, b; De Angelis and Kleman, 2007; Kleman and Glasser, 2007; Li et al., 2011; Q. Simon et al., 2014; MacLean et al., 2017; Furze et al., 2018). Foxe ice was drained to the north, northeast and east by a large number of ice streams that fed off numerous fjords in the mountainous north-eastern coast of Baffin Island and crossed the continental shelf in Baffin Bay (Figs. 1, 5; Løken and Hodgson, 1971; Sugden, 1978; Briner et al., 2005; Briner et al., 2006; De Angelis, 2007b; De Angelis and Kleman, 2007; Briner et al., 2008; Briner et al., 2009; Batchelor and Dowdeswell, 2014; Margold et al., 2015a, b; Brouard and Lajeunesse, 2017). Southern portions of the Foxe-Baffin sector were drained by the Hudson Strait Ice Stream (no. 24 in Figs. 1, 5; see Supplementary Data and Andrews and MacLean [2003] and Margold et al. [2015a] for reviews of the large body of literature pertaining that ice stream) that functioned as the only eastern outlet for Keewatin ice and also drained the north-western portions of the Québec-Labrador sector. The Hudson Strait Ice Stream is thought to have been the dominant source of ice discharged during the Heinrich Events (Andrews and MacLean, 2003; Hemming, 2004), discussed in Section 4.2.3.

The Québec-Labrador Ice Dome was drained to the Labrador Sea by several ice streams on the continental shelf (nos. 167-171 in Figs. 1, 5; Josenhans et al., 1986; Josenhans and Zevenhuizen, 1989; Margold et al., 2015a, b), as were the local ice dispersal centres of the

Appalachian Ice Complex in Atlantic Canada (nos. 45, 130 and 133 in Figs. 1, 5; Shaw, 2003; Shaw et al., 2006; Shaw et al., 2009; Rashid et al., 2012; Margold et al., 2015a, b). The most important ice discharge route for the Québec-Labrador Ice Dome was the Laurentian Channel Ice Stream (no. 25 in Figs. 1, 5; Grant, 1989; Occhietti, 1989; Mathews, 1991; Keigwin and Jones, 1995; Piper and Macdonald, 2001; Shaw et al., 2006; Shaw et al., 2009; Rashid et al., 2012; Eyles and Putkinen, 2014; Margold et al., 2015a, b), which is evidenced by its long and distinct trough (see Figs. 1 and 2 in Shaw et al., 2006).

The south-eastern marine margin in Atlantic Canada and the north-eastern United States was the first portion of the ice sheet to start retreating from its local LGM positions on the outer shelf in the Gulf of Maine (Schnitker et al., 2001; Clark et al., 2009; Stokes, 2017). While we draw the ice stream in the Northeast Channel as still operating at 21.8 cal ka (18 ^{14}C ka), which is based on the 18 ^{14}C ice margin position of Dyke et al. (2003), it might have already switched off before this time. Shaw et al. (2006) suggested that the ice stream might have only operated during the early LGM and the downstream part of the channel might have become deglaciated as early as 23.5 cal ka. Another ice stream, identified by Siegel et al. (2012) offshore Massachusetts (no. 135 in Fig. 1) is thought to precede the last glacial, possibly being of MIS 12 age (Siegel et al., 2012).

The southernmost limit of the Wisconsinan LIS ice margin was reached in the Great Lakes region, where ice streams occupied the lake basins (nos. 30, 31, 49, 183, 184 in Figs. 1, 5; Whittecar and Mickelson, 1979; Clayton et al., 1985; Beget, 1986; Karrow, 1989; Clark, 1992; Hicock, 1992; Hicock and Dreimanis, 1992; Breemer et al., 2002; Lian et al., 2003; Kehew et al., 2005; Jennings, 2006; Eyles, 2012) and drained ice towards the digitate ice margin. The ice lobes of the Great Lakes are thought to have experienced several cycles of advance and retreat over the LGM time, with changing dominance of particular lobes (Kehew et al., 2005; Larson, 2011; Syverson and Colgan, 2011). As such, our depiction of the ice drainage network in this region at the LGM and its reconstructed development through time are likely to be over-simplified. The long-axis of the lake basins might not have always been the preferred route for fast ice flow. This applies, for example, to the Lake Superior basin, where the ice flow might have initially been oriented more north-south, crossing the northeast-southwest oriented lake basin, and for the basins of lakes Ontario and Erie that might have been crossed by ice flowing from the north (see Fig. 5a-d; Dyke and Prest, 1987b). Locally, there is also evidence of fast ice flow where the timing of the ice streaming is disputed, such as for the Wadena drumlin field related to the Alexandria moraine in central Minnesota (Wright, 1962; Goldstein, 1989; Sookhan et al., 2016; see Knaeble [2006] for a discussion of the timing). In the north, the onset zones of the Great Lakes ice streams extended to the hard beds of the Canadian Shield (Eyles, 2012; Krabbendam et al., 2016).

Further west along the southern margin, on the Interior Plains, we draw two major lobes, Des Moines Lobe and James Lobe, as active ice streams at the LGM (nos. 27, 28 in Figs. 1, 5; Clayton et al., 1985; Mathews, 1991; Clark, 1992; Patterson, 1997, 1998; Hooyer and Iverson, 2002; Jennings, 2006; Carlson et al., 2007; Lusardi et al., 2011). Ross et al. (2009) associate these lobes with the smaller ice streams (nos. 152 and 153 in Figs. 1, 5; Ó Cofaigh et al., 2010; Evans et al., 2014; Margold et al., 2015a, b) cross-cutting the Maskwa Ice Stream (no. 153 in Figs. 1, 5; Ó Cofaigh et al., 2010; Lusardi et al., 2011; Evans et al., 2014; Margold et al., 2015a, b), assuming that the small ice streams functioned as tributaries of the Des Moines and James lobes and the whole system was only active from about 16.5 cal ka. We agree with Ross et al.'s (2009) interpretation for the time period following 16.5 cal ka; for the time preceding that period, we follow the logic that because the lobes feature in the

LGM ice margin of Dyke et al. (2003), the existence of these large protrusions in the ice sheet margin (based on glacial geological evidence) implies fast ice flow to sustain them. We thus draw them as operating synchronously with the Maskwa Ice Stream and with all three ice streams (Des Moines, James, Maskwa) draining the Keewatin Ice Dome (Fig. 5a). However, we acknowledge that the Des Moines and James lobes have undergone oscillations that were not synchronous and which we are unable to capture in our pan-ice –sheet study. Moreover, the limit of the two ice lobes is not well constrained for the time period preceding 17 cal ka (see Clayton and Moran, 1982; Dyke and Prest, 1987a). Consequently, we have little information on the source areas of the Des Moines and James lobes at the LGM. Dyke and Prest (1987b) draw them to be fed by Hudson ice that was flowing southwest, which would for the Des Moines Lobe imply being fed by the Wadena Lobe, and for the James Lobe being fed from the area of Lake Winnipeg. In contrast, the regional geomorphology, with well-developed (though shallow) troughs bounded by massive moraines, perhaps hints at the stability of the ice drainage network in this region. Indeed, Patterson (1997, p. 251) states "The Des Moines Lobe followed the course of the 20-30 ka BP advance through western Minnesota along the Minnesota River valley", which implies that ice followed a south-eastern course in the down-ice portion of the lobe. Patterson (1997) further suggests that the catchment area of the Des Moines Lobe roughly coincided with the extent of the province of Manitoba (see Figs. 6 and 7 in that paper). Based on lithological properties of distinct till sheets, Lusardi et al. (2011) were able to distinguish a gradual shift in the catchment of the Des Moines Lobe from a northwestern (Buffalo Corridor, no. 159 in Figs. 1, 5) to northern source during its evolution in the Late Glacial , but their study does not bring any evidence for an early northern source of the lobe (cf. Dyke and Prest, 1987b). A possible explanation might be in the position of the major saddle between the Keewatin and Québec-Labrador ice domes. Dyke and Prest (1987b) drew this saddle directly north of Lake Winnipeg, roughly over Southern Indian Lake (Fig. 5a), but if this was the source area of the Des Moines Lobe, as portrayed in Patterson (1997), the saddle might have instead been positioned farther east, approximately where the main ice divide crossed the Ontario-Manitoba border or east of it (see Fig. 5a).

In the west, the LIS coalesced with the CIS at the LGM (Dyke et al., 2003). However, the precise timing of the coalescence of the two ice sheets remains unclear (Dyke, 2004; Stokes, 2017). A ^{14}C dated horse bone from the Edmonton area indicates that the coalescence must have occurred more recently than ~25.5 cal ka (Young et al., 1994). Traces of fast ice flow from the period before the coalescence occur throughout the region (Winefred Lake fragment, no. 157; Pre-Maskwa Ice Stream, no. 150; Saskatchewan River Ice Stream, no. 162; Margold et al., 2015a, b; Fig. 1) and indicate ice drainage in NE-SW direction, i.e. the sourcing of ice from a more easterly location than that displayed by the LGM ice streams (Fig. 6a). The build-up of the Keewatin Ice Dome, its dominance over Hudson ice, and later the influence of the Cordilleran ice and the coalescence of the two ice sheets caused a gradual change in the ice flow over southern Saskatchewan from NE-SW to N-S and NW-SE (Fig. 6; Clayton and Moran, 1982; Dyke and Prest, 1987a, b; Margold et al., 2015a).

Fig. 5 here (full-page width)

3.3. Changes in the ice stream activity during late LGM and early Late Glacial

Changes in the ice streaming activity in the late LGM time and during the early Late Glacial were mainly related to the retreat of the marine margin in Atlantic Canada, a region where the ice sheet started to retreat early (Schnitker et al., 2001; Shaw et al., 2006), and to the continuing build-up of the CIS-LIS ice saddle on the Interior Plains that lagged behind the global LGM because the CIS reached its maximum extent in the south later than other ice sheets of the Northern Hemisphere (Booth et al., 2003; Clark et al., 2009; Clague and Ward, 2011).

3.3.1. 21.8-20.5 cal ka (18-17 ^{14}C ka)

In Atlantic Canada, the Bay of Fundy Ice Stream (no. 185 in Figs. 1, 5b-d) commenced operation in connection with the retreat and eventual shut-down of the Northeast Channel Ice Stream (no. 134 in Fig. 5b). However, the exact configuration of ice divides in the southern part of the Appalachian Ice Complex (over New Brunswick and Nova Scotia) is not well understood. Stea et al. (1998) and Stea et al. (2011) draw the ice divide of the so-called Escuminac Phase (local LGM) in a roughly W-E direction, running towards Magdalen Islands in the Gulf of St. Lawrence with ice crossing the long axis of Nova Scotia at a right angle, including the Bay of Fundy. In contrast, Shaw et al. (2006) draw the ice divide of the Escuminac Phase in a NW-SE direction, running across Nova Scotia to the continental shelf, with ice drawn into the Bay of Fundy (Fig. 5a). These studies also differ in the timing of the establishment of the Scotian ice divide that stretched along the long axis of the peninsula in the late LGM time. Shaw et al. (2006) portray the Scotian ice divide at about 23.5 cal ka, whereas Stea et al. (1998, 2011) place the reconfiguration from the Escuminac Phase to ~ 20.5 cal ka. Here, we follow Shaw et al. (2006) and draw the Scotian ice divide from the first time step we depict (Fig. 5a) because we have adopted their regional model of deglaciation for Atlantic Canada.

In the southwest of the ice sheet, a series of three ice streams; the Central Alberta, High Plains, and Rocky Mountain Foothills ice streams (nos. 14,15, 151 in Fig. 5b), forming together an anastomosing ice stream system with one tributary ice stream from the CIS, came into operation at about 20.5 cal ka (the timing is not well-constrained; see Supplementary Data). They marked the growth of the western portion of the Keewatin Ice Dome and/or the migration of the dome to the west, and the build-up of the CIS-LIS ice saddle, facilitated by the flow of the Cordilleran ice over the main ridge of the Rocky Mountains (Bednarski and Smith, 2007; Margold et al., 2013; Seguinot et al., 2016). Cordilleran ice was drained by the Rocky Mountain Foothills Ice Stream into the High Plains Ice Stream (Fig. 5b). In the north, Cordilleran ice likely supplied some of the ice drained by the Mackenzie Trough Ice Stream system.

3.3.2. 20.5-19.3 cal ka (17-16 ^{14}C ka)

We depict few changes between 20.5 and 19.3 cal ka. We note the recently published early deglaciation dates and a complex glacial record from Magdalen Islands in the Gulf of St. Lawrence (Rémillard et al., 2016) that led the authors to consider unorthodox ice configurations within the Appalachian Ice Complex. Rémillard et al. (2016) suggest an early shutdown of the Laurentian Channel Ice Stream, possibly associated with an ice advance from Newfoundland crossing the Laurentian Channel as a result of the debuttressing of the Newfoundland Ice Cap. However, we defer the incorporation of these new findings into our broad-scale model until they can be incorporated in regional reconstructions of the Late Glacial ice dynamics in Atlantic Canada.

While the oscillations of the ice margin in the region of the Great Lakes in relation to the varying dynamics of the particular ice lobes were likely rather complex (Mickelson and Colgan, 2003; Kehew et al., 2005), we depict all of the ice lobes as streaming throughout deglaciation of the area and only the Saginaw Lobe (no. 184) as switching off earlier (Fig. 5b, c), at about 20.5 cal ka, after the early advance over the southern Michigan Upland (see Supplementary Data for comments; Kehew et al., 2005).

3.3.3. 19.3-18.2 cal ka (16-15 ^{14}C ka)

At about 19 cal ka, following a retreat of the ice margin towards the Newfoundland coast, fast ice flow, driven by opening calving bays, reached back to Conception Bay (no. 182 in Figs. 1, 5c-e). Progressing retreat of the marine ice margin, the decay of this sector of the LIS, and a transition to individual local ice caps marked the cessation of ice stream activity in the troughs on the continental shelf in Atlantic Canada south of the Labrador Sea. Based on the available regional model of deglaciation (Shaw et al., 2006), and on new deglaciation ages provided by John Shaw for the ongoing effort to update the ice margin chronology of Dyke et al. (2003), we draw the Notre Dame Channel Ice Stream (no. 45 in Figs. 1, 5a-c) and the Hawke Saddle Ice Stream (no. 169 in Figs. 1, 5a-c) as switching off at about 19 cal ka (note that the tracks of the both ice streams were largely beyond the maximum ice extent of Dyke et al., 2003). The Trinity Trough and Placentia Bay–Halibut Channel ice streams (nos. 130 and 133, respectively, in Figs. 1, 5a-d) continued to function to about 18 cal ka.

3.3.4. 18.2-17 cal ka (15-14 ^{14}C ka)

Shortly after ice retreat from the outer shelf in Atlantic Canada, the sea transgressed into the glaciated areas of the inner shelf. Large calving bays opened in the Gulf of Saint Lawrence and Bay of Fundy at around 17.5 cal ka and the Laurentian Channel and Bay of Fundy ice streams ceased to exist (nos. 25 and 185 in Fig. 5e). On the southern Interior Plains, the Maskwa Ice Stream (no. 153 in Fig. 5a-d) might have continued to operate between the Central Alberta Ice Stream (no. 14 in Fig. 5b-e) and the James Lobe (no. 28 in Fig. 5a-f), but further growth of the LIS-CIS ice saddle likely caused ice drained by the James Lobe to become sourced from a more westerly location, with its tributaries cutting across the now inactive Maskwa Ice Stream track (Figs. 5e, 6; see also Fig. 12 in Ross et al., 2009). We reconstruct the shutdown to occur just before 17.5 cal ka, and this might have also been the peak phase of the CIS-LIS saddle, since both the CIS and the western portions of the LIS were at or close to their maximum extents at this time (Jackson et al., 1999; Kennedy et al., 2010; Seguinot et al., 2016).

Other sectors of the LIS saw little change during the early Late Glacial in terms of ice stream activity or the location of the major ice domes and saddles. The ice drainage network was mainly topographically controlled with ice streams located in glacial troughs on the continental shelf that were formed during earlier glacial cycles. The lobes of the southern terrestrial margin underwent oscillations (Clayton and Moran, 1982; Clark, 1994; Dyke et al., 2003; Mickelson and Colgan, 2003; Dyke, 2004) that were earlier described as surging (Clayton et al., 1985; Marshall et al., 1996; Marshall and Clarke, 1997b; Evans et al., 1999; Evans and Rea, 1999; Colgan et al., 2003). The oscillations of the lobe fronts were related to the activity of the ice streams feeding the lobes. This is documented by a series of lithologically distinct till sheets deposited by the lobes, which indicate changes in the source area of the sediments and changes in the flow paths (e.g., Lusardi et al., 2011). However, the complexity of the till stratigraphy of the southern lobes has thus far precluded a broader regional reconstruction that would combine the till stratigraphy with the geomorphological record.

