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The Laurentide and Innuitian ice sheets during the Last Glacial Maximum

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Abstract

The Late Wisconsinan advance of the Laurentide Ice Sheet started from a Middle Wisconsinan interstadial minimum 27–30 ^{14}C ka BP when the ice margin approximately followed the boundary of the Canadian Shield. Ice extent in the Cordillera and in the High Arctic at that time was probably similar to present. Ice advanced to its Late Wisconsinan (stage 2) limit in the northwest, south, and northeast about 23–24 ^{14}C ka BP and in the southwest and far north about 20–21 ^{14}C ka BP. In comparison to some previous reconstructions of ice extent, our current reconstruction has substantially more Late Wisconsinan ice in the High Arctic, where an Innuitian Ice Sheet is generally acknowledged to have existed, in the Atlantic Provinces, where ice is now thought to have extended to the Continental Shelf edge in most places, and on eastern Baffin Island, where ice probably extended to the fiord mouths rather than to the fiord heads. Around most of the ice margin, the Late Wisconsinan maximum ice extent either exceeded the extent of earlier Wisconsinan advances or it was similar to the Early Wisconsinan advance. Ice marginal recession prior to 14 ^{14}C ka BP occurred mainly in deep water and along the southern terrestrial fringe. However, Heinrich event 1 probably drew down the entire central ice surface at 14.5 ^{14}C ka BP sufficiently to displace the Labrador Sector outflow centre 900 km eastward from the coast of Hudson Bay. The onset of substantial ice marginal recession occurred about 14 ^{14}C ka BP in the northwest, southwest, and south but not until about 10–11 ^{14}C ka BP in the northeast and in the High Arctic. Thus, the period of maximum ice extent in North America generally encompasses the interval from $\sim$24/21 to 14 ^{14}C ka BP, or considerably longer than the duration of the LGM defined as occurring during a period of low global sea level as well as during a time of relative climate stability $\sim$18 ^{14}C ka BP. The interval of advance of much of the Laurentide Ice Sheet to its maximum extent (between $\sim$27 ^{14}C ka BP and $\sim$24 ^{14}C ka BP) coincides with a suggested interval of rapid fall in global sea level to near LGM levels.

1. Introduction

This review restricts itself to addressing questions raised by the workshop on Ice Sheets and Sea Level of the Last Glacial Maximum convened by the EPILOG Project (Environmental Processes of the Ice Age: Land, Oceans, Glaciers). The workshop was a continuation of efforts to reconcile global ice volume histories with relative sea level (RSL) of sites far from glaciated regions and with other proxy ice volume records. The Last Glacial Maximum (LGM) is defined as occurring during a period of low global sea level as well as during a time of relative climate stability $\sim$18 ^{14}C ka BP (Clark and Mix, 2000). At the global scale, these efforts began more than 20 years ago with the CLIMAP Project reconstructions of ice sheets at the LGM (CLIMAP, 1976; Denton and Hughes, 1981). The histories of the North American ice sheets, and especially that of the Laurentide Ice Sheet, are of central importance to these efforts because of the large proportion of the Late Wisconsinan (i.e., marine isotope stage 2) and Holocene (stage 1) ice-equivalent sea level change that they account for. The range of estimates of ice volume and

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ice-equivalent sea-level of the LGM Laurentide Ice Sheet is $15.9–37 \times 10^6\,km^3$ and 40–92 m, respectively (Table 2 of Licciardi et al., 1998).

Several questions structure our review. What was the ice extent in North America east of the Cordillera during the interval of the global LGM centred on 18 ^{14}C ka BP (21.5 cal ka BP)? What was the probable extent of the ice sheet during the Middle Wisconsinan (isotope stage 3)? When did the margin advance to its LGM limit from the more restricted Middle Wisconsinan position? When did the ice margin retreat substantially from the LGM limit? And what was the general ice thickness and flow pattern at LGM? We use the compilations of Prest (1969), CLIMAP (Denton and Hughes, 1981), and Dyke and Prest (1987) as our points of comparison.

Since the CLIMAP reconstructions, new interpretations of North American glacial history for the LGM time interval have been stimulated mainly by progress resulting from continued field studies, including a renewed focus on the interpretation of offshore late Quaternary sediments within a glacial and glaciological context, and particularly by advances in dating methods. Primary among these has been the rapid growth of the AMS radiocarbon database, which seems to have had three main effects. First, the higher dating precision now possible has led to the rejection of large sets of conventional radiocarbon dates on bulk sediment organic matter, now recognized as anomalously old. Bulk dates on marine sediment are generally no longer used and similar dates on basal lake sediments, an important means of inferring time of deglaciation, are treated with increasing caution. The net effect of these reassessments of radiocarbon dates has been a decrease in many of the ages assigned to deglaciation, including the assigning of Late Wisconsinan ages to sites previously thought to have been glaciated earlier. The second important impact of AMS dating has been that it has enabled the investigation of marine sedimentation along the North Atlantic margin of the ice sheet in much more detail than was previously possible. Consequently, the timing and Laurentide sources of Heinrich events have largely been determined, and along with them, the times and at least general positions of ice advances to the LGM limit(s). Heinrich events (Heinrich, 1988) are inferred from widespread and distinctive sediment layers that contain an iceberg rafted coarse component. However, in the Labrador Sea the layers are dominated by fine-grained detrital carbonate sediments (Chough, 1978; Kirby, 1996) that appear to record large meltwater or turbidite flows emanating from Hudson Strait (Hesse et al., 1997; Hesse and Khodabakhsh, 1998). Sudden iceberg and meltwater release could have accompanied either a surge or a breakup of the Hudson Strait Ice Stream (cf. MacAyeal, 1993), events conceivably triggered by storage and release of subglacial meltwater (Schoemaker, 1991; Dowdeswell and Siegert, 1999). The third impact of AMS dating, as it pertains to this review, has been that we now have available many more finite age determinations on single organisms of Middle Wisconsinan age. Conventional radiocarbon dates on bulk samples (multiple organisms) in that age range were intrinsically ambiguous because of the possibility that the finite age determinations resulted from inclusion in the sample of a mixture of old and younger materials. The second important dating advance has been the advent of surface exposure dating using cosmogenic nuclides. Although these results are somewhat more ambiguous, they have allowed testing of earlier ideas about the antiquity of weathered and otherwise older-looking terrains around the margins of the Laurentide and Innuitian ice sheets.

Because of the large number of results available, radiocarbon dates are not tabulated in this paper. Instead, we present data only in summary form and mention only key dates specifically. However, we have attempted elsewhere a listing of all those dates known to us that help constrain the timing of the advances to the LGM limit and of the initial recession from it, where this occurred before 14 ^{14}C ka BP (Dyke et al., 2001).

The discussion below addresses the questions set out above. Discussion generally proceeds around the ice sheet margins counterclockwise from the far north. First we discuss recent interpretations of ice extent and age, then internal ice configurations.

2. Ice extent and age

A large fraction of the Laurentide Ice Sheet apparently reached its maximum extent prior to the LGM (19–22 cal ka BP; Mix et al., 2001) by up to several thousand years, and most of the ice sheet remained at this limit well after the LGM, with large-scale ice margin retreat beginning only after 14 ^{14}C ka BP. For this reason, we refer to the timing of the last advance of the Laurentide ice margin to its maximum extent as occurring during the Late Wisconsinan, which roughly corresponds to marine isotope stage 2, rather than during the LGM, as conventionally defined. In most places, the Laurentide ice margin was at or near its maximum extent during the LGM, but we wish to emphasize the point that it was also there before and after the LGM.

The compilation of North American LGM ice sheet margins by Dyke and Prest (1987) was generally similar to that of Prest (1969, 1970) and of CLIMAP (Denton and Hughes, 1981) for the western and southern margins, but Dyke and Prest (1987) showed much less ice than did Prest (1969) or CLIMAP in the Canadian Atlantic Provinces and in the High Arctic. The two more recent compilations expressed the range of interpretations being debated in the early to middle 1980s and, perhaps predictably, interpretations of ice extent at LGM have since changed most significantly in the two regions where disagreement was greatest. In addition, LGM ice extent on eastern Baffin Island is now thought to have been larger than that portrayed by Dyke and Prest (1987), though the change there is much smaller than in the other two regions. Firm LGM limits have yet to be resolved in any of these regions. Elsewhere, the LGM limit is thought to be similar to that portrayed in 1987, though there is rarely good evidence regarding the exact placement of the 18 ka margin (Fig. 1). Our discussion focuses on the larger changes in interpretation and on remaining uncertainties.