Fig. 6 here (column width)

3.4. Evolution of the ice stream network during the Late Glacial

3.4.1. 17-15.5 cal ka (14-13 ^{14}C ka)

Whereas the early part of the Late Glacial saw little reduction in the ice sheet extent, the pace of the ice margin retreat increased after about 16 cal ka in the terrestrial part of the ice margin (Dyke et al., 2003; Dyke, 2004). As a consequence, the CIS-LIS ice saddle had weakened and separation of the two ice sheets began. According to our reconstruction, the collapse of the ice saddle followed immediately after its maximum phase, which might explain the highly complex succession of ice streams, where, for a brief period during the ice saddle collapse, several ice streams with NW-SE orientation existed before the ice-flow direction turned to the south and later south-west (Margold et al., 2015a). Shortly before 16 cal ka, the Central Alberta and High Plains ice streams (nos. 14 and 15 in Fig. 5b-e) were replaced by "IS2" (no. 152 in Fig. 5f, after Ó Cofaigh et al., 2010) that had its onset zone close to that of the Central Alberta Ice Stream but had a NW-SE orientation instead of N-S (Fig. 5f). This ice stream likely functioned first as a tributary of the James Lobe (Ross et al., 2009, their Fig. 12) and, when that retreated above the junction with IS2, the latter might have functioned as an independent ice stream, before they both switched off at about 15.5 cal ka.

To the north of the CIS-LIS ice saddle, ice was drained in the direction of the present day Mackenzie River. It is, however, unclear how far up-ice the fast ice flow extended, and whether there was an extensive ice stream system changing its trajectory only in its down-ice portion, or whether there was alternating flow between the Mackenzie Ice Stream and the Anderson Ice Stream (no. 2 in Fig. 5f; Kleman and Glasser, 2007; Brown, 2012; Batchelor et al., 2014). Another possibility is that there was a succession of less extensive ice streams operating in a limited distance from the ice margin (see Brown, 2012; Margold et al., 2015a; and Sections 4.3.1 ad 4.3.2. in this study). Based on the sedimentary record in the SW portion of the Amundsen Gulf , Batchelor et al. (2014) inferred that the Anderson Ice Stream could have possibly reached to the shelf north of Cape Bathurst (see Fig. 1 for location) “subsequent to the retreat of the last, Late Wisconsinan Amundsen Gulf ice stream” (Batchelor et al., 2014, p. 140).

In the Great Lakes area, the Port Huron phase advance, one of the major oscillations of the ice margin in this region, is dated to ~15.5 cal ka (Karrow et al., 2000; Larson, 2011). Elsewhere in the ice sheet, there were no major changes.

3.4.2. 15.5-13.9 cal ka (13-12 ^{14}C ka)

Just prior to the collapse of the CIS-LIS saddle, the short-lived IS2, operating on the southern Interior Plains, was replaced by IS 3 (no. 155 in Fig. 5f), which itself stopped operating at about 14.7 ka, at about the same time when the Des Moines Lobe retreated and its streaming ceased. The marked easterly direction of IS 3 indicates that Cordilleran ice was likely still drained onto the Interior Plains at the time of IS 3 operation. With the progressing ice margin retreat, the onset zone of ice streaming on the Interior Plains migrated further north and the ice flow changed into a more southerly direction. The thinning of the ice sheet in this region is evidenced by traces of fast ice flow on high ground, with ice flow unconstrained by topography (ice stream fragments no. 154, 160 in Fig. 5f), succeeded by ice streaming steered by the low local relief (ice streams no. 156, 158 in Fig. 5g).

North of the weakening CIS-LIS saddle, ice streaming moved further south in connection with ice margin retreat and ice streams of E-W orientation, likely active before the saddle formation, were reactivated (no. 145, 175-178 in Fig. 5f, g). Similar to the south-flowing ice streams in Saskatchewan described above (ice stream fragments no. 154, 160 succeeded by ice streams no. 156, 158), evidence exists of earlier fast ice flow at the western margin of the ice sheet, unrestricted by topography, in the form of patches of streamlined terrain on the plateau surfaces of the Cameron Hills (no 145 in Fig. 5g) and the Birch Mountains (no. 148 in Fig. 5f, g; Margold et al., 2015a; Paulen and McClenaghan, 2015; Krabbendam et al., 2016). With the lowering of the ice surface, ice streaming became confined to the broad troughs between the plateaux (nos. 175-178 in Fig. 5f-h).

At the southern margin of the ice sheet, fast ice flow has been reconstructed in the upper St. Lawrence Valley shortly before the separation of the Appalachian Ice Complex from the LIS and its subsequent decay into several remnant ice caps. Ice streaming in a south-western direction formed the last phase of ice flow through the Lake Ontario basin (Fig. 5f; Ross et al., 2006; Sookhan et al., in press). Further down the St. Lawrence Valley, ice was drawn in a north-western direction towards an ice margin in the St. Lawrence estuary (Parent and Occhietti, 1999).

3.4.3. 13.9-12.9 cal ka (12-11 ^{14}C ka)

In between the Mackenzie ice stream system and the Amundsen Gulf Ice Stream, at the far north-western margin of the ice sheet, a series of smaller ice streams switched on in the retreating ice margin, each likely only for decades to a few centuries (nos. 3, 4 in Fig. 5g). Successors of the Mackenzie Trough Ice Stream system operated at the rapidly retreating margin of the ice sheet, first as the Bear Lake Ice Stream (no. 5 in Fig. 5g) and then in another branch, possibly simultaneously with the Bear Lake Ice Stream, as the Fort Simpson Ice Stream (no. 143 in Fig. 5h). However, the pattern of ice streaming in this portion of the ice sheet is still little understood and our reconstruction should be viewed as a hypothesis to be tested (cf. Kleman and Glasser, 2007 and Brown, 2012).

Whereas the operation of the Amundsen Gulf Ice Stream appears to have been stable throughout deglaciation, ice streaming in M'Clure Strait likely ceased after a rapid ice retreat from the strait at some point between 15.2 and 13.5 cal ka (Stokes et al., 2009; Lakeman and England, 2012) only to be reactivated later as an ice stream in the M'Clintock Channel (nos. 19 and 10 in Fig. 5a-h; Clark and Stokes, 2001). An additional ice stream track of similar direction exists on Prince of Wales Island (no. 11 in Fig. 5g). It possibly documents a phase of the streaming flow through the M'Clintock Channel that operated after the cessation of the M'Clure Strait Ice Stream and before the commencement of the flow of the M'Clintock Channel Ice Stream (De Angelis and Kleman, 2005; Stokes et al., 2009). The first ice streams to switch off in the Innuitian Ice Sheet were the Gustaf Adolf Sea and Massey Sound ice streams (nos. 129 and 123 in Fig. 5a-h). We reconstruct their cessation at about 13 cal ka, just prior to the Younger Dryas, in connection with marine transgression into the broad marine channels that the ice streams occupied.

At the southern LIS margin, ice streams in the Great Lakes region ceased operating in connection with the ice margin pulling away from the lake basins. The only exception might have been the Superior Lobe that might have possibly reached further up-ice through the Albany Bay Ice Stream (reconstructed by Hicock, 1988; Hicock et al., 1989), even though little is known about the latter ice stream. In the southwest, the reorientation of the fast ice flow, discussed above, proceeded with the retreating ice margin and the orientation of ice streams changed from N-S to NE-SW. The largest of these was the Red River Lobe that operated in what used to be an onset zone of the Des Moines Lobe (no. 163 in Fig. 5h). The western margin of the ice sheet on the Interior Plains was characterised by fast flowing and rapidly retreating lobes in the wide, shallow troughs between the higher elevated plateaux (nos. 175-178 in Fig. 5g, h; Margold et al., 2015a, b). These ice streams have thus far received minimal attention and recently available high-resolution digital elevation data should permit a closer investigation of the Late Glacial ice dynamics in this region (see Atkinson et al., 2014a, b).

The high ablation along the southern margin of the ice sheet (Ullman et al., 2015) and the resulting retreat of the southern margin are reflected in the migration of the main Keewatin-Labrador ice divide to the north. At the same time, the Keewatin Dome was migrating to the east, following the collapse of the CIS-LIS saddle and the resumption of fast ice flow at the western ice margin.

3.4.4. 12.9-11.5 cal ka (11-10 ^{14}C ka)

During the Younger Dryas (which characterises this time step) ice streaming resumed in the upstream part of the former M'Clure Strait Ice Stream track (M'Clintock Channel Ice Stream, no. 10 in Fig. 5h; Dyke, 2004). Other smaller ice streams operated on Victoria Island at about the time when streaming in M'Clintock Channel commenced (nos. 7, 8 in Fig. 5h; Stokes et al., 2009). On the mainland, in the north-western, western, and south-western portions of the ice sheet, a series of ice streams operated briefly during this period: Kugluktuk (no. 142 in Fig. 5h), Horn (no. 143 in Fig. 5h), Buffalo River (no. 146 in Fig. 5h) and Suggi Lake (no. 161 in Fig. 5h; Margold et al., 2015a, b). Two large lobes came into operation at the south-western portion of the ice margin in the latter part of the Younger Dryas: the Hayes Lobe that drained Hudson ice in northern Manitoba and its southern, smaller contemporary (or possibly predecessor), the Rainy Lobe (nos. 179 and 180, respectively, in Fig. 5j). While several ice streams can be constrained to the Younger Dryas cooling, and many of the large moraine systems are also assigned a Younger Dryas age (see Dyke, 2004), the existing ice retreat chronology indicates an overall ice margin retreat during this period (Dyke et al., 2003;

Dyke, 2004), although some notable advances have also been recorded (Dyke and Prest, 1987a, Miller and Kaufman, 1990; Dyke and Savelle, 2000). It was at this time when the western and south-western ice margin retreated onto the Canadian Shield and the numerous ice streams of the Interior Plains shut down (cf. Fig. 5h and i). In the Great Lakes region, the Marquette phase, dated to about ~ 11.5 ka, marked the last readvance of the Superior Lobe (Larson, 2011), followed by a retreat of the lobe out of the Lake Superior basin.

The ice drainage network remained stable in the northern and north-eastern ice margins (i.e. the Innuitian and the Foxe-Baffin sector of the LIS) during the Late Glacial, even though the ice front likely retreated on the continental shelf in the warm period (Bølling-Allerød) prior to the Younger Dryas (Furze et al., 2018) and might have readvanced, in places, in the subsequent cold stadial conditions (e.g., in the Cumberland Sound; Jennings et al., 1996; Andrews, 1998; Dyke, 2004). We further discuss the climate forcing of ice streaming in Section 4.2.

3.5. Ice streaming in the Early and Middle Holocene

3.5.1. 11.5-10.1 cal ka (10-9 ^{14}C ka)

The northern portion of the ice sheet, covering the CAA, retreated prior to the Younger Dryas in its westernmost part where the Amundsen Gulf Ice Stream and M’Clintock Channel ice streams drained the ice margin retreating through their respective marine channels. By the end of the Younger Dryas, ice streaming from the saddle connecting the IIS to the Greenland Ice Sheet likely ceased (at about 11.5 cal ka), as did streaming into Baffin Bay from Jones Sound. The major ice discharge route in the region, entering Baffin Bay through Lancaster Sound, likely weakened and began its retreat with the post-Younger Dryas warming, but the grounded ice stream in the outer part of Lancaster Sound may have changed into a floating ice shelf already during the late Allerød (Furze et al., 2018). The sea transgressed rapidly through Lancaster Sound and Gulf of Boothia and, at about 10.5 cal ka, these marine channels were free of ice and streaming had ceased (De Angelis and Kleman, 2007). With the changing ice configuration, ice streaming might have briefly occurred in Peel Sound (no. 13 in Fig. 5j; MacLean et al., 2010) and the first of a succession of small, ephemeral ice streams drained ice on Prince of Wales Island (no. 102 in Fig. 5j; De Angelis and Kleman, 2005). The smaller ice streams draining the high relief coasts of Ellesmere and Baffin Islands ceased to exist when ice drew back from the continental shelf, although fast ice flow in individual fjords likely continued. This transition is dated locally to occur ~12-10 cal ka (Briner et al., 2009) and, in the absence of retreat chronologies from individual fjords, we simply assume that the aforementioned ice streams switched off around 11.5 cal ka, after the Younger Dryas (Fig. 5i-j). The Foxe Ice Dome continued to support diminished ice streams in Cumberland Sound and Frobisher Bay, and the Hudson Strait Ice Stream might have also continued operating, albeit with a shorter extent, having vacated the outer parts of Hudson Strait (De Angelis and Kleman, 2007).

Following the end of the Younger Dryas, the ice sheet resumed its rapid retreat and fully withdrew from the Interior Plains onto the hard bed of the Canadian Shield (Dyke, 2004). The pattern of ice streaming along the terrestrial ice margin changed markedly. Instead of numerous, often small, ice streams that existed on the Interior Plains, infrequent but broad, fan-like ice streams occurred on the Canadian Shield. The best examples of these were the Hayes and Rainy lobes, operating during the latter part of the Younger Dryas. South of the Rainy Lobe, the Albany Bay Ice Stream might still have continued to function after the Younger Dryas. In the Québec-Labrador sector, a series of ice streams drained ice to Ungava Bay in the Younger Dryas or post-Younger Dryas time (nos. 17 and 16 in Fig. 5i, j), although the dating of these ice streams is equivocal. The configuration of the ice divides also changed, and the main ice divide connecting the Keewatin and Québec-Labrador domes was succeeded by an intervening dome located above southern Hudson Bay (Dyke and Prest, 1987a, b). The ice divides connecting it to the Keewatin and Québec-Labrador domes lay further north than the preceding Keewatin-Labrador ice divide (cf. Fig. 5h and i).

3.5.2. 10.1-8.9 cal ka (9-8 ^{14}C ka)

After further retreat of the Keewatin sector, a broad ice stream switched on in its western part (no. 6 in Fig. 5j). The Dubawnt Lake Ice Stream has attracted much attention (e.g., Kleman and Borgström, 1996; Stokes and Clark, 2003a, b, 2004; De Angelis and Kleman, 2008; Greenwood and Kleman, 2010; Ó Cofaigh et al., 2013b; Stokes et al., 2013) for its large size in a decaying ice sheet and its proximity to the centre of the Keewatin Ice Dome. The duration of its operation was likely rather short, from several decades to a few centuries (see Stokes and Clark, 2003a; Kleman and Applegate, 2014). To the south, ice streaming commenced in James Bay after the ice margin retreated across the Shield to the vicinity of the bay (no. 33 in Fig. 5j). The initiation of ice streaming here is unclear, but we place it at about 10 cal ka. Subsequently, an ice stream with a southern direction activated to the west of James Bay (no. 165 in Fig. 5k; termed Ekwan River Ice Stream in Margold et al., 2015a, b and Winisk Ice Stream in Veillette et al., 2017). The later phase of ice streaming out of James Bay was associated with the Cochrane readvances (Dyke and Dredge, 1989; Thorleifson and Kristjansson, 1993; Dyke, 2004). Veillette et al. (2017) suggested that the Ekwan River/Winisk Ice Stream continued to operate in the area west of James Bay after ice streaming out the bay itself ceased.

Further northwest, the Keewatin Dome was associated with an ice stream that flowed in a southerly location at about 9.2 cal ka in northern Manitoba (no. 164 in Fig. 5k), which formed a distinct lobe at the ice margin (Dyke and Prest, 1987b). It is unclear whether this fast flow event was conditioned by the brief cold spell around 9.3 ka (Rasmussen et al., 2014), to which the Québec-Labrador sector supposedly also reacted (Ullman et al., 2016), or whether the ice streaming was caused by the interaction between the ice margin and glacial Lake Agassiz, in which the ice front stood. With the continuous retreat of the ice sheet from the marine channels of the central parts of the CAA, local ephemeral ice caps were left on the islands and these were drained by short-lived ice streams, reconstructed on Prince of Wales Island (no. 12 in Fig. 5j) and in western parts of Baffin Island (no. 103 in Fig. 5j; De Angelis and Kleman, 2007).

3.5.3. 8.9 cal ka to final deglaciation in the Middle Holocene (8 ^{14}C ka onwards)

Towards the end of the Early Holocene, only a narrow body of ice connected the Keewatin and Québec-Labrador ice remnants (Fig. 5k). At about 8.5 cal ka, waters of Lake Agassiz-Ojibway broke through (or escaped beneath) this ice dam and the remnant ice caps in Keewatin and Québec-Labrador became separated (Josenhans and Zevenhuizen, 1990; Barber et al., 1999; Dyke et al., 2003; Dyke, 2004; Lajeunesse and St-Onge, 2008). Shortly after, the Keewatin ice cap became separated from the Baffin Ice Cap, a remnant of the Foxe-Baffin Dome (Dyke et al., 2003; Dyke, 2004). The final breakup of the ice sheet produced several smaller ice streams in the region of Hudson Bay (nos. 121, 122, 166 in Fig. 5k; De Angelis and Kleman, 2005; De Angelis, 2007). Elsewhere, ice streaming is documented only in the Baffin Ice Cap, which continued retreating throughout the Holocene, and where a number of smaller ice streams occurred over the time (nos. 106, 107, 118 in Fig. 5k, 118,119 in Fig. 5m, 120 in Fig. 5n; De Angelis and Kleman, 2007; Dyke, 2008).

4. Discussion

4.1. Topographical and geological controls on the ice drainage network

Early in deglaciation, the main controls on the configuration of the LIS ice drainage network were the broad-scale topography and the existence of marine-terminating margins, in combination with the underlying geology (soft or hard beds), as has been noted in previous studies (Margold et al., 2015a; Stokes et al., 2016a). The topographic controls mostly consisted of glacial troughs on the continental shelf (Margold et al., 2015a). These troughs had formed through selective linear erosion (Løken and Hodgson, 1971; Sugden, 1978; Kessler et al., 2008) and are as much a product of fast ice flow as a control on it. Glacial troughs are assumed to develop over multiple glacial periods (Glasser and Bennett, 2004; though see, e.g., Montelli et al., 2017) and it can therefore be expected that they had largely been in place at the beginning of the last glacial period to act as conduits for LIS ice drainage. In the outermost regions that are thought to have only seen ice sheet glaciation a few times during the Quaternary, and where the Pleistocene glacial maximum was reached only during the LGM such as the lower reaches of the Mackenzie Trough (Batchelor et al., 2013a), the glacial overdeepening and the overall glacial modification of the landscape are less distinct although even here the remoulding of the relief by the ice sheet has completely altered the drainage network (Duk-Rodkin and Hughes, 1994). Indeed, broad troughs were also carved by ephemeral ice streams in the continental settings of the Interior Plains or the region of the Great Lakes.

The existence of a marine-terminating ice margin strongly influenced the character of the ice drainage network (Margold et al., 2015a). Fast ice flow was induced by a calving margin and by topographic steering in marine channels or cross-shelf troughs with their weak beds of marine sediments providing a low basal drag (Winsborrow et al., 2010b). Similar to present-day ice sheets, the configuration of ice streams differed between high-relief coasts and coasts with more subdued topography. Large, broad ice streams existed in regions where ice from relatively low-relief landscapes was drained across extensive areas of the continental shelf (e.g., northern shelf of Newfoundland and eastern Labrador shelf, western portion of the Innuitian Ice Sheet). In contrast, ice drainage through high relief coasts led to the formation of smaller, more closely spaced ice streams (e.g., north-eastern Baffin Island, north-western Ellesmere Island). Nevertheless, these ice streams still concentrated ice flow from multiple fjords, indicating a self-organisation of the ice drainage network where the regional ice sheet configuration, surface mass balance and the resulting ice surface slope likely governed the catchment size of individual ice streams (Kessler et al., 2008). While most of the LGM ice streams in the LIS were marine-terminating, a large number of terrestrial ice streams during the Late Glacial terminated in glacial lakes and were thus also exposed to ice calving at their fronts, which may have triggered and helped sustain ice streaming (Andrews, 1973; Cutler et al., 2001; Stokes and Clark, 2004; Margold et al., 2015a).

The role of geology on Laurentide ice drainage has long been recognised: softer rocks of the Interior Plains and the marine sediments on the continental shelf made for weaker beds and were thus instrumental to fast ice flow (Fisher et al., 1985; Hicock et al., 1989; Clark, 1994; Marshall et al., 1996; Licciardi et al., 1998; Margold et al., 2015a; Stokes et al., 2016a). In contrast, more resistant rocks of the Canadian Shield made for harder beds that caused higher basal drag and were thus less favourable to fast ice flow (Clark, 1994; Marshall et al., 1996). These differences in ice bed conditions are clearly reflected in the changing pattern and lower frequency of ice streaming once the ice margin retreated to the Shield first in the SE and S, and subsequently in the SW and W (Fig. 5a-i). However, it has been suggested that increased subglacial water pressure had the potential to weaken the coupling with the bed and allow for fast ice flow even in the regions where the ice sheet had a rigid bed (Clayton et al., 1985; Kamb, 1987; Stokes and Clark, 2003a, b), possibly combined with high ice deformation within a thick temperate ice layer (Krabbendam, 2016). Furthermore, beds made of resistant rocks are subject to substantial glacial erosion when debris-rich basal ice grinds the bedrock surface. The beds progressively become smooth and streamlined, which reduces the friction at the ice-bed interface (Krabbendam et al., 2016). The Canadian Shield, that had repeatedly been glaciated prior to the last LIS, displays extensive regions of areal scouring, which smoothed the bed in the direction of ice flow (Sugden, 1978; Krabbendam et al., 2016). These regions were likely in place even before the last deglaciation, during which they could have provided, in combination with large amounts of meltwater drained to the bed, conditions relatively favourable for fast ice flow. We thus posit that subglacial geology acted as a modulator of fast ice flow and influenced the character of Late Glacial terrestrial ice streams on the Shield (generally broader and less frequently occurring than earlier ice streams operating on the Interior Plains) but, at the same time, we note that the hard beds of the Canadian Shield did not prevent ice streams occurring. Indeed, Stokes et al. (2016a) argued that although underlying geology and topography are important controls on ice streams, their cumulative discharge is strongly linked to ice sheet volume, which, in turn, is influenced by climatically-driven changes in ice sheet mass balance during deglaciation.

4.2. Relationship between ice stream activity and climate change

The most detailed climate record for the northern hemisphere, the Greenland ice cores, indicates a prolonged period of cold climate spanning from ~23 cal ka to the abrupt start of the Bølling warming at 14.6 cal ka (GS-2.1; Rasmussen et al., 2014). The period of 23-14.6 cal ka clearly shows high-frequency oscillations in $\delta^{18}O$ but no climatic peaks that are classified as interstadials. $\delta^{18}O$ data from a marine sediment core in the Gulf of Alaska indicate a gradual warming trend from about 16 cal ka that is followed by an abrupt warming – the Bølling – that is in tune with, and of the same scale, as the Greenland data (Praetorius and Mix, 2014). The Gulf of Alaska climate record then follows the Greenland ice core data closely for the rest of the Late Glacial and into the Holocene (Praetorius and Mix, 2014; Rasmussen et al., 2014). The whole of northern North America, the region that hosted the North American Ice Sheet Complex (NAISC), thus appears to have experienced broadly similar climate trends, although these could have had different intensity depending on the specific location and setting (e.g., continental vs. oceanic regions/sectors of the LIS). We now discuss how broad changes in climate may have influenced ice stream activity.

4.2.1. The LGM and early Late Glacial

The early maximum extent in Atlantic Canada and the Great Lakes region (Johnson et al., 1997; Shaw et al., 2006; Curry and Petras, 2011) coincided with the coldest period in Greenland (27.5-23.5 cal ka; Rasmussen et al., 2014). The subsequent retreat of the south-eastern LIS and the shutdown of some of the ice streams in this region (no. 134 in Fig. 5a, b) might have been triggered either by a slight mid-LGM atmospheric warming (short interstadials GI 2.1, 2.2; Rasmussen et al., 2014) or possibly by an incursion of warmer water to the marine ice front, a mechanism that has been invoked to influence ice stream retreat in both Greenland and Antarctica (e.g., Payne, 2004; Shepherd et al., 2004; Holland et al., 2008; Rignot et al., 2012; Mouginot et al., 2015), but which has only recently been invoked to explain LIS dynamics (Bassis et al., 2017). No seawater temperature data are available for the concerned region but incursions of warmer water have been documented elsewhere in the North Atlantic during OIS 2 (e.g., Rasmussen and Thomsen, 2008). The reduction in the extent of the Appalachian Ice Complex continued in the early Late Glacial with the resulting shut-down of the Notre Dame and Hawke Saddle ice streams (nos. 45 and 169 in Fig. 5a-d) that used to drain Newfoundland ice to the north. Shaw (2006) suggested that increased calving and lowered basal drag due to gradually increasing sea level might have been responsible. In this respect, the Appalachian Ice Complex likely shared a fate similar to that of the other marine-based ice sheet sectors of the Northern Hemisphere that disintegrated at the time of global sea level rise from the LGM low-stand (western portion of the CIS [Clague and James, 2002; Hendy and Cosma, 2008; Clague and Ward, 2011; Margold et al., 2013], Barents Sea Ice Sheet [Elverhøi et al., 1993; Winsborrow et al., 2010a], British-Irish Ice Sheet [Bradwell et al., 2008]). Empirical reconstructions (Clark et al., 2012) and numerical modelling (Patton et al., 2016, 2017) have shown similar pattern of early ice advance and retreat at the maritime western margin of the British-Irish Ice Sheet as the dynamics of the Appalachian Ice Complex discussed here. Elsewhere in the LIS, few changes are observed throughout this time, which appears consistent with the relatively stable climatic conditions.

4.2.2. Heinrich event 1

Numerical modelling studies (Tarasov et al., 2012; Ullman et al., 2015) model high dynamic discharge from the Laurentide Ice Sheet in the period of Heinrich event 1 and after, ca 17-14.5 cal ka. Indeed, the binge-purge model for Heinrich events specifically implicates instabilities in the Hudson Strait Ice Stream related to thermomechanical feedbacks between ice sheet thickness and basal thermal regime (MacAyeal, 1993; Marshall and Clarke, 1997b; Calov et al., 2002). However, we do not find any evidence in the reconstructed ice drainage network for this period that would point to substantial changes in the ice stream network (such as switching-on or off of ice streams adjacent to Hudson Strait), something that might be expected in order to facilitate increased ice discharge. Historically, there have been only a few studies that claim to have identified changes in the terrestrial LIS drainage network and configuration associated with the assumed Heinrich event Hudson Strait surges. Mooers and Lehr (1997) attempted to correlate the advances of the Rainy Lobe in Minnesota with Heinrich Events 2 and 1, and Clark et al. (2000) considered the Gold Cove advance from Labrador to arise from ice stream destabilisation in Hudson Strait via expansion of the warm-based zone (during H-0). Dyke et al. (2002) also interpreted the changes in the ice sheet geometry over Labrador, reconstructed by Veillette et al. (1999), as resulting from post-Heinrich event ice sheet reorganisation. We cannot exclude the possibility that ice streams, such as the Hudson Strait Ice Stream, which has been suggested as the source of the ice rafted debris-carrying icebergs, increased their velocity at the time of Heinrich event 1. Indeed, there is some evidence that might indicate a deepening and widening of its catchment, although this is difficult to date and correlate to a Heinrich event (Parent et al., 1995). However, our glacial geomorphological data is unable to resolve whether ice streams accelerated over relatively short (centennial) time-scales.

4.2.3. Late Glacial climate oscillation: the Bølling-Allerød warming and the Younger Dryas cold period

The substantial and abrupt Bølling warming appears to have broadly coincided with the collapse of the CIS-LIS ice saddle (cf. Rasmussen et al., 2014 and Dyke et al., 2003, Dyke 2004). The warmer climate likely drew the surface mass balance in the saddle region to negative values and this led to the saddle collapse (Gregoire et al., 2012). The separation of the Cordilleran and Laurentide ice sheets and an accompanied rapid retreat of the north-western, western and south-western margin of the LIS caused a rapid reorganisation of the ice-drainage network (Fig. 5f-g). At this time, the western portion of the ice sheet spread onto the Interior Plains and its soft sedimentary rocks aided fast ice flow. That could have been further boosted by lowered basal drag due to excess meltwater lubricating the bed and possibly an over-steepened ice surface profile back to the Keewatin Dome (see Section 5.4.4. in Margold et al., 2015a, for a detailed discussion).

The LIS reaction to the Younger Dryas cooling has not yet been described in detail at the ice sheet scale. Dyke (2004) placed the formation of some of the major moraine systems, which occur mostly closely up-ice from the Shield boundary and well within the LGM LIS extent, in connection with the Younger Dryas, and he also attributed some of the locally-recognised readvances to the Younger Dryas cooling. Nevertheless, the changes in the ice dynamics associated with this most pronounced cooling of the Late Glacial have thus far been little studied other than at a local to regional scale (see, e.g., Stea and Mott, 1989; Lowell et al., 1999; Dyke and Savelle, 2000; Occhietti, 2007; Occhietti et al., 2011; Young et al., 2012). The largest of the ice streams that switched on during the Younger Dryas cooling was the M’Clintock Channel Ice Stream that resumed drainage of Keewatin ice following the cessation of the M’Clure Strait Ice Stream (Stokes et al., 2009). The configuration of the ice drainage network north of the Keewatin Ice Dome was a subject of complex interaction between the Amundsen Gulf, M’Clure/M’Clintock, and Gulf of Boothia/Lancaster Sound ice streams, with these major drainage routes likely competing for adjacent ice catchments (De Angelis and Kleman, 2005; Stokes et al., 2009; Margold et al., 2015a; Fig. 7). It can be assumed that a change towards a more positive mass-balance could have played a role in the activation of the M’Clintock Channel Ice Stream. However, the activity of the M’Clintock Channel Ice Stream only lasted a couple of centuries, and may have shut-down as a consequence of sediment exhaustion rather than for any climate-related reason (Clark and Stokes, 2001). After this, according to the available data, the ice margin in the marine channels of the western CAA started to retreat rapidly, still within the Younger Dryas (Dyke et al., 2003; Stokes et al., 2009; Lakeman et al., in press).

A number of smaller ice streams in the north-western sector of the LIS operated during the Younger Dryas (see Section 3.4.4.). However, it is this sector of the ice sheet where the constraints on the ice retreat chronology are the weakest, and where the ice sheet records one of the highest retreat rates despite the colder climate of the Younger Dryas (Dyke et al., 2003). The connection of these smaller ice streams (Kugluktuk [no. 142 in Fig. 5h], Horn [no. 143 in Fig. 5h], Buffalo River [no. 146 in Fig. 5h] and Suggi Lake [no. 161 in Fig. 5h]) with possible mass balance changes stemming from the temporarily colder climate of the Younger Dryas is therefore only tentative. More intriguing are the two large, broad lobes farther southeast: the Hayes Lobe and the Rainy Lobe (nos. 179 and 180, respectively, in Fig. 5j). These lobes came into existence towards the end of the Younger Dryas and represent the penultimate advance of Hudson ice before its collapse (the last being the Cochrane readvances and the associated ice streaming around James Bay – see Section 3.5.2. and Dyke and Dredge, 1989). The mechanism that drove the advance of the lobes has not yet been fully explained and several processes might be involved: (1) ice flow was reinvigorated as a result of a more positive surface mass balance during the Younger Dryas, (2) rapid ice flow was triggered by a decreased basal friction as a result of large amounts of supraglacial meltwater reaching the bed in a warming climate during the latter part of the Younger Dryas and in the early Holocene – note that this goes against point 1, (3) a portion of the ice sheet was destabilised through water level changes of glacial Lake Agassiz, which abutted a substantial portion of the ice margin in the southwest. Lake Agassiz rose from its Moorhead low-phase to the Upper Campbell Beach, marking the Emerson phase, at about 11.5 cal ka (Björck and Keister, 1983; Clayton, 1983; Bajc et al., 2000; Boyd, 2007). This is within the uncertainty of the timing of the start of the Hayes Lobe activity but it postdates the inferred start of operation of the Rainy Lobe by about 900 years (see Supplementary Data). Points 2 and 3 are not mutually exclusive and we thus deem their combination as a more credible explanation for the activity of the broad ice streams in the SW LIS than the changes in surface mass balance (point 1).

Another region that saw considerable ice stream activity at and after the Younger Dryas is the Labrador Peninsula. Several ice stream flow-sets oriented towards Ungava Bay have been reconstructed there (Veillette et al., 1999; Clark et al., 2000; Jansson et al., 2003). Margold et al. (2015a, b) classify these flow sets into two main generations, with an additional smaller ice stream track at Payne Bay farther northwest (nos. 16, 17, and 188 resp. in Fig. 1). Andrews and MacLean (2003) mention three episodes of ice discharge from Labrador in a northern and north-eastern direction: the first at 11-10.5 ^{14}C ka (12.9-12.4 cal ka), associated with the Heinrich-like H-0 event and likely coeval with an ice advance in Cumberland Sound, the second at 9.9-9.6 ^{14}C ka (11.4-10.9 cal ka) known as the Gold Cove advance and the third, and spatially most limited, at 8.9-8.4 ka (10-9.3 cal ka), known as the Noble Inlet advance (Stravers et al., 1992; Andrews et al., 1995a; Andrews et al., 1995b; Manley, 1996; Kleman et al., 2001; Rashid and Piper, 2007). Moraines formed at the southern margin of the Labrador Dome document a slow-down of the ice retreat or ice margin stabilisation during the Younger Dryas at around 10.3 ka and 9.3 ka (Occhietti, 2007; Ullman et al., 2016). The more dynamic northern section of the Labrador Dome could have readvanced at these times. Alternatively, the advances across Hudson Strait could have been caused by the loss of buttressing effect that the ice stream in the Hudson Strait exerted on Labrador ice. The interaction between the dynamics of ice flow in Hudson Strait and advances across the strait from Ungava is not well understood and requires further research.

In summary, we note that the LIS reaction to the Younger Dryas is still poorly understood, especially at the ice sheet scale. The greatest effect of the temporary cooling was the preservation of the north-eastern portions of the Laurentide and Innuitian ice sheets that, apart from a possible retreat from the outer continental shelf, saw little change until after the Younger Dryas. Nevertheless, we note here that the portrayal of the Innuitian Ice Sheet has possibly gone from “little” ice to “big ice *perhaps too long*” (see section 3.1. and Miller et al., 2002, for a discussion of the LGM ice extent in the Canadian Arctic). Whereas previously it was portrayed as partly unglaciated, even during the LGM (largely due to faulty ^{14}C ages; Dyke and Prest, 1987a; see discussion in Hodgson, 1989), the most recent reconstruction shows the IIS remaining at its LGM position even at the time when substantial retreat occurs elsewhere (cf. Dyke, 2004 and England et al., 2006).