From Melville Island in the northwest to Cape Cod in the southeast, the LGM Laurentide limit follows a mapped set of moraines or till sheet limits (Flint et al., 1959; Prest et al., 1968; Dyke and Prest, 1987), or otherwise it follows the line of confluence with Cordilleran ice. East and north from Cape Cod, the limit is largely offshore and its placement, as illustrated in this review (Fig. 1), is constrained by mapped geological features only in places.

2.1. Innuitian region

After more than a century of debate, intensively for the last 25 years, about the *existence* of an Innuitian Ice Sheet during the LGM over the northern half of the Canadian Arctic Archipelago, a consensus has emerged that such an ice sheet did in fact cover most of that region. The most convincing evidence of extensive Innuitian ice comes from mapping of two sets of features. The first are the flow traces striations, flutings and erratic dispersal trains of ice streams that flowed along several of the inter-island channels and large fiord systems (Blake, 1992a, b, 1993; Bell, 1996; Dyke, 1999; Lamoureux and England, 2000; O'Cofaigh et al., 2000). The second are sets of regionally extensive lateral meltwater channels, which everywhere descend to a marine limit of early Holocene age (Bednarski, 1998; Dyke, 1999; England, 1999; Smith, 1999; England et al., 2000). Putative higher and older marine shorelines (e.g., England et al., 1981; Retelle, 1986; Bell, 1996), a keystone in the earlier interpretation of much more restricted LGM ice, are now reinterpreted as recessional morainal or outwash features (England, 1999). Removal of an LGM Innuitian ice load had long been the simplest interpretation of regional Holocene rebound (Blake, 1970, 1975). Although indications of long-stable relative sea levels at or close to the Holocene marine limit at both peripheral and central Innuitian sites (England, 1983, 1992, 1997) confounded that simple interpretation, that apparent stability of RSL now appears to result in part from an exaggerated reservoir-age effect on the infaunal, deposit-feeding mollusc, *Portlandia arctica* (England, Dyke, and McNeely, unpublished data). Thus, in the current view, the glacial and sea-level histories of the Canadian High Arctic are in accordance with the notion of an Innuitian Ice Sheet.

However, with the exception of a few small probable nunataks along the Nares Strait coast of Ellesmere Island (England, 1999), the margin of the Innuitian Ice Sheet at its maximum extent evidently lay offshore and has yet to be mapped. Innuitian ice was probably fully coalescent with Greenland along Nares Strait in the east and with Laurentide ice along Parry Channel in the south, as proposed by Blake (1970) and illustrated by Prest (1969) and Funder and Hansen (1996). Although Innuitian ice was clearly extensive in the eastern and central parts of the archipelago, some ice-free terrain may have existed in the western part, for at least one "old" Laurentide till lies beyond the Late Wisconsinan Laurentide limit on Melville Island, and there is no present indication of that till being subsequently covered by Innuitian ice (Hodgson et al., 1984). Evidence for only local ice caps is known on western Melville Island (Hodgson, 1992) and on Prince Patrick Island (Hodgson, 1990) following a regional Laurentide glaciation of suggested Early Pleistocene age. Although the northwestern limit of the Innuitian Ice Sheet has not been mapped, and the line of Fig. 1 is therefore speculative, the uplift pattern (Blake, 1970; Dyke, 1998) and the location of the outflow centre or ice divide (Lamoureux and England, 2000) indicate that this margin lay on the polar continental shelf. General support for an ice limit in this position can be drawn from the description of morainal topography on the polar shelf (Pelletier, 1966). A Late Wisconsinan age for such an ice limit is inferred from marine shells that were probably ice transported onto Brock Island and dated at $27\,860 \pm 210$ ^{14}C yr BP (TO-3561; Hodgson et al., 1994). An ice-thrust moraine on the same island contains marine shells dated $38\,590 \pm 1340$ ^{14}C yr BP (GSC-381; Lowdon et al., 1967), indicating that the last ice advance may have followed an ice-free interval of 10 000 years or more.