Fig. 7 here (full-page width)

4.2.4. Ice dynamics during the LIS collapse in the Early Holocene

The rapidly retreating ice sheet, which would have been in a highly negative mass balance (Carlson et al., 2008; Carlson et al., 2009), produced several large ice streams during the Early Holocene. The most well-known of these was the Dubawnt Lake Ice Stream, draining the remnant of the Keewatin Ice Dome to the west (no. 6 in Figs. 1, 5j), but others included the Quinn Lake Ice Stream, also in Keewatin, and the Ekwan River (Winisk in Veillette et al., 2017) and James Bay ice streams in the south. It might be assumed that all these ice streams must have had dynamic triggers (such as rapid changes in ice sheet configuration, flotation/calving of the ice front in glacial lakes, or basal de-coupling due to excessive meltwater reaching the ice sheet bed) because ice streaming is unlikely to have been driven by surface mass balance at this time (Ullman et al., 2015). Each of these ice streams terminated in a glacial lake. The role of the glacial lakes as one of the factors triggering fast ice flow has been repeatedly noted for the Dubawnt Lake Ice Stream (Kleman and Borgström, 1996; Stokes and Clark, 2004) and ice surging into Lake Agassiz-Ojibway at this time was mentioned in earlier studies (Dyke and Prest, 1987a; Dyke and Dredge, 1989). The occurrence of glacial lakes therefore appears to be of particular importance for ice streaming in the Early Holocene LIS.

Similarly, the other smaller deglacial ice streams also most probably came into existence as a result of dynamic readjustment to rapidly changing ice geometries and their connection to climate was thus only indirect. The Early Holocene warming caused a rapid ice retreat (mostly in the marine channels) that left isolated ice masses stranded on the islands of the CAA. Small ice streams then switched on in these ice caps, likely in response to the adjusting ice geometry and the removal of buttressing ice in the channels, but their activity only lasted for a short time because these ice caps were in a strong negative mass balance and rapidly downwasted. Nonetheless, these ephemeral ice streams were an integral part of the ice sheet collapse (De Angelis and Kleman, 2007).

4.2.5. Summary

In summary, we note that the first ice streams to switch off were those in the south-eastern marine margin (Fig. 4), which started to retreat already during the LGM. Whether this was due to an incursion of warm waters to the ice front or in reaction to the global sea level rise is as yet unclear. The separation of the LIS from the CIS was completed in the Bølling-Allerød warm interval and the north-western, western and south-western Laurentide margin appears to have continued retreating even in the Younger Dryas (Dyke et al., 2003) despite the fact that some moraine systems (see Dyke, 2004) indicate a possible stabilisation of the ice margin and, locally, where these moraines were overridden (e.g., the Pas Moraine; Dyke and Dredge, 1989; Stokes et al., 2016b), even a readvance of the ice sheet into the previously deglaciated area. Subsequently, in the warm climate of the early Holocene, dynamic triggers related to the abrupt deglaciation are seen as a more likely drivers of ice streaming than ice mass turnover related to the ice sheet mass balance.

4.3. Relationship between ice streaming and ice divides

4.3.1. Onset zones of major Laurentide ice streams

A fundamental question with regards to the overall ice sheet geometry and dynamics is the issue of how far up-ice ice streams propagated, both during the LGM and later in the shrinking ice sheet of the Late Glacial and Early Holocene. We have noted this in connection with the uncertainty about the character of the Mackenzie Trough Ice Stream system (Section 3.4.1), but it applies to most of the large Laurentide ice streams. The implication of this issue for the Keewatin Ice Dome is the question of whether it was connected to the Innuitian Ice Sheet and the Foxe-Baffin Dome by high-elevated ice divides or whether these major ice dispersal centres were divided by a region of low-elevated, fast-flowing ice extending from the Kent Peninsula in the west to the Rae Isthmus in the east, where the major ice streams draining the Keewatin Dome to the northwest, north, northeast, and east, likely had their onset zones (Figs. 5a; 7). Based on a sediment-landform assemblage of a thick till layer and a convergent pattern of drumlins in northern Keewatin, inferred to have been formed under the LGM ice flow directions, Hodder et al. (2016) suggest that zones of rapid ice flow propagated close to the ice divides during the LGM. A large ice catchment and significant up-ice influence has also been reconstructed for the Hudson Strait Ice Stream based on the glacial geomorphology of islands in northern Hudson Bay (Ross et al., 2011). This is similar to the ice drainage network of modern West Antarctic Ice Sheet where tributaries of ice streams that feed into the Ross and Ronne ice shelves reach close to the central ice divide (Rignot et al., 2011), which has, as a consequence of this effective ice drainage, an elevation of about 2000 m lower than the East Antarctic Plateau (Fretwell et al., 2013).

While no high-resolution numerical modelling of the LIS has been undertaken thus far (with cell-size at the order of single km, such as, e.g., Golledge et al. [2012]), the current numerical models of relatively low spatial resolution might still provide some clues about the character of the Laurentide ice drainage, even if they only capture the largest ice streams in a rather rudimentary manner (Stokes and Tarasov, 2010; Tarasov et al., 2012). For example, the results of ensemble N5a mean of Tarasov et al. (2012) indicate the existence of a bifurcation zone in northern Keewatin between M'Clure and a tributary of the Gulf of Boothia ice streams, which has been reconstructed by De Angelis and Kleman (2005) and Margold et al. (2015a, b; here discussed in Section 4.2.3.; see Fig. 7). The same model also indicates an extensive up-ice reach of the Mackenzie Trough and Hudson Strait ice streams (Fig. 8b). We therefore draw the large ice streams draining the Keewatin Dome as extending close to the dome area, with onset zones of the Hudson Strait and Gulf of Boothia ice streams meeting over the Rae Isthmus and with the above mentioned bifurcation zone in northern Keewatin (Figs. 5a-h, 7). On the other hand, our depiction of the ice drainage network possibly underestimates the up-ice extent of the ice streams that drained the Québec-Labrador Dome to the northeast (nos. 167-171 in Fig. 5a-g). While the trunks of these ice streams carved distinct troughs on the continental shelf, their upper reaches were located on the hard beds of the Canadian Shield, which, combined with the expected lower ice-flow velocity in the upper portions of these ice streams, makes their recognition in the glacial landform record difficult. It might be noted that whilst Krabbendam et al. (2016) regard the number and size of the reconstructed hard-bedded palaeo-ice streams as underestimated, there have thus far been few studies from hard-bed areas of the LIS that would provide unequivocal information on ice stream outlines, directions and configuration. Future work that attempts to model, individually, some of the larger ice streams in the LIS, as has been done with success in Antarctica (Golledge and Levy, 2011; Jamieson et al. 2012), would enable a refined understanding of their extent and impact on ice sheet geometry.

A numerical modelling study by Robel and Tziperman (2016), using an idealised model set-up, found that climate warming from the LGM conditions resulted in acceleration of ice streams during the early part of deglaciation. This was caused by an increased driving stress in the ice stream onset zone due to a steepened ice surface profile (because an increase in the surface melting was higher in low elevations than in high elevations farther up-ice). While these modelling results appear consistent with numerical modelling of surface energy balance (Ullman et al., 2015), the identified mechanism needs to be further tested in a more complex simulation that would include glacioisostasy and realistic ice sheet bed topography.

Fig. 8 here (full-page width)

4.3.2. Connection between ice streaming and changing ice sheet geometry

The ice sheet sector for which we reconstruct the most complex evolution of its ice drainage network was the Keewatin sector. This is largely because it was the largest and the least topographically constrained ice-dispersal centre of the ice sheet, and one which was also most affected by the temporary coalescence with the CIS. While our reconstruction, starting at the LGM, pictures a fully-developed Keewatin Dome from its beginning (Fig. 5a), mapped ice stream tracks, other components of the glacial landform record, and the pattern of terrestrial sediment dispersal, indicate that in the build-up phase to the LGM or in the earlier stages of the Wisconsinan LIS, the ice sheet geometry might have been markedly different, with an initial ice dispersal from the Labrador Peninsula, northernmost terrestrial Canada and the CAA (Shilts, 1980; Adshead, 1983; Prest et al., 2000; Kleman et al., 2010; Stokes et al., 2012). In the latter stages of the LGM, the Keewatin Ice Dome migrated west in connection with the build-up of the CIS-LIS saddle (Fig. 9). This is evidenced by the switch-on of the Albertan ice streams. The ice dome migration and growth was likely reflected also in the ice drainage to the north. However, there is large uncertainty about the ice dynamics and the ice extent in that region (see Section 3.2.). In addition, we do not fully understand the timing of the ice saddle build-up and the associated strengthening and migration of the Keewatin Dome; this relates to our uncertainty about the source areas of the Des Moines and James lobes and their relation to the Maskwa Ice Stream at the start of our reconstruction (see Section 3.2., Fig. 6). Dyke and Prest (1987b) portray the Albertan ice streams (in a simplified version) fully operational at about 22 cal ka, but that predates the maximum extent of the CIS and thus a fully grown LIS-CIS saddle by several thousand years. Intriguingly, they draw the Des Moines and James lobe as fed by Hudson ice at that time. For the 17 cal ka map, Dyke and Prest (1987b) draw the Albertan ice streams as inactive and the Des Moines and James lobe fed by Keewatin ice at this time. As stated above, we are uncertain about the source areas of the Des Moines and James lobes early on (and acknowledge that their westward migration was likely) but concur with Ross at al. (2009) that, in the later phase, the Albertan ice streams were fed by the CIS-LIS saddle (from about 20.5 cal ka), as might have also been the Des Moines and James lobes shortly before their switch-off and the collapse of the ice saddle.

Fig. 9 here (column width)

Our empirically-based reconstruction of the LIS's geometry can be compared against the available numerical modelling and glacial isostasy studies. Most modelling studies support the notion of a dominant Keewatin Ice Dome at the LGM: the modelling study of Tarasov et al. (2012; Fig. 8b in this paper), and the glacial isostasy-based reconstructions of K.M. Simon et al. (2014, 2016) and Lambeck et al. (2017; Fig. 8d) indicate ice surface elevations of 3000-3500 m for the Keewatin Dome. An older glacial isostasy-based reconstruction ICE-5G (Fig. 8c), suggests ice surface elevations to reach over 4000 in Keewatin (Peltier, 2004). All these reconstructions display a northwest-southeast elongated Keewatin Dome with a secondary Hudson Dome southeast of it (which in empirical reconstructions appears only in the Late Glacial, as we discuss further below). In contrast, the new isostasy model (ICE-6G) of Peltier et al. (2015; Fig. 8e in this paper) depicts the traditional location of the CIS-LIS saddle as the highest and the most dominant feature of the whole NAISC with the Keewatin Dome forming a high, broad shoulder of the common CIS-LIS dome. While such a configuration is a radical diversion from the picture of the LIS geometry that has crystallised since the abandonment of the single-domed LIS, and its plausibility is contested by several types of empirical data at 21 cal ka (such as ice flow indicators [Ross et al., 2009; Brown, 2012] or other interpretations of the glacial isostasy record [Lambeck et al., 2017]), something similar to this ice sheet geometry (if not absolute elevation) might have been briefly attained about 17 cal ka or shortly after, just before the collapse of the CIS-LIS saddle. For this time, there is evidence of an extremely west-based source areas of the James Lobe (see Fig. 12B, C of Ross et al., 2009; Fig. 5e this study).

With the CIS-LIS saddle collapse and a contraction of the accumulation zone of the Keewatin Dome, the main trunk of long lived ice streams, such as the Amundsen Gulf Ice Stream, moved up-ice. In case of the Mackenzie Trough Ice Stream, the migration up-ice might have involved major lateral migration of the main trunk of the ice stream or, alternatively, there may have been a series of smaller ice streams operating in succession (see Section 3.4.1.). The onset of the west-flowing ice streams on the Interior Plains forced a migration of the Keewatin Dome back to the east (Fig. 5f-g).

Dyke and Prest (1987b) depict a gradual formation of the Hudson Dome from about 12.9 cal ka (11 ^{14}C ka; Fig. 9) by a lowering of the long-lived saddle that existed east of the Keewatin Dome (with its final position defined by the Burntwood-Knife Interlobate Moraine; Dyke and Dredge [1989]) and by a formation of another saddle farther east on the main Keewatin-Labrador divide (final position defined by the Harricana Interlobate Moraine; Dredge and Cowan [1989]). Dredge and Cowan (1989) placed the formation of the major segments of these interlobate moraines that separated the Hudson Dome from the rest of the ice sheet to 11.5-9.5 cal ka (10-8.5 ^{14}C ka). The Hudson Dome sourced several large, if short-lived, ice streams: Red River, Hayes and Rainy lobes (nos. 163, 179 and 180 in Figs. 1, 5h-j). As we discuss in Section 4.2.4., the drivers of these large ice streams are not well understood. What is clear is that these ice streams, even though they could have partially been forced by the more positive mass balance of the ice sheet during the Younger Dryas, must have exhausted the Hudson Dome. They (in combination with higher surface melt on the southern slopes) pushed the main ice divide in the Hudson section northeast to the area over Hudson Bay (Fig. 9). We speculate that such a configuration, with the ice divide and the whole northern slope on the slippery bed of Hudson Bay, and with both the north-eastern and the south-western margins influenced by calving, could not sustain a sufficiently steep ice surface profile, which in the warm post-Younger Dryas climate led to a rapid weakening of the whole Hudson section of the LIS.

Post-Younger Dryas changes in ice sheet geometry were dramatic also in the northern part of the LIS. The Lancaster Sound/Gulf of Boothia ice stream system retreated extremely rapidly and a marine transgression might have reached the head of Gulf of Boothia by 10 cal ka. This deep calving bay must have drawn down the surface of both the Keewatin and Foxe-Baffin domes, which further increased their exposure to the warm post-Younger Dryas climate. The short-lived but massive Dubawnt Lake Ice Stream likely pushed the Keewatin ice divide to the east (Boulton and Clark, 1990 a,b; McMartin and Henderson, 2004). At the same time or shortly after, the Keewatin ice mass, and its connection to the Foxe-Baffin Dome, were drained by increased fast ice flow into Hudson Bay (nos. 121, 122, 166, 174 in Fig. 5j, k), which forced the Keewatin Ice Divide to migrate back to the north-west (McMartin and Henderson, 2004). Increased calving to Lake Agassiz-Ojibway also drained ice from the Keewatin and Hudson ice masses to the south. The Québec-Labrador sector might have been less affected by the dynamic response to the changes in the ice sheet configuration but, even here, the series of Ungava advances (see Section 4.2.3.) must have contributed to the lowering of the ice sheet surface with all its negative consequences in the warm climate of the Early Holocene.

An interesting aspect of the configuration of the LIS during the last glaciation is that a large dome was never positioned over Hudson Bay itself. Rather, we see the development of domes in Keewatin, Québec-Labrador and above the Foxe Basin (Prest, 1969; Shilts, 1980; Dyke and Prest, 1987a, b; Boulton and Clark, 1990a, b; Kleman et al., 2010; Stokes et al., 2012). This is likely due to the influence of the Hudson Strait Ice Stream whose existence was conditioned by the presence of marine sediments on the floor of Hudson Bay (Clark, 1994; Marshall and Clarke, 1997b, Calov et al., 2002).

4.3.3. Relationship between ice streaming and ice divides – Summary

In summary, we note that a picture of large ice streams propagating far into the accumulation zone is emerging for the fully-developed LIS at the LGM and in the early Late Glacial. This is supported by both empirical (Hodder et al., 2016) and numerical modelling studies (Stokes and Tarasov, 2010; Tarasov et al., 2012). The most dynamic sector of the LIS was the Keewatin Dome where a build-up and subsequent collapse of an ice saddle with the CIS caused a succession of ice streams of varied ice flow directions. Uncertainty remains about the prominence of the CIS-LIS saddle at its peak phase, with different glacioisostasy-based models providing different estimates of ice geometry and thickness (cf. Peltier et al., 2015; Lambeck et al. 2017). The LIS retreat in the early Holocene was marked by rapid ice divide migrations and reconfiguration of the ice drainage network. Some of the ephemeral ice streams arose in reaction to calving into the transgressing sea and debuttressing caused by ice retreat from the marine channels of the CAA.

5. Future work

An important component for reconstructing the Late Wisconsinan LIS configuration and ice drainage network is an improved knowledge of the ice sheet extent through time. Arguably, the most sparsely covered region is the NW portion of the Keewatin sector, even though the need for additional dates was highlighted in a paper by Bryson et al. in 1969 (see Stokes, 2017). Here, uncertainty remains about the timing of the maximum stage (Section 3.2.) and the subsequent period (14.7-12.5 cal ka) when some of the highest ice margin retreat rates were reached. It is also a critical area in terms of dating the northern opening of the ice-free corridor. Indeed, improved chronological constraints in this region could, among other issues, contribute to determining the primary source of the Meltwater Pulse 1-A, for which this region has been cited as a candidate (cf. Tarasov and Peltier, 2005; Tarasov and Peltier, 2006; Gregoire et al., 2016; and Carlson and Clark, 2012). Ice sheet chronology could also be improved along most areas of the continental shelf, although here the dating methods face substantial problems, such as the lack of datable organic material (Kaufman et al., 2004; England and Furze, 2008; Lakeman and England, 2013) or the poorly constrained radiocarbon marine reservoir effect (see Section 2.1.).

5.1. Processes and specific issues

In addition, improved understanding of several processes and factors involved in ice streaming is required in order to understand deglaciation dynamics:

- The role of glacial lakes and rising sea level on the flotation of outlet glacier snouts, decreased basal drag along the ice front, and calving, needs more consideration. The issues that connect to this point are, for example, the occurrence of the fan-like ice streams at the south-western LIS margin, abutting Lake Agassiz, or the rapid deglaciation of the Lancaster Sound / Gulf of Boothia ice stream system, where a calving bay might have propagated under extreme rates into the interior of the ice sheet. Model simulations might attempt to assess calving into glacial lakes, which would provide a useful insight into the importance of lacustrine calving in pacing ice sheet retreat.
- The role of subglacial meltwater in decreasing the basal drag and triggering ice streams. This might also involve an improved understanding of what percentage of supraglacial meltwater is drained to the bed and what is the form of the subglacial drainage system (Storrar et al., 2014; Greenwood et al., 2016). Further quantification of melt rates during the LIS deglaciation, in line with the studies of Carlson et al.

(2008, 2009, 2012), and their comparison to present day situation in Greenland and Antarctica may help to understand the ice dynamics of the deglaciating ice sheet.

- Ice streaming on hard beds: it has recently been argued that even hard beds can be conducive to fast ice flow (Eyles and Putkinen, 2014; Krabbendam, 2016; Krabbendam et al., 2016), but fewer ice stream tracks have thus far been mapped in the zone of areal scouring on the Canadian Shield. The region of the Canadian Shield should therefore be scrutinised for evidence of hard-bed ice streams that have not yet been identified.
- Improved knowledge on the extent of ice shelves along the LIS/IIS margin and an assessment of the buttressing effect they provided: Ice shelf buttressing is known to be a key control on the stability of ice streams in Antarctica (Dupont and Alley, 2005; Fürst et al., 2016), but there is a dearth of literature on Laurentide ice shelves. A possible occurrence of an ice shelf off the Hudson Strait has been discussed in connection with Heinrich events (Hulbe, 1997; Hulbe et al., 2004; Alvarez-Solas et al., 2010; Marcott et al., 2011), but these studies did not consider a full scale ice shelf in Baffin Bay due to earlier views that have seen the Greenland Ice Sheet as not reaching the continental shelf edge at the LGM. With the new information on the LGM extent of the Greenland Ice Sheet (Ó Cofaigh et al., 2013a; Dowdeswell et al., 2014; Hogan et al., 2016; Slabon et al., 2016), the possibility of a large ice shelf in Baffin Bay ought to be revisited. The existence of an ice shelf covering Baffin Bay would have implications for the ice surface profile of the Foxe-Baffin Dome (and of the Greenland Ice Sheet as well) due to the buttressing effect that the ice shelf covering the bay would provide. Ice shelves have also been suggested to have formed adjacent to the LIS and IIS in the Arctic Ocean (Jakobsson et al., 2014), although their LGM extent is uncertain and is assumed to have been much more limited than that of

earlier glacial stages, e.g. MIS 6 (Engels et al., 2008; Jakobsson et al., 2016). Ice shelves have also been reported for the ice retreat phase through the channels of the CAA (Hodgson and Vincent, 1984; Hodgson, 1994; Furze et al., 2018) and both the LGM ice shelves in the Arctic Ocean and the more limited deglacial ice shelves in the CAA would have provided buttressing effect for the ice streams that fed them with ice.

- Changes in the general circulation pattern affect the distribution of precipitation, and thus the mass balance of the ice sheet: studies investigating the interaction between palaeo-ice sheets and the climate system (e.g., Löfverström et al., 2014) are needed to improve our understanding of palaeoclimate, which will in turn help us to disentangle the resulting ice sheet dynamics.
- Incursion of warm waters to the marine ice front: the Appalachian Ice Complex started to retreat during the cold LGM climate and thus the possibility of ocean forcing of its retreat ought to be investigated. Moreover, Bassis et al. (2017) suggested the incursion of warm waters to the front of the Hudson Strait Ice Stream to be responsible for the Heinrich events – if this is to resolve the long-debated issue, more data on water temperature from the Labrador Sea are needed.

5.2. Numerical modelling

An ever growing body of research is directed at numerical modelling of ice sheets and these models continue to advance in terms of their resolution and ability to capture key processes (see Stokes et al., 2015). There is thus a growing importance to efforts that attempt to reconcile numerical modelling results with available empirical and proxy data (Stokes et al., 2015; see also Seguinot et al., 2016, and Patton et al., 2016, Patton et al., 2017). To this end, our reconstruction can serve as a useful test for higher order numerical models that can simulate ice streaming within the LIS. A study comparing empirically reconstructed Laurentide ice streams with those produced by a numerical ice sheet model of the LIS has been carried out by Stokes and Tarasov (2010), but our review presents a significantly higher number of ice streams than known in 2010 and also their timing of operation, which was largely unknown at the time of that study. As a consequence, numerical models with higher spatial resolution will be required to simulate the ice sheet dynamics in order to match the detail of the empirical information that we present here. In addition, numerical models of the surface energy balance, predicting the amount of ice to be drained as a dynamic discharge to maintain the ice sheet geometry, can now be tested against empirical reconstructions of dynamic discharge through time. At present, there is a significant disagreement between these two approaches early during the deglacial cycle (cf. Ullman et al., 2015, and Stokes et al., 2016a).

5.3. Individual ice stream systems

In addition to the points above, future studies should seek a better understanding of individual ice stream systems:

- Hudson Strait Ice Stream, for its connection to the Heinrich events. The mechanism that released the armadas of icebergs into the North Atlantic is still debated and may include an accelerated flow of the Hudson Strait Ice Stream, either due to a change in the basal temperature regime in the area of Hudson Bay (MacAyeal, 1993; Alley and MacAyeal, 1994; Calov et al., 2002) or due to an incursion of warm water to its ice front (Bassis et al., 2017). Another question is the degree of activity of the Hudson Strait Ice Stream between the individual Heinrich events: was it an ice stream with a steady ice discharge and if yes, how did its discharge compare to the other large Laurentide ice streams?

- Mackenzie Trough ice stream system: it remains unresolved whether it operated as an extensive ice stream system with the downstream section migrating laterally across the region (from the Mackenzie Trough to the Anderson ice stream track and back), or whether a collection of smaller, time-transgressive ice streams operated within a limited distance from the ice margin.

- The large Late Glacial lobes of the south-western margin (Hayes and Rainy lobes) need to be explained within the context of the retreating LIS: what were their drivers, how long did they operate, and how did they impact the ice sheet mass balance and specifically the stability of the Hudson Dome?

These and other Laurentide ice streams might be targeted by higher resolution modelling studies that will (i) examine their influence on the ice sheet configuration and mass balance, and (ii) determine the causes of ice stream switch on and shut-down (e.g., Jamieson et al., 2012), which will help to understand and predict the behaviour of modern ice streams.

5.4. LIS configuration and response to Late Glacial climate fluctuations

Questions remain not only about the LIS ice drainage but also about the ice sheet configuration. The build-up, the role at its peak, and the collapse of the CIS-LIS ice saddle is still not properly understood (cf. Peltier, 2004; Peltier et al., 2015; Lambeck et al., 2017). While it has been suggested that the early ice-dispersal centres lay in the east (Dyke et al., 2002; Kleman et al., 2010) and the ice sheet growth in the west followed later on, the exact timing of this is not well resolved.

Finally, the LIS response to the Late Glacial climate fluctuations needs to be better understood. This applies both to the peak retreat rates during the Bølling warming and to the reaction of the ice sheet to the subsequent Younger Dryas cooling. While the reaction of the Fennoscandian Ice Sheet is reasonably well understood, with a major readvance in the southwest and ice margin stabilisation along the rest of the ice sheet perimeter (e.g., Mangerud et al., 2016), the behaviour of the LIS during this time is little known, despite the fact that some large moraine systems have been tentatively connected with the Younger Dryas (Dyke, 2004). Occhietti (2007) reviewed this issue for the southern margin of the Québec-Labrador sector, but few other regions have received similar attention. Some of the large moraines that have been described by Dyke (2004) as of Younger Dryas age are known to have been overridden (such as the Pas moraine in Manitoba), yet the nature of these readvances, which occurred in the context of a continued ice margin retreat elsewhere, is not known.

6. Conclusions

This paper represents the first attempt to reconstruct the transient evolution of ice streams in the LIS from the LGM throughout the deglaciation, together with a series of palaeogeographic maps. The LIS (including the Innuitian Ice Sheet) had an ice drainage network that resembled the organisation of ice streams in the modern ice sheets of Antarctica and Greenland. At the LGM, the majority of the ice streams were marine terminating and topographically-controlled and many of these continued to function late into the deglaciation. The terrestrial ice margin in the south was drained by several large ice streams that fed extensive ice lobes protruding from the ice margin and which were only loosely constrained by the underlying topography. These ice streams have no present-day analogue. The only marine sector that underwent an early retreat was the south-eastern portion of the ice sheet in

Atlantic Canada and Gulf of Maine, where the marine ice margin likely started retreating during the LGM. Ice drainage towards the terrestrial ice margin in the west and south was more transient during the LGM and in the early stages of the deglaciation. This was largely due to a combination of the build-up and decay of the CIS-LIS ice saddle, with the associated changes in the ice sheet geometry, and the low-relief, soft-bedded topography that allowed dynamic changes in the ice drainage network in connection with the changing ice sheet configuration. The CIS-LIS ice saddle likely reached its maximum phase after the LGM, at about 17 cal ka, and started to weaken within a few hundreds of years. The late peak phase of the CIS-LIS saddle is reflected in the evolution of the ice drainage in the south-western portion of the ice sheet, where the source areas of the ice streams draining towards the southern margin gradually shifted west, even at the time when the southern margin was already undergoing rapid retreat.

A rapid reorganisation in the configuration of the ice drainage network ensued after the saddle collapse, with new ice streams draining ice towards the opening corridor between the LIS and the CIS (as of ~14 cal ka). These ice streams were sourced from the Keewatin Dome and they probably contributed to the lowering of the saddles in the main Keewatin-Labrador ice divide, thus establishing an independent ice dispersal centre – the Hudson Dome – by ~ 12 cal ka. In contrast, the ice drainage network towards the marine ice margins in the northern Labrador Sea, in Baffin Bay, and in the Beaufort Sea appears to have remained stable throughout most of the Late Glacial. With the possible exception of the retreat from the outer continental shelf, this portion of the ice sheet margin probably survived largely unchanged until after the Younger Dryas. This, however, is in contrast to the north-western extremity of the ice sheet: ice might have only reached the area of the present day Mackenzie delta after the LGM and retreated soon afterwards. A considerable retreat prior to the

Younger Dryas occurred also in Amundsen Gulf, M'Clure Strait, and on intervening Banks Island.

The Younger Dryas, a period of pronounced cooling during the Late Glacial, clearly influenced the dynamics of some ice streams during deglaciation. We tentatively ascribe the switching-on of some ice streams, most importantly the M'Clintock Ice Stream, to the effects of the Younger Dryas. For other large ice streams, whose timing partially overlaps with the Younger Dryas, the drivers are little understood and ice-dynamic causes of the fast ice flow, rather than effects of climate and surface mass balance are viewed as more likely. Following the Younger Dryas, the ice sheet retreated onto the Canadian Shield. The change in the character of its bed (from weak, fast-ice-flow-conducive to more resistant), together with the shift to sharply negative mass balance, brought about a different pattern of the ice drainage network. Ice streams became less frequent, broader, and probably only operated for a relatively brief period of time. The final ice sheet collapse that split it into remnant ice domes was characterised by the occurrence of small, ephemeral deglacial ice streams. Indeed, the activity of ice streams is clearly associated with the migration and eventual demise of some of the major divides and domes of the LIS. Ice stream activity appears to have influenced the migration and eventual demise of the Keewatin Dome and changes in ice stream trajectory in the south-western sector of the ice sheet can be explained by the build-up and collapse of the ice saddle between the LIS and CIS.

Our empirically-based reconstruction acts as a useful template for the numerical modelling community to test the results of their simulations at the timescales of thousands of years. This is important for realistic simulations of the ice-sheet climate interactions during the last deglaciation, as well as for numerical modelling studies attempting to predict the long-term evolution of modern ice sheets. In addition, future work might fruitfully target issues such as ice streaming in hard-bed areas, the influence of subglacial meltwater and calving on fast ice flow, the extent and role of ice shelves, and the changes in the general circulation pattern, which affected the distribution of precipitation and ice sheet mass balance. Topics that require continued efforts are an improved ice margin chronology during deglaciation and the reaction of the LIS to the Late Glacial climate fluctuations. A better understanding of the behaviour of the ephemeral ice sheets of the Northern Hemisphere is important for predicting the processes that might affect the modern ice sheets in the warming world.

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Figure captions

Fig. 1. LGM ice extent of the Laurentide Ice Sheet (including the Innuitian Ice Sheet) with all its identified ice streams (in blue shading with dashed orange lines marking the ice flow direction). See Table 1 for the names of ice stream numbers (after Margold et al., 2015a, b). The evolution of the ice sheet throughout the last deglaciation, with associated changes in the ice drainage network is drawn in Fig. 5. Abbreviations: AB – Alberta; BF – Bay of Fundy; CB – Cape Bathurst; DI – Devon Island; EI – Ellesmere Island; MB – Manitoba; MI – Magdalen Islands; NB – New Brunswick; PCI – Prince Charles Island; SK – Saskatchewan. This figure is modified from Stokes et al. (2016a). Ice flow velocity for the modern Greenland Ice Sheet is reproduced from the data released by Joughin et al. (2010a, b).

Fig. 2. Ice retreat pattern of the Laurentide Ice Sheet reconstructed by Dyke et al. (2003) drawn in red, with our updated LGM extent in bolder red based on new evidence that the ice sheet extended to the shelf edge (Briner et al., 2006; England et al., 2006; Shaw et al., 2006; Li et al., 2011; Lakeman and England, 2012; Jakobsson et al., 2014; Brouard and Lajeunesse, 2017) and our inference in areas that have yet to be scrutinised. Black dots mark the

distribution of the ^{14}C data on which the ice margin chronology is based. Modified from Stokes et al. (2016a).

Fig. 3. Schematic illustration of the dating of ice stream activity from Stokes et al. (2016a) based on the mapped ice stream tracks (Margold et al., 2015b) and the ice margin chronology of Dyke et al. (2003). The bedform imprint, indicative of fast ice flow in the centre of the figure, is in grey. Isochrones of Dyke et al. (2003) are in orange and the individual ^{14}C ages, on which the ice retreat chronology is based, are in black; note the time-lag between the two. Stokes et al. (2016a) derived maximum, best-estimate, and minimum time of operation for each ice stream – see Section 2.2 for the description of dating methodology.

Fig. 4. Timing of operation of individual ice streams: maximum, best-estimate, and minimum duration of operation from Stokes et al. (2016a). Note that, in some cases, the earliest time of ice stream shut-down precedes the latest time of its switch-on (this happens for smaller ice stream tracks in regions with high reconstructed ice margin retreat rate); this is depicted by the category “Minimum inversed”. In these cases, Stokes et al (2016a) assign each ice stream a minimum time of operation of 100 years.

Fig. 5. Evolution of ice sheet geometry and ice stream drainage network during deglaciation of the Laurentide Ice Sheet. This new reconstruction is based on the ice stream inventory of Margold et al. (2015a, b) and the timing of ice stream activity from Stokes et al. (2016a). The locations of the 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 kyr are shown in grey and those that switched on during the subsequent 1 kyr are shown in dark blue with numbers in red (see Table 1 to cross reference ice stream numbers to their names). Ice sheet geometry

(ice divides, domes [K – Keewatin, Q-L – Québec-Labrador, F-B – Foxe-Baffin], and saddles [S]) are drawn in dark blue. We expect the marine terminating ice streams to have reached the highest ice flow velocities at the ice front whereas terrestrially terminating ice streams were likely fronted by lobes of slower ice and high ablation rates. Note that some of the ice stream tracks drawn do not fit the depicted ice sheet extent and configuration (e.g. the Dubawnt Lake Ice Stream [no. 6 in panel j]) because there was often a substantial ice retreat and reconfiguration in-between the isochrones and some of the pictured ice streams operated only briefly in-between the two isochrones drawn. White numbered question marks indicate areas where high uncertainty remains regarding ice extents and dynamics: 1 (panels a-d) the timing of maximum ice extent in the Mackenzie Delta (see section 3.2); 2 (panels a-d) the location of the source areas of the Des Moines and James lobes (see section 3.2); 3 (panels a-d) ice flow directions in the basins of the Great Lakes (see section 3.2); 4 (panels e-i) extent and timing of operation of the Albany Bay Ice Stream and its connection to the Lake Superior Lobe (see section 3.4.3.); (5) timing of the Remnant Dubawnt Ice Stream corridor (no. 174, see Fig. 4 and Supplementary Data). Abbreviations in panel (a): SIL – Southern Indian Lake; LW – Lake Winnipeg.

Fig. 6. Schematic depiction of the CIS-LIS saddle area evolution and associated ice streaming. Simplified maps with ice sheet outlines and reconstructed ice stream tracks are drawn in the panels on the right. (a) Ice free corridor exists between the two ice sheets prior to the LGM; ice streams occur at the W LIS margin located on the Interior Plains. (b) Growth phase of the CIS-LIS saddle during the LGM: ice streaming in the western direction at the LIS margin ceases and is succeeded by larger ice streams draining ice to the south; little or no Cordilleran ice is drained across the main ridge of the Rocky Mountains. (c) Peak phase of the CIS-LIS saddle during the Late LGM: Cordilleran ice is drained across the main ridge of

the Rocky Mountains and is deflected south by the LIS margin shortly east of the mountain front; Rocky Mountain Foothills Ice Stream of the CIS forms a tributary of the High Plains Ice Stream draining ice from the saddle area and the Keewatin Dome. (d) CIS-LIS saddle collapse phase around 15.5 cal ka: as climate warms and ice stops being drained across the main ridge of the Rocky Mountains, the ice saddle collapses. Equilibrium line retreats to the higher portions of the Keewatin Dome and a rapid succession of ice streams occurs at the W LIS margin. For a brief period, ice is still drained from the saddle area by ice streams of southeasterly direction, succeeded by ice streams of southerly and southwesterly directions. (e) Ice free corridor is re-established by < 14 cal ka and a number of ice streams operate in the W LIS margin on the soft beds of the Interior Plains. (f) LIS ice margin retreats to the hard beds of the Canadian Shield (>11.5 cal ka), which leads to the change in the character of ice streaming: ice streams become less frequent but broader.

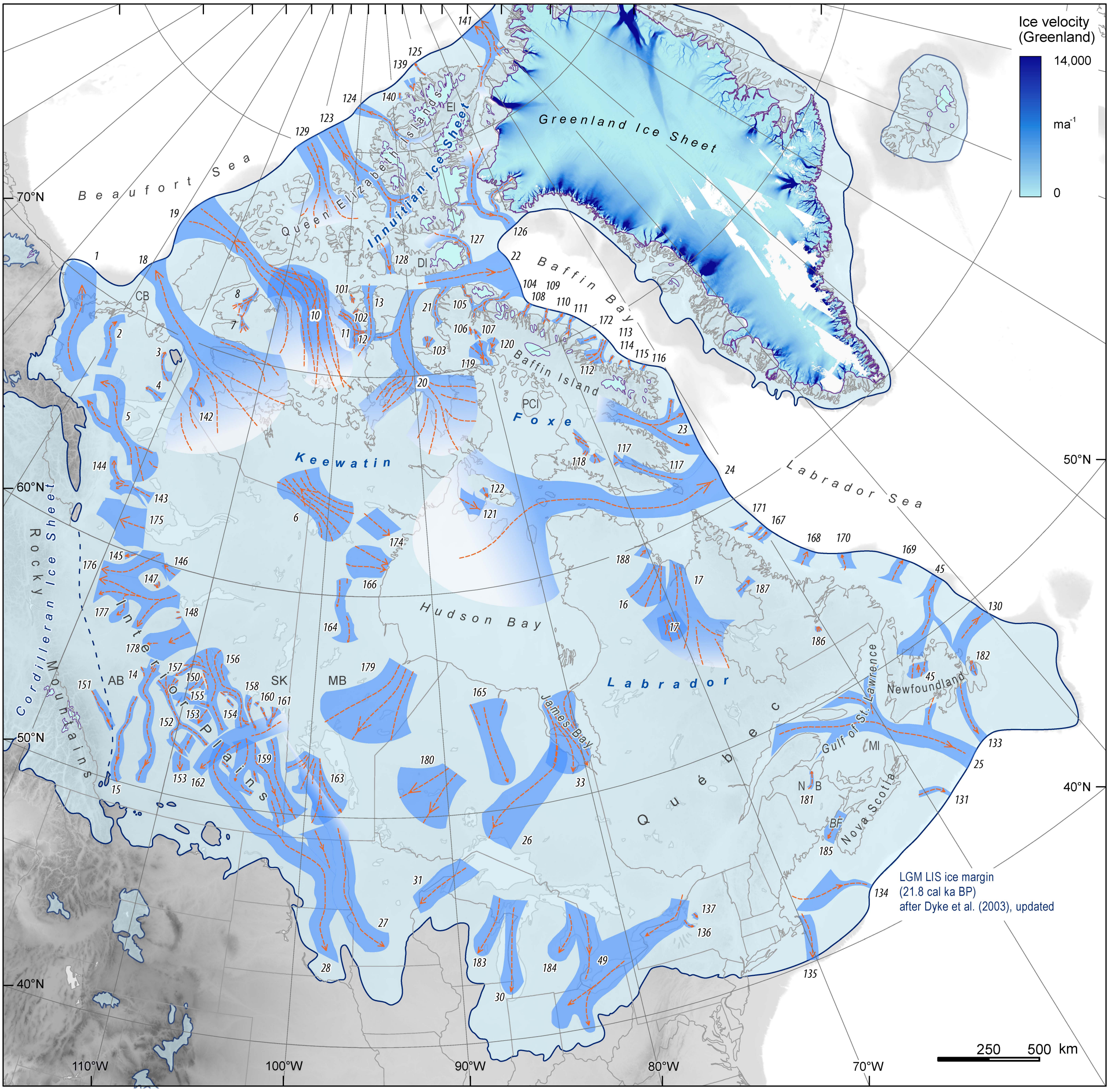
Fig. 7. Comparison of the reconstructed/inferred ice stream onset zones north of the Keewatin Dome with numerical modelling results for the ensemble mean (N5a) of Tarasov et al. (2012; pictured at 20 cal ka). Note that the bifurcation of ice between the Gulf of Boothia and M’Clure Strait ice streams, discussed by Angelis and Kleman (2005), is well reproduced in the numerical simulation. An onset zone up-ice of the bifurcation area reaches close to the Keewatin Dome (K) in the modelling; the possibility of this far-reaching tributary is supported by field evidence recently reported by Hodder et al. (2016). Note also that the onset zones of the Hudson and Mackenzie ice stream systems reach close to the Keewatin Dome in the modelling data.

Fig. 8. Comparison of LGM LIS geometry: (a) empirical reconstruction (this study), (b) numerical ice sheet modelling (ensemble mean (N5a) of Tarasov et al., 2012; pictured at 20

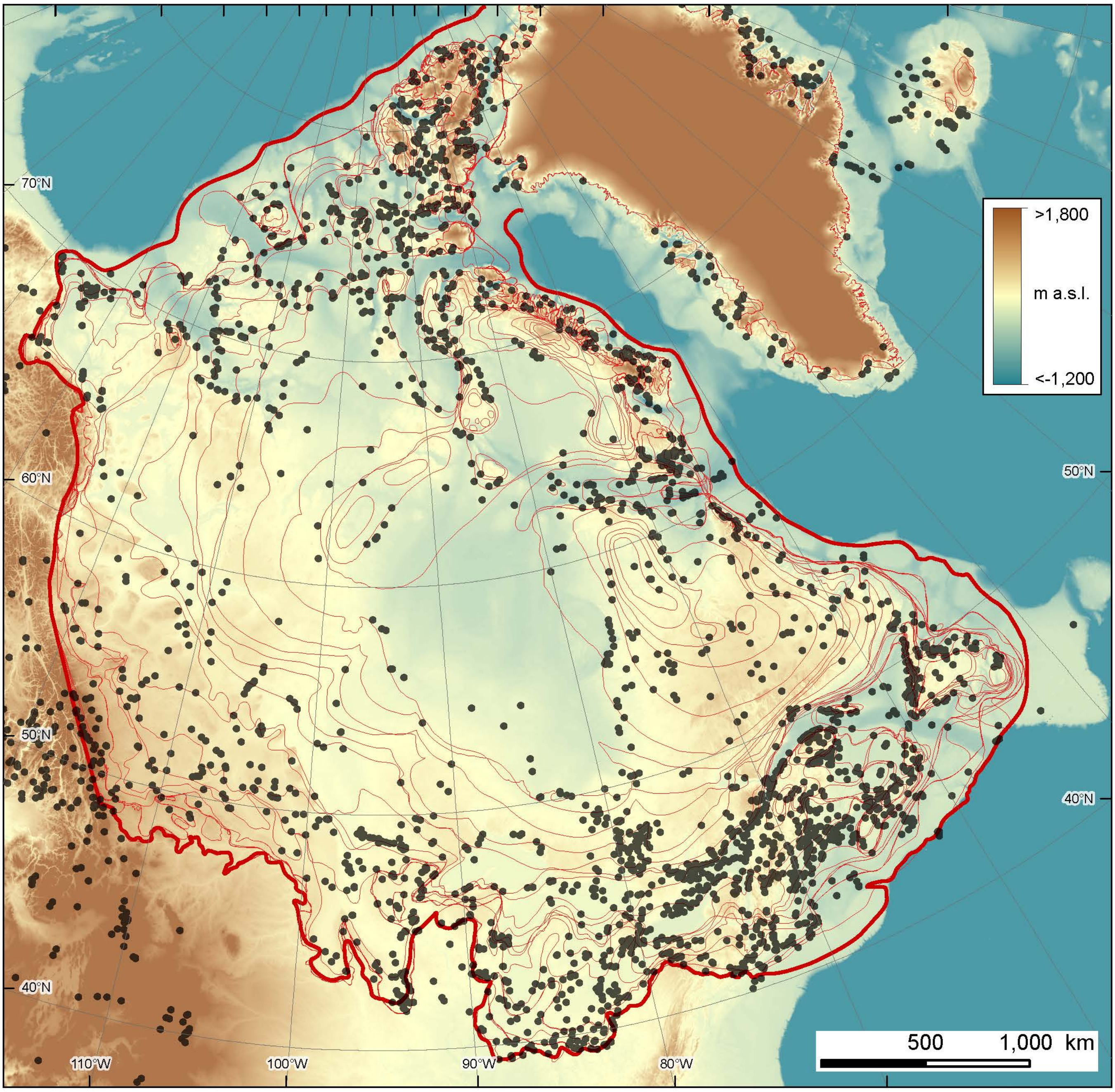
cal ka; colour displays ice velocity; ice elevation in contours of 100 m interval); (c) glacioisostasy-based reconstruction ICE-5G (Peltier et al., 2004); (d) glacioisostasy-based reconstruction of Lambeck et al. (2017); (e) glacial isostasy-based reconstruction ICE-6G (Peltier et al., 2015). Notation of the main ice domes is the same as in Fig. 5. Note the main differences: while the empirical study only envisages the Hudson Dome to exist late in the deglaciation (Fig. 5i), both the numerical modelling in panel b and glacioisostasy reconstructions ICE-5G and Lambeck et al. (2017) in panels c and d infer the existence of the Hudson Dome at the LGM time. In all cases, the Hudson Dome is located south of Hudson Bay, not directly above it. Glacioisostasy reconstruction ICE-6G departs radically from the other reconstructions – see Section 4.3.2. for a more detailed discussion.

Fig. 9. Evolution of ice sheet dome geometry from the LGM throughout deglaciation (updated from Dyke and Prest, 1987b). Major ice domes (Keewatin [K], Québec-Labrador [Q-L], and Foxe-Baffin [F-B] are drawn as ellipses during their peak time. Disappearance of individual ice divides and domes is noted.

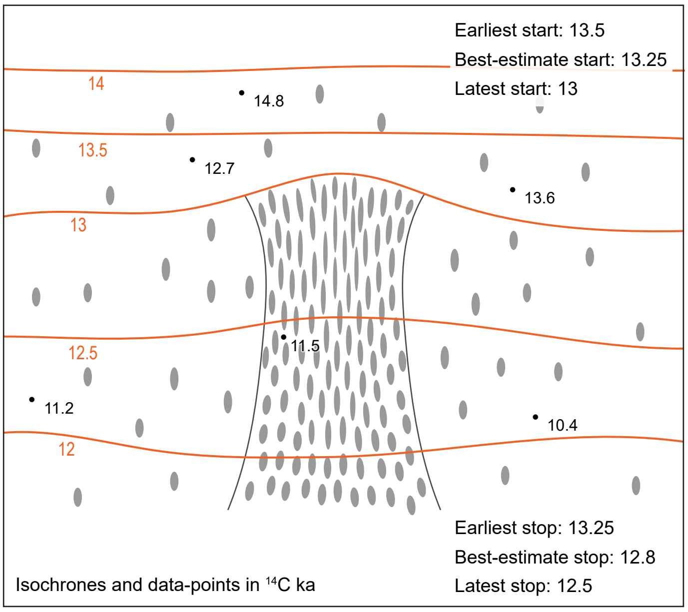
Table 1. Cross referencing between ice stream identification numbers and names (after Margold et al., 2015b).

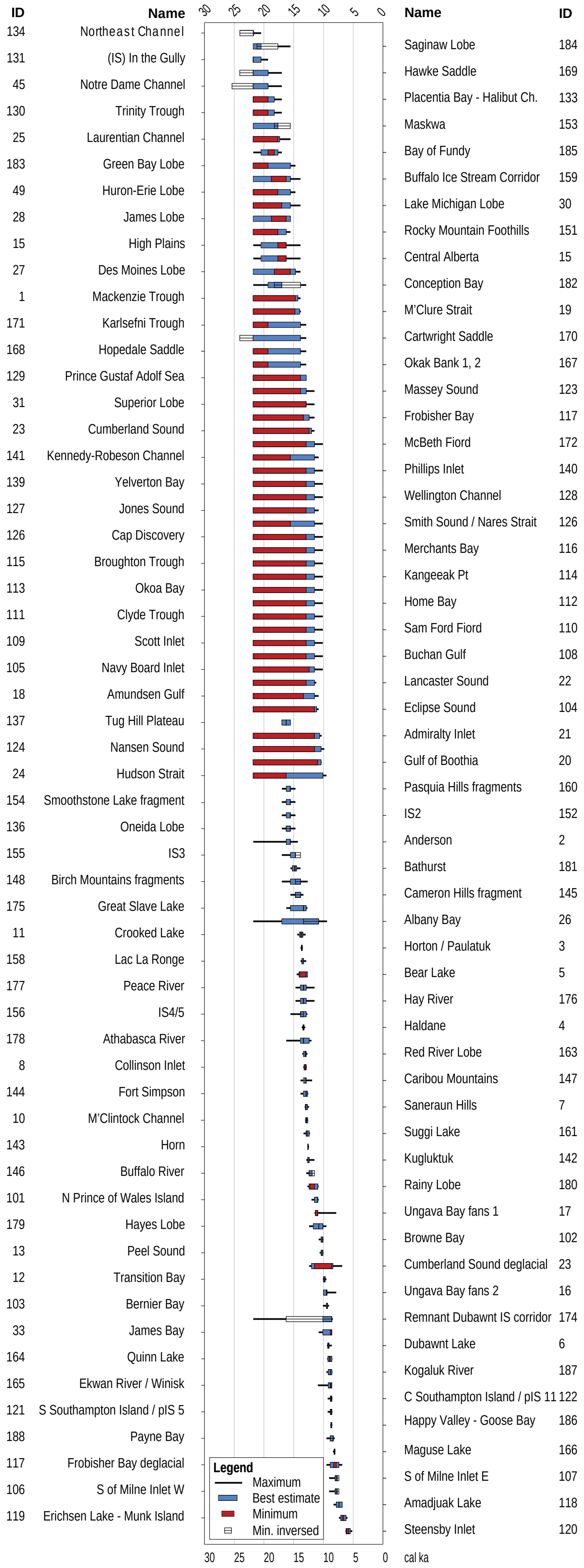


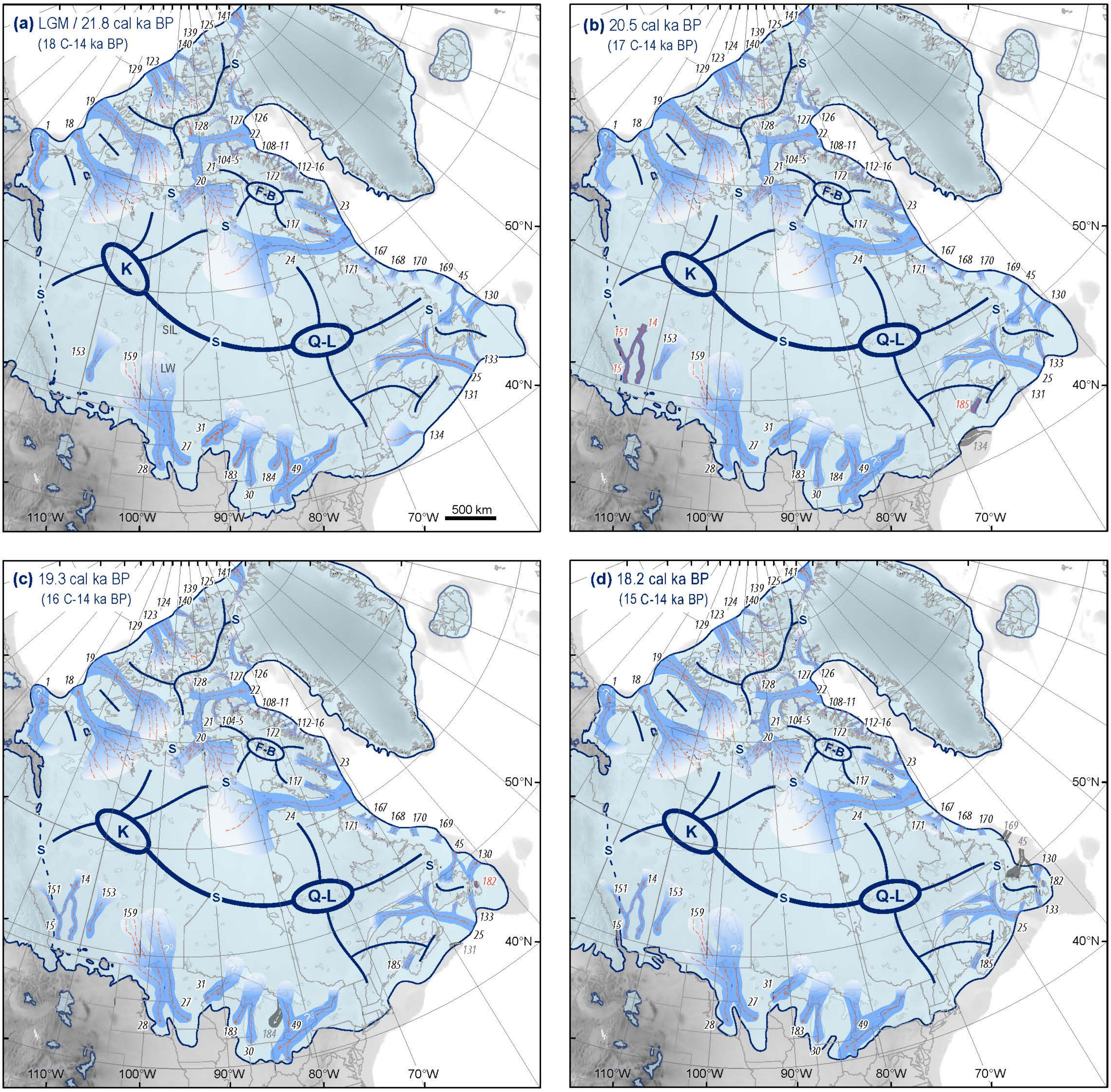
Ice velocity (Greenland)
14,000
ma-1
0
Greenland Ice Sheet
Innuitian Ice Sheet
Queen Elizabeth Islands
Beaufort Sea
Baffin Bay
Baffin Island
Foxe
Keewatin
Labrador Sea
Hudson Bay
Labrador
James Bay
Québec
Gulf of St. Lawrence
Newfoundland
Nova Scotia
Interior Plains
Rocky Mountains
Cordilleran Ice Sheet
AB
SK
MB
CB
EI
DI
PCI
NB
MI
BF
70°N
60°N
50°N
40°N
110°W
100°W
90°W
80°W
70°W
250 500 km
LGM LIS ice margin
(21.8 cal ka BP)
after Dyke et al. (2003), updated



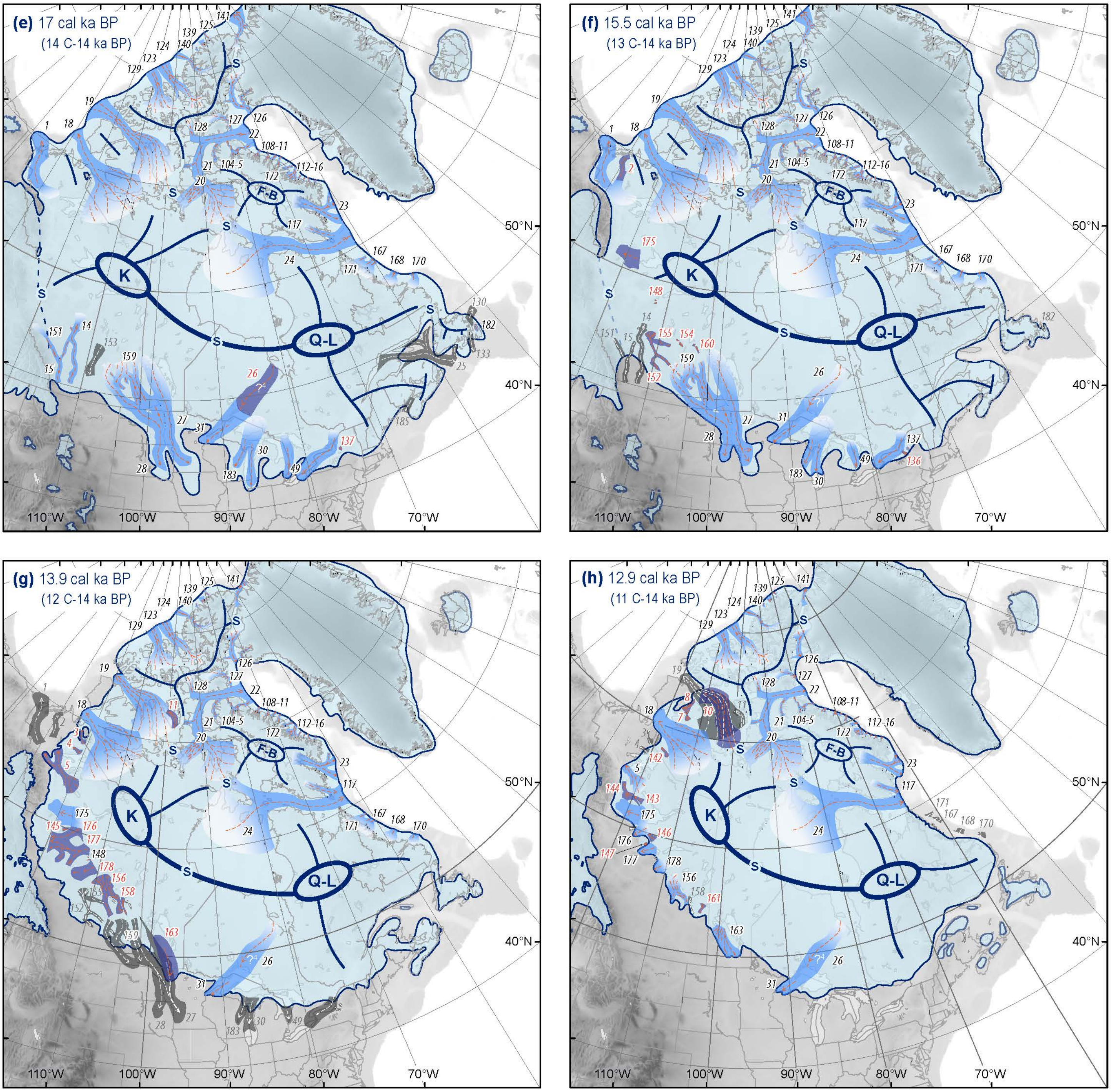
70°N
60°N
50°N
40°N
50°N
40°N
110°W
100°W
90°W
80°W
>1,800
m a.s.l.
<-1,200
500
1,000 km

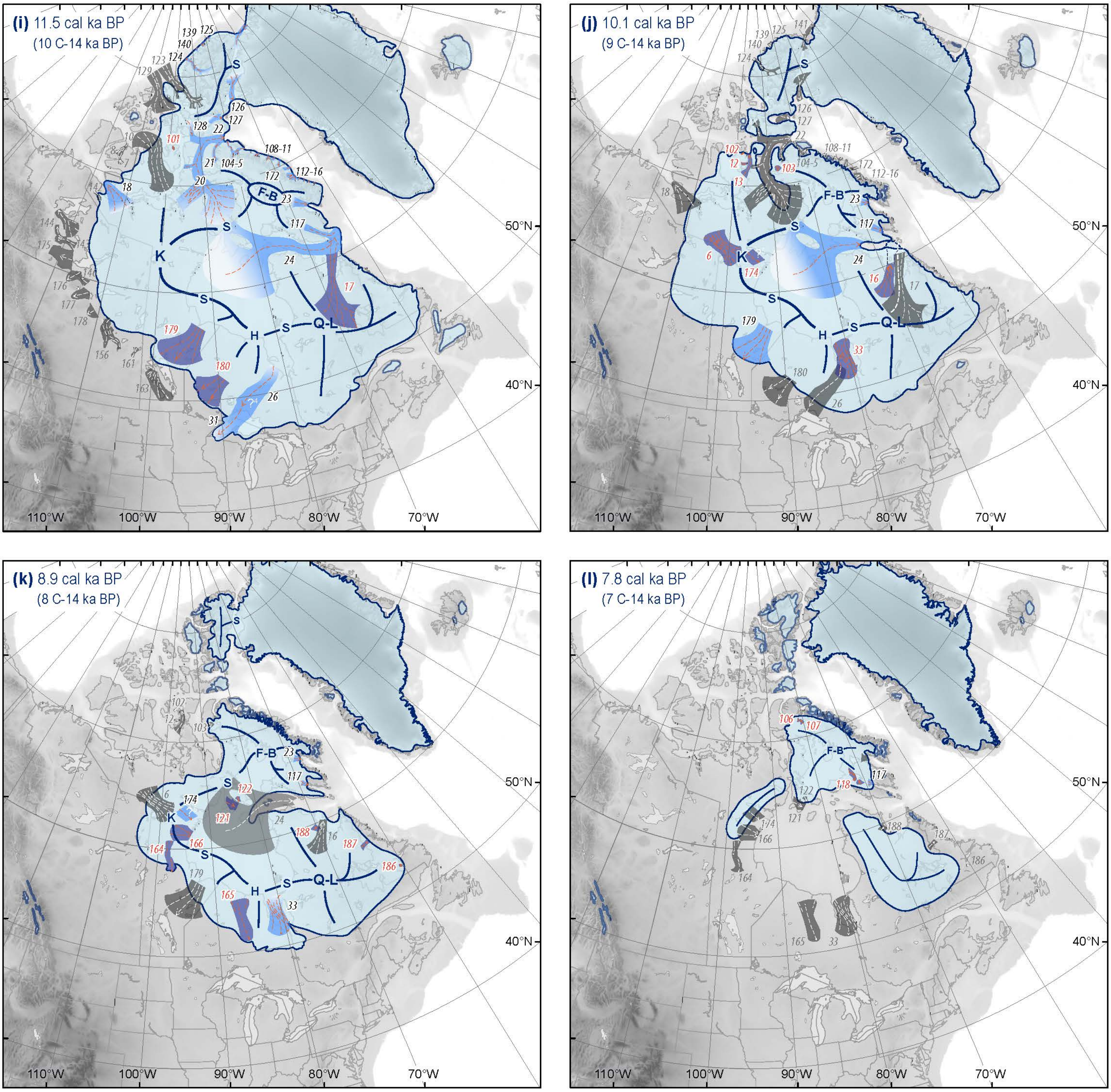

Earliest start: 13.5
Best-estimate start: 13.25
Latest start: 13
14
14.8
13.5
12.7
13.6
13
12.5
11.5
11.2
10.4
12
Earliest stop: 13.25
Best-estimate stop: 12.8
Latest stop: 12.5
Isochrones and data-points in ¹⁴C ka

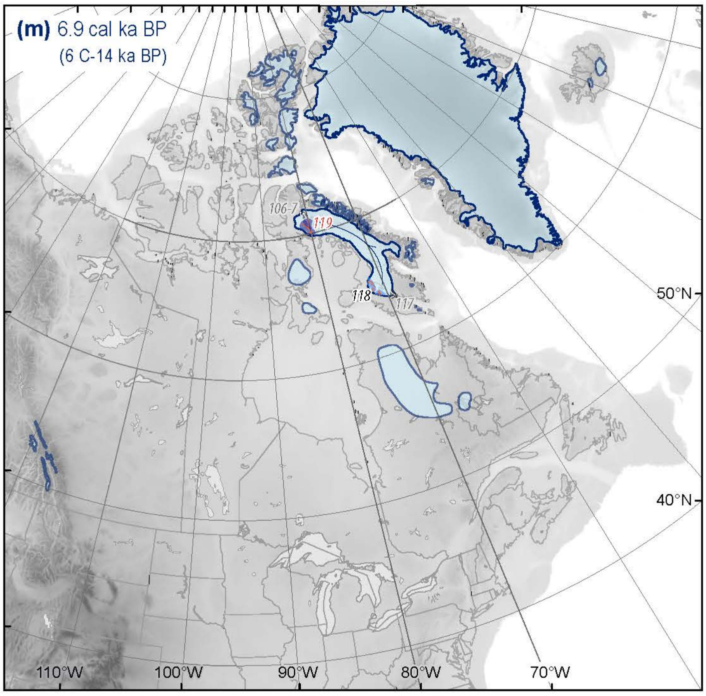

ID Name
134 Northeast Channel
131 (IS) In the Gully
45 Notre Dame Channel
130 Trinity Trough
25 Laurentian Channel
183 Green Bay Lobe
49 Huron-Erie Lobe
28 James Lobe
15 High Plains
27 Des Moines Lobe
1 Mackenzie Trough
171 Karlsefni Trough
168 Hopedale Saddle
129 Prince Gustaf Adolf Sea
31 Superior Lobe
23 Cumberland Sound
141 Kennedy-Robeson Channel
139 Yelverton Bay
127 Jones Sound
126 Cap Discovery
115 Broughton Trough
113 Okoa Bay
111 Clyde Trough
109 Scott Inlet
105 Navy Board Inlet
18 Amundsen Gulf
137 Tug Hill Plateau
124 Nansen Sound
24 Hudson Strait
154 Smoothstone Lake fragment
136 Oneida Lobe
155 IS3
148 Birch Mountains fragments
175 Great Slave Lake
11 Crooked Lake
158 Lac La Ronge
177 Peace River
156 IS4/5
178 Athabasca River
8 Collinson Inlet
144 Fort Simpson
10 M'Clintock Channel
143 Horn
146 Buffalo River
101 N Prince of Wales Island
179 Hayes Lobe
13 Peel Sound
12 Transition Bay
103 Bernier Bay
33 James Bay
164 Quinn Lake
165 Ekwan River / Winisk
121 S Southampton Island / pIS 5
188 Payne Bay
117 Frobisher Bay deglacial
106 S of Milne Inlet W
119 Erichsen Lake - Munk Island
30 25 20 15 10 5 0
Name ID
Saginaw Lobe 184
Hawke Saddle 169
Placentia Bay - Halibut Ch. 133
Maskwa 153
Bay of Fundy 185
Buffalo Ice Stream Corridor 159
Lake Michigan Lobe 30
Rocky Mountain Foothills 151
Central Alberta 15
Conception Bay 182
M'Clure Strait 19
Cartwright Saddle 170
Okak Bank 1, 2 167
Massey Sound 123
Frobisher Bay 117
McBeth Fiord 172
Phillips Inlet 140
Wellington Channel 128
Smith Sound / Nares Strait 126
Merchants Bay 116
Kangeeak Pt 114
Home Bay 112
Sam Ford Fiord 110
Buchan Gulf 108
Lancaster Sound 22
Eclipse Sound 104
Admiralty Inlet 21
Gulf of Boothia 20
Pasquia Hills fragments 160
IS2 152
Anderson 2
Bathurst 181
Cameron Hills fragment 145
Albany Bay 26
Horton / Paulatuk 3
Bear Lake 5
Hay River 176
Haldane 4
Red River Lobe 163
Caribou Mountains 147
Saneraun Hills 7
Suggi Lake 161
Kugluktuk 142
Rainy Lobe 180
Ungava Bay fans 1 17
Browne Bay 102
Cumberland Sound deglacial 23
Ungava Bay fans 2 16
Remnant Dubawnt IS corridor 174
Dubawnt Lake 6
Kogaluk River 187
C Southampton Island / pIS 11 122
Happy Valley - Goose Bay 186
Maguse Lake 166
S of Milne Inlet E 107
Amadjuak Lake 118
Steensby Inlet 120
Legend
Maximum
Best estimate
Minimum
Min. inversed
30 25 20 15 10 5 0 cal ka

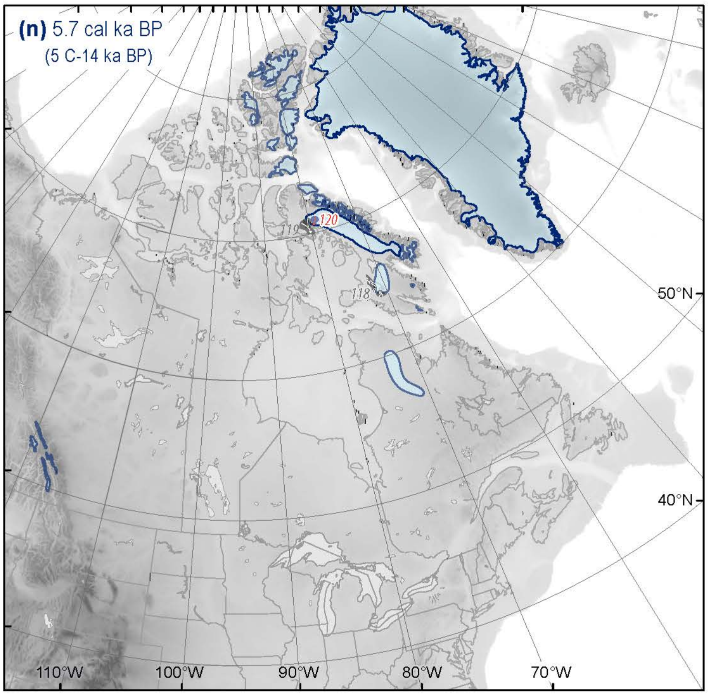


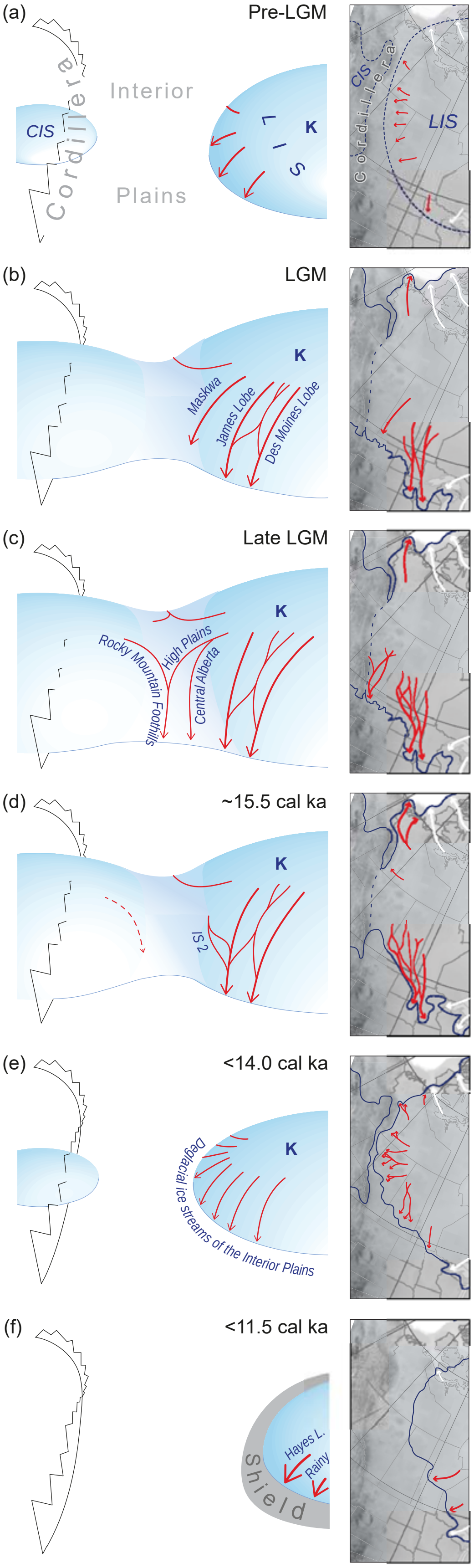
(a) LGM / 21.8 cal ka BP
(18 C-14 ka BP)
(b) 20.5 cal ka BP
(17 C-14 ka BP)
(c) 19.3 cal ka BP
(16 C-14 ka BP)
(d) 18.2 cal ka BP
(15 C-14 ka BP)
K
Q-L
F-B
S
SIL
LW
500 km
50°N
40°N
110°W
100°W
90°W
80°W
70°W
141
125
139
140
124
123
129
19
1
18
128
127
126
22
108-11
21
104-5
112-16
20
172
23
117
167
168
170
169
45
24
171
130
133
25
131
134
185
182
151
14
15
153
159
31
27
28
183
30
184
49


(e) 17 cal ka BP
(14 C-14 ka BP)
K
Q-L
F-B
S
50°N
40°N
110°W
100°W
90°W
80°W
70°W
(f) 15.5 cal ka BP
(13 C-14 ka BP)
K
Q-L
F-B
S
50°N
40°N
110°W
100°W
90°W
80°W
70°W
(g) 13.9 cal ka BP
(12 C-14 ka BP)
K
Q-L
F-B
S
50°N
40°N
110°W
100°W
90°W
80°W
70°W
(h) 12.9 cal ka BP
(11 C-14 ka BP)
K
Q-L
F-B
S
50°N
40°N
110°W
100°W
90°W
80°W
70°W

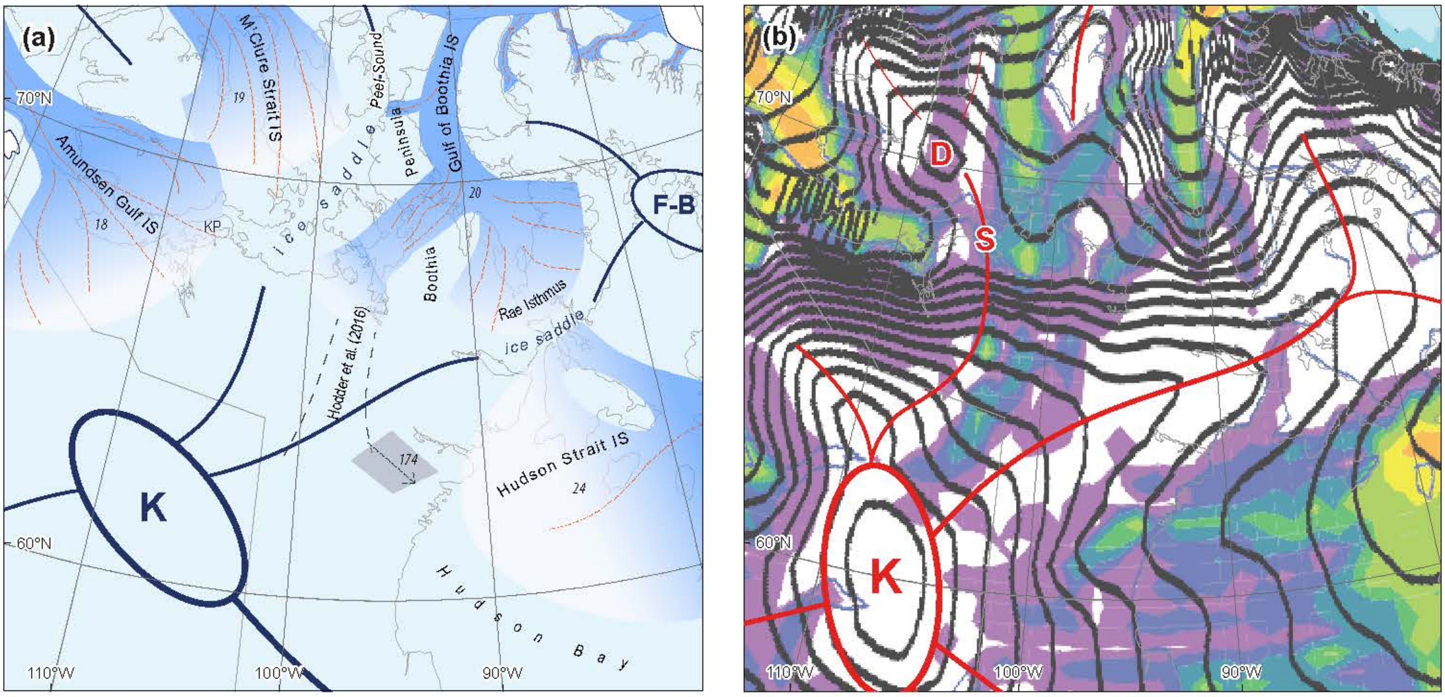

(i) 11.5 cal ka BP
(10 C-14 ka BP)
(j) 10.1 cal ka BP
(9 C-14 ka BP)
(k) 8.9 cal ka BP
(8 C-14 ka BP)
(l) 7.8 cal ka BP
(7 C-14 ka BP)
K
S
H
Q-L
F-B
50°N
40°N
110°W
100°W
90°W
80°W
70°W

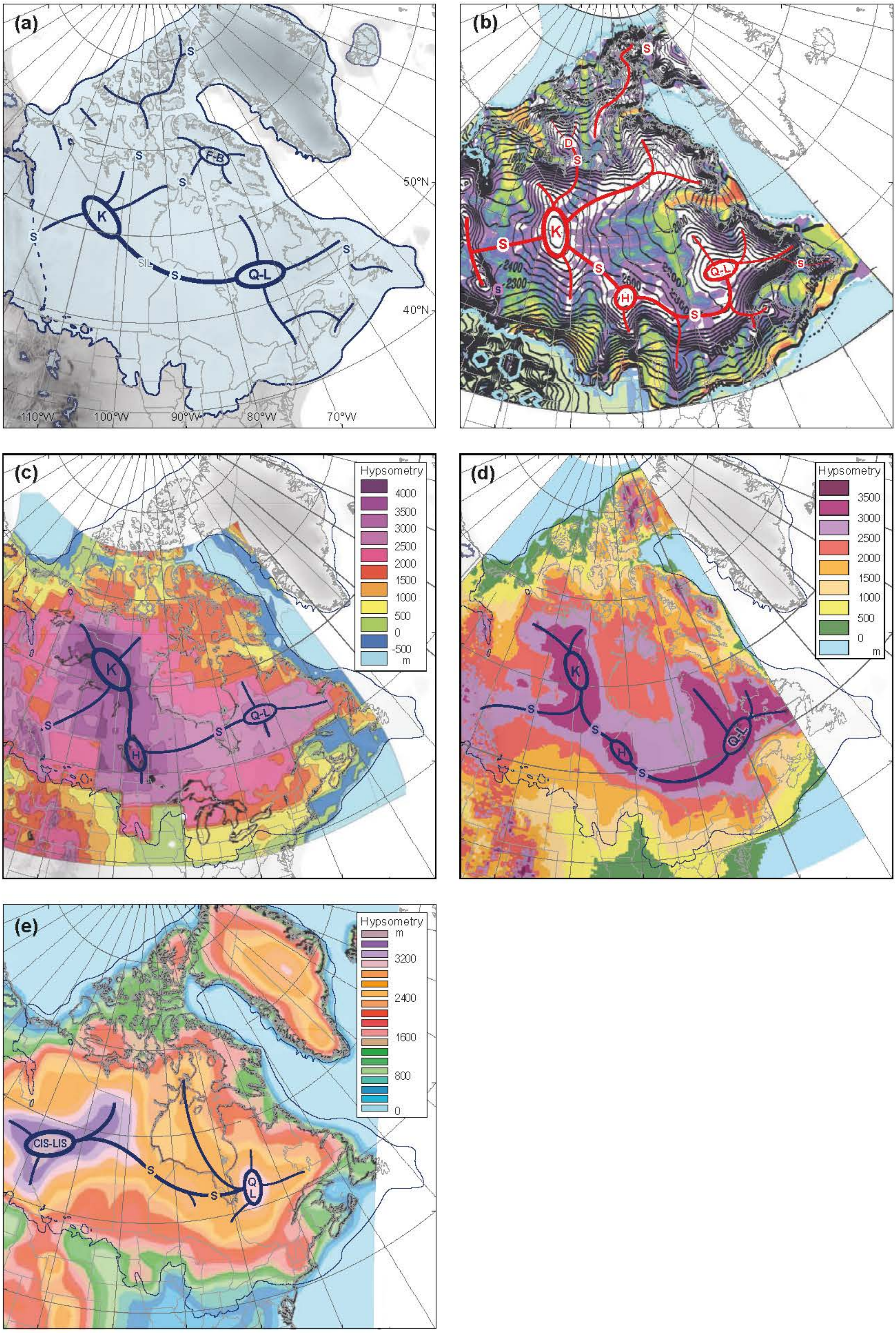

(m) 6.9 cal ka BP
(6 C-14 ka BP)
106-7
119
118
117
50°N
40°N
110°W
100°W
90°W
80°W
70°W


(n) 5.7 cal ka BP
(5 C-14 ka BP)
119
120
118
50°N
40°N
110°W
100°W
90°W
80°W
70°W

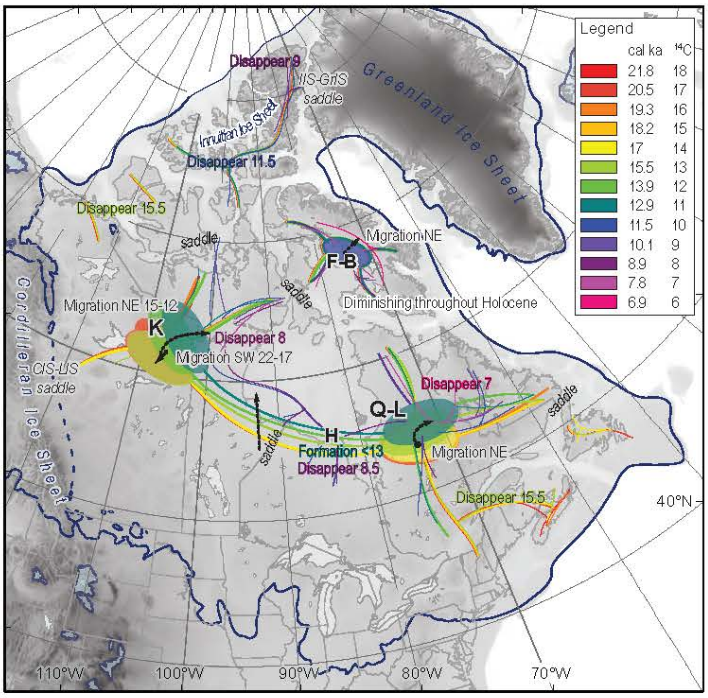


(a)
Pre-LGM
Cordillera
Interior
Plains
CIS
LIS
K
(b)
LGM
K
Maskwa
James Lobe
Des Moines Lobe
(c)
Late LGM
K
Rocky Mountain Foothills
High Plains
Central Alberta
(d)
~15.5 cal ka
K
IS 2
(e)
<14.0 cal ka
K
Deglacial ice streams of the Interior Plains
(f)
<11.5 cal ka
Hayes L.
Rainy
Shield


(a)
70°N
M'Clure Strait IS
19
Peel Sound
Peninsula
Gulf of Boothia IS
Amundsen Gulf IS
18
KP
ice saddle
20
F-B
Boothia
Rae Isthmus
ice saddle
Hodder et al. (2016)
174
Hudson Strait IS
24
K
60°N
Hudson Bay
110°W
100°W
90°W
(b)
70°N
D
S
K
60°N
110°W
100°W
90°W



(a)
S
F-B
K
SIL
Q-L
50°N
40°N
110°W
100°W
90°W
80°W
70°W
(b)
S
D
K
H
Q-L
2400
2300
2500
(c)
Hypsometry
4000
3500
3000
2500
2000
1500
1000
500
0
-500
m
K
H
Q-L
S
(d)
Hypsometry
3500
3000
2500
2000
1500
1000
500
0
m
K
H
Q-L
S
(e)
Hypsometry
m
3200
2400
1600
800
0
CIS-LIS
Q
L
S


Legend
cal ka ¹⁴C
21.8 18
20.5 17
19.3 16
18.2 15
17 14
15.5 13
13.9 12
12.9 11
11.5 10
10.1 9
8.9 8
7.8 7
6.9 6
Greenland Ice Sheet
Disappear 9
IIS-GrIS saddle
Innuitian Ice Sheet
Disappear 11.5
Disappear 15.5
saddle
Migration NE
F-B
Diminishing throughout Holocene
Migration NE 15-12
K
Disappear 8
Migration SW 22-17
CIS-LIS saddle
Cordilleran Ice Sheet
Disappear 7
Q-L
H
Formation <13
Disappear 8.5
Migration NE
Disappear 15.5
40°N
110°W
100°W
90°W
80°W
70°W

ID	Ice stream name
1	Mackenzie Trough
2	Anderson
3	Horton / Paulatuk
4	Haldane
5	Bear Lake
6	Dubawnt Lake
7	Saneraun Hills
8	Collinson Inlet
10	M'Clintock Channel
11	Crooked Lake
12	Transition Bay
13	Peel Sound
14	Central Alberta
15	High Plains
16	Ungava Bay fans 2
17	Ungava Bay fans 1
18	Amundsen Gulf
19	M'Clure Strait
20	Gulf of Boothia
21	Admiralty Inlet
22	Lancaster Sound
23	Cumberland Sound
24	Hudson Strait
25	Laurentian Channel
26	Albany Bay
27	Des Moines Lobe
28	James Lobe
30	Lake Michigan Lobe
31	Superior Lobe
33	James Bay
45	Notre Dame Channel
49	Huron-Erie Lobe
101	N Prince of Wales Island
102	Browne Bay
103	Bernier Bay
104	Eclipse Sound
105	Navy Board Inlet
106	S of Milne Inlet W
107	S of Milne Inlet E
108	Buchan Gulf
109	Scott Inlet
110	Sam Ford Fiord
111	Clyde Trough
112	Home Bay
113	Okoa Bay
114	Kangeeak Pt
115	Broughton Trough
116	Merchants Bay
117	Frobisher Bay
117	Frobisher Bay deglacial
118	Amadjuak Lake
119	Erichsen L - Munk Island
120	Steensby Inlet
121	S Southampton Island / pIS 5
122	C Southampton Island / pIS 11
123	Massey Sound
124	Nansen Sound
125	Cap Discovery
126	Smith Sound / Nares Strait
127	Jones Sound
128	Wellington Channel
129	Prince Gustaf Adolf Sea
130	Trinity Trough
131	(IS in) The Gully
133	Placentia Bay - Halibut Channel
134	Northeast Channel IS
135	(IS) offshore Massachusetts
136	Oneida Lobe
137	Tug Hill Plateau
139	Yelverton Bay
140	Phillips Inlet
141	Kennedy - Robeson Channel
142	Kugluktuk
143	Horn
144	Fort Simpson
145	Cameron Hills fragment
146	Buffalo River
147	Caribou Mountains
148	Birch Mountains fragments
150	pre-Maskwa
151	Rocky Mountian Foothills
152	IS2
153	Maskwa
154	Smoothstone Lake fragment
155	IS3
156	IS4/5
157	Winefred Lake fragment
158	Lac La Ronge
159	Buffalo Ice Stream Corridor
160	Pasquia Hills fragments
161	Suggi Lake
162	Saskatchewan River
163	Red River Lobe
164	Quinn Lake
165	Ekwan River
166	Maguse Lake
167	Okak Bank 1, 2
168	Hopedale Saddle
169	Hawke Saddle
170	Cartwright Saddle
171	Karlsefni Trough
172	McBeth Fiord
174	Remnant Dubawnt ice stream corridor
175	Great Slave Lake
176	Hay River
177	Peace River IS
178	Athabasca River IS
179	Hayes Lobe
180	Rainy Lobe
181	Bathurst IS
182	Conception Bay IS
183	Green Bay Lobe
184	Saginaw Lobe
185	Bay of Fundy
186	Happy Valley-Goose Bay IS
187	Kogaluk River
188	Payne Bay