The pre-LGM ice configuration in the Innuitian region has received little attention because there are few stratigraphic sections that elucidate events prior to LGM. Blake (1980) proposed a Middle Wisconsinan non-glacial interval for the Cape Storm area of Ellesmere Island. Since then, a large number of marine shells redeposited as erratics in till and a few sub-till marine and terrestrial deposits have been dated by radiocarbon (Figs. 2 and 3; see Dyke et al., 2001 for references). Because most of these are AMS dates on single organisms, the dates cannot be dismissed as blended ages of postglacial and older materials. Furthermore, many have ages well below the upper limits of radiocarbon dating (~ 50 ka) and are determinations run on materials lacking any evident contamination. If the dates are therefore accepted, it appears

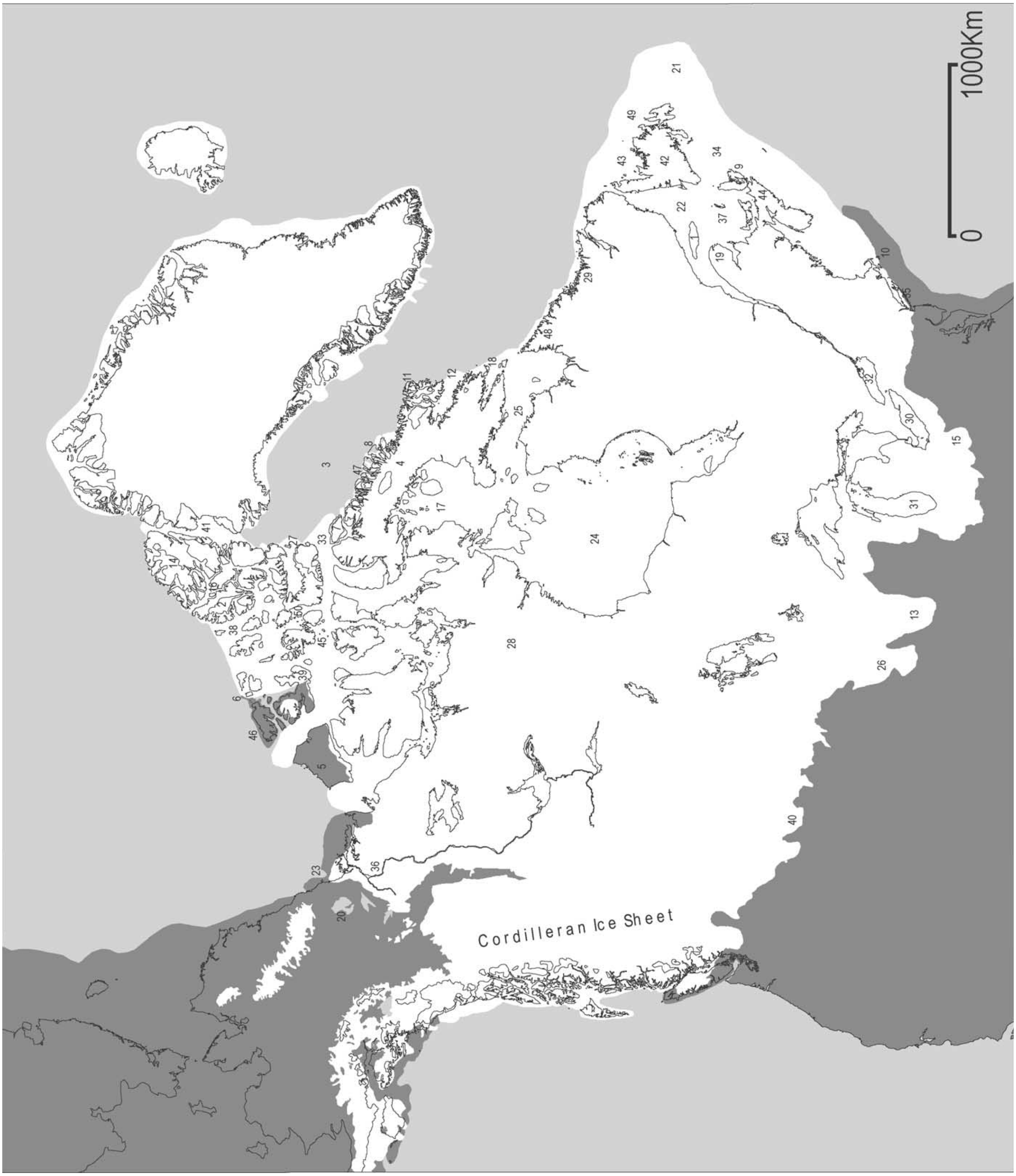

Fig. 1. Ice sheet margins in North America at LGM with major place names and ice lobe names. 1, Anticosti Island; 2, Axel Heiberg Island; 3, Baffin Bay; 4, Baffin Island; 5, Banks Island; 6, Brock Island; 7, Bylot Island; 8, Cape Aston; 9, Cape Breton Island; 10, Cape Cod; 11, Cumberland Peninsula; 12, Cumberland Sound; 13, Des Moines Lobe; 14, Ellesmere Island; 15, Erie Lobe; 16, Eureka Sound; 17, Foxe Basin; 18, Frobisher Bay; 19, Gaspé Peninsula; 20, Glacial Lake Old Crow; 21, Grand Banks; 22, Gulf of St. Lawrence; 23, Herschel Island; 24, Hudson Bay; 25, Hudson Strait; 26, James Lobe; 27, Jones Sound; 28, Keewatin; 29, Labrador; 30, Lake Erie; 31, Lake Michigan; 32, Lake Ontario; 33, Lancaster Sound; 34, Laurentian Channel; 35, Long Island; 36, Mackenzie River and Lobe; 37, Magdalen Islands; 38, Massey Channel; 39, Melville Island; 40, Montana; 41, Nares Strait; 42, Newfoundland; 43, Notre Dame Channel; 44, Nova Scotia; 45, Parry Channel; 46, Prince Patrick Island; 47, Remote Lake; 48, Torngat Mountains; 49, Trinity Trough; 50, Wellington Channel.

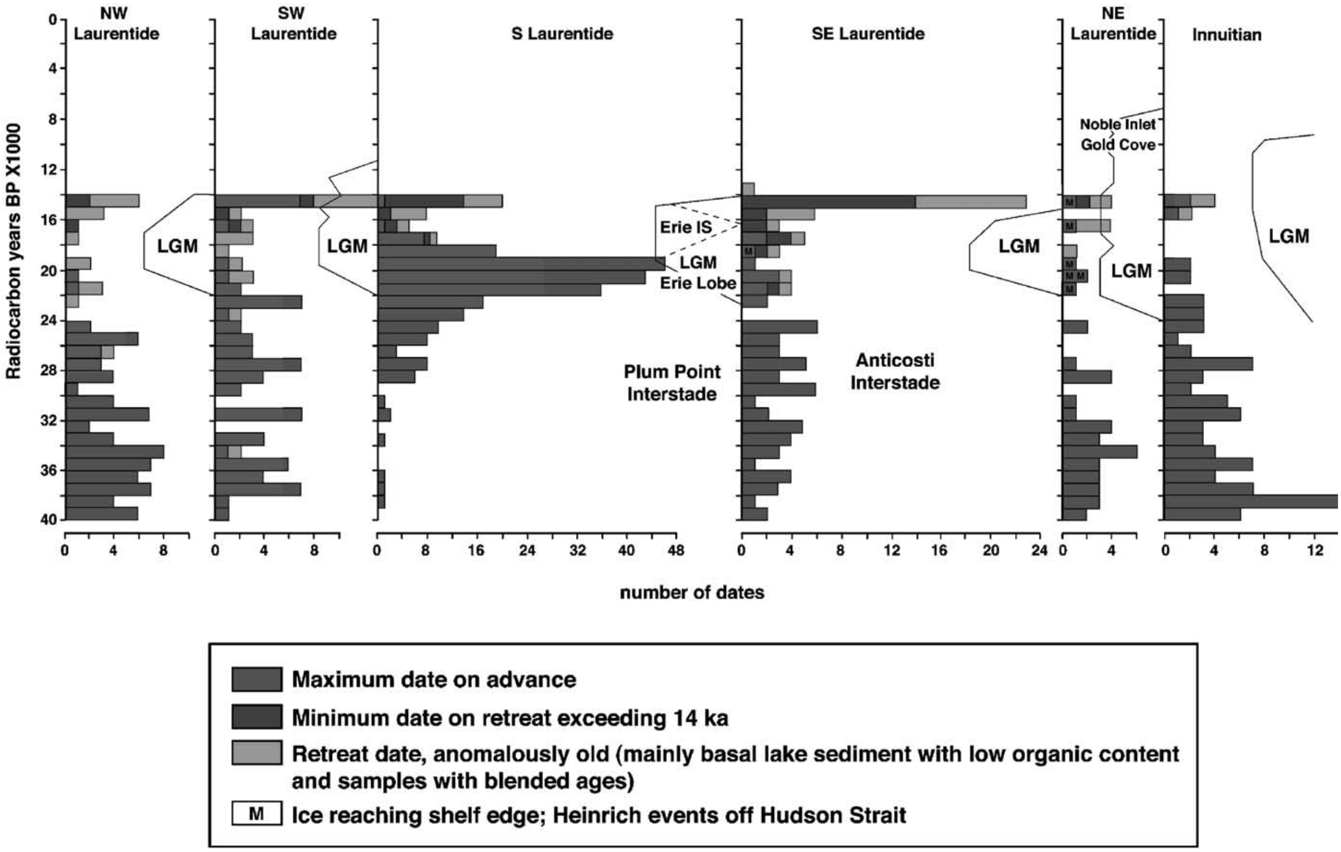

Fig. 2. Histograms of radiocarbon dates that provide maximum limiting ages and minimum limiting ages on the Late Wisconsinan ice advances and recessions. The line to the right of each histogram indicates the probable movement of the ice margin.

that late in the Middle Wisconsinan ice extent in the Innuitian region was similar to present and that the ice sheet built to its LGM limit at approximately 20 ^{14}C ka BP (Blake, 1992a, b; England, 1999). Nowhere in this region is there evidence of recession prior to 11 ^{14}C ka BP (Hodgson, 1989) and most recession occurred after 10 ^{14}C ka BP.

2.2. Northwestern Laurentide margin

The northwest Laurentide margin shown here is similar to earlier portrayals although it has been mapped in greater detail along the mountain front west of the Mackenzie River and the relative ages of Laurentide and Cordilleran moraines have been determined in areas of overlap (Duk-Rodkin and Hughes, 1991). Currently, the most important question regarding this margin is its age, whether it is generally of global LGM age or older. Dyke and Prest (1987) assigned this limit an age of 25 ^{14}C ka BP, placing it in the earliest Late Wisconsinan, whereas Lemmen et al. (1994) assigned it an age of 30 ^{14}C ka BP, which places it in the Middle Wisconsinan, a time of reduced ice cover elsewhere in glaciated North America (Fig. 3) and globally (Shackleton, 1987). Both reconstructions showed recession starting before 18 ^{14}C ka BP. Different interpretations are possible because of the ambiguity of a few radiocarbon-dated sites (see Dyke et al., 2001 for details). The older age assignment is based on several radiocarbon dates in the 30–40 ^{14}C ka BP range on plant material below drift near the glacial limit and below sediments of Glacial Lake Old Crow, which formed in unglaciated Yukon Territory when ice stood at the limit and blocked eastward drainage. However, these dates do not necessarily closely limit the age of the last ice advance. In fact, the youngest dated material below Glacial Lake Old Crow sediments is a mammoth tusk with AMS dates of 25,170 $\pm$ 630 ^{14}C yr BP (CRNL-1232) and 24,700 $\pm$ 250 ^{14}C yr BP (RIDDL-229; Morlan, 1986; Morlan et al., 1990). Furthermore, a continuous series of AMS mammal bone collagen dates between 24 and 40 ^{14}C ka BP from Old Crow Basin precludes existence of the lake during that interval and the abrupt termination of the series at $\sim$24 ^{14}C ka BP likely signifies flooding of the basin, as suggested by Thorson and Dixon (1983).

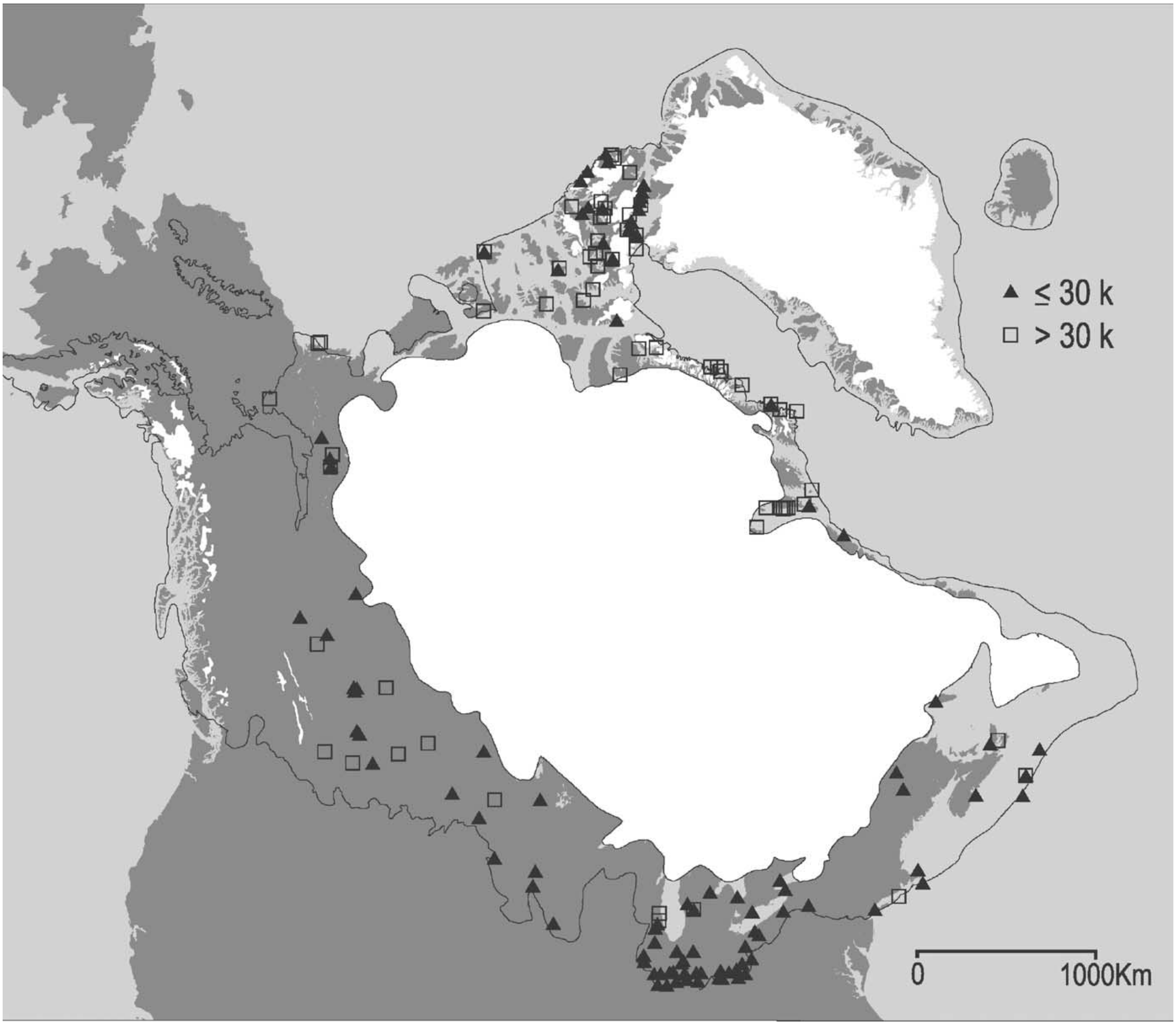

Fig. 3. The distribution of Middle Wisconsinan (stage 3) sites that have yielded finite radiocarbon ages. The probable interstadial ice margin at 27–30 ^{14}C ka BP approximately follows the margin of the Canadian Shield.

In addition, plant detritus from basal silt, which is thought to be either the lower part of a nearshore glaciolacustrine facies or the top of underlying alluvial channel sediment, in the Bluefish basin of Glacial Lake Old Crow dated 20,800 ± 200 ^{14}C yr BP (GSC-3946; Blake, 1987). That result is compatible with an LGM age assignment. Furthermore, well-dated alluvial sediments along the Porcupine River downstream from Old Crow indicate that a period of fluvial aggradation began before 30.6 ^{14}C ka BP and ended after 26 ^{14}C ka BP. Overflow from Glacial Lake Old Crow evidently did not occur during that interval. Instead, down cutting to form the Ramparts of the Porcupine River, which resulted from overflow of the glacial lake, occurred after 26 ^{14}C ka BP, which is fully in agreement with the series of mammal dates from the basin itself (Thorson and Dixon, 1983). Finally, AMS ^{14}C dates on wood redeposited in glaciofluvial sediment along the upper Mackenzie River, well behind the LGM limit, indicate that ice-free conditions existed there until at least 27.2 ^{14}C ka BP (Smith, 1992).

Several dates indicate that recession from the LGM limit in the northwest may have started prior to 18 ^{14}C ka BP (Lemmen et al., 1994). All of these dates are ambiguous, however, because of uncertainties about stratigraphic origin, probabilities of blended age determinations on mixed assemblages, stratigraphic dating reversals, and hardwater errors (see discussions in Dyke et al., 2001). Putting these dates aside, the earliest date on recession that appears acceptable is $16\,200 \pm 150$ ^{14}C yr BP (RIDDL-765) on bone of an extinct horse from colluvial sand overlying till on Hershel Island, Yukon Territory, which is part of the terminal moraine (Harington, 1989). Otherwise, there is no firm radiocarbon evidence of significant regional recession of the Mackenzie Lobe until $14\,400 \pm 180$ ^{14}C yr BP (GSC-1792), a date on peat overlying till. Laurentide lobes northeast of the Mackenzie Lobe evidently held near maximal positions until $\sim$13 ^{14}C ka BP (Dyke and Prest, 1987; Vincent, 1989). Chlorine-36 exposure dating of four boulders at the LGM limit along the west side of the Mackenzie Lobe has produced ages of 20–28 ^{36}Cl ka BP ($\sim$17–24 ^{14}C ka BP if ^{36}Cl ages are comparable to calendar years). These ages are compatible with the LGM age inferred from the radiocarbon record discussed above. The earliest deglacial phase is $\sim$19 ^{36}Cl ka BP ($\sim$17 ^{14}C ka BP; Duk-Robin et al., 1996).

2.3. Southwestern Laurentide margin

There has been much contention about whether Laurentide and Cordilleran ice coalesced at LGM, particularly in the south, but more recently farther north. Dyke and Prest (1987) showed full coalescence based on evidence of coalescence of alpine and Laurentide ice in Montana (Fullerton and Colton, 1986) and recognition of the problematic nature of the radiocarbon dates indicating the contrary in Alberta (see Dyke et al., 2001). Recent ^{36}Cl dating of erratics in the Foothills Erratics Train, which formed along the coalescence of the two ice sheets, confirms that the erratics were deposited during the Late Wisconsinan (Jackson et al., 1997).

However, problems remain with respect to the timing and nature of Laurentide–Cordilleran coalescence north of the erratics train. In contrast to the clear evidence of coalescent flow along the erratics train, farther north the flow patterns of the two ice sheets are opposed or cross cutting (Prest et al., 1968). This pattern might indicate expansions of one or both ice sheets across the zone of coalescence after separation. Nevertheless, AMS-dated wood below Cordilleran till indicates that the east-central Cordilleran ice margin lay well west of the Laurentide limit at 18.7 ^{14}C ka (TO-709) and at 15.2 ^{14}C ka BP (TO-708). Bobrowski and Rutter (1992) concluded, therefore, that coalescence did not occur in that region during the last glaciation. Because the last glaciation is thought by them and by others (e.g., Liverman et al, 1989; Young et al., 1994) to be the only Laurentide glaciation to have reached western and southern Alberta, the two ice sheets may never have coalesced in this part of the 'ice-free corridor'. It is, however, difficult to conceive of how the two ice sheets might have flowed in coalescence for nearly 1000 km southward along the erratics train while remaining separated northward from the head of the train. We therefore continue to show the ice sheets as fully coalescent at LGM.

The timing of the last ice advance across central and southern Alberta and Saskatchewan is tightly constrained by many radiocarbon dates on sub-till materials, including numerous recent dates on faunal remains (e.g., Burns, 1996) in addition to those on wood and plant remains from the longer-known sections (Figs. 2 and 3; Dyke et al., 2001). The advance evidently started close to the edge of the Canadian Shield perhaps as late as 22–23 ^{14}C ka BP and reached close to its limit in the Des Moines Lobe by 20 ^{14}C ka BP (Ruhe, 1969; Clayton and Moran, 1982). The southwestern Laurentide margin thus advanced 500–1000 km in about 2–3 ka. Oscillations of the margin after 20 ^{14}C ka BP are not well defined (Clayton and Moran, 1982), but the final advance to the local glacial limit in the Des Moines lobe is well dated at 14 ^{14}C ka BP (Lowell et al., 1999).

Any substantial recession prior to 14 ^{14}C ka BP was necessarily limited to the margin west of the Des Moines Lobe and likely west of the James Lobe. If such recession occurred, there is no current radiocarbon evidence for it, with all pre-14 ka dates from the region (16 ages mainly on lake sediments) being suspect. Minor recession probably was underway in Minnesota, east of the Des Moines Lobe by 16.5 ^{14}C ka (Matsch and Schneider, 1988).

2.4. Southern Laurentide margin

The Laurentide margin south of the Great Lakes is more extensively studied and better dated than any other segment of the margin. Because the ice here advanced across a landscape that was at least partly forested, there are numerous exposures of sub-till forest and peat remains. The till also contains wood inclusions, some of the wood evidently sheared off while green by the advancing ice. Fig. 2 includes a plot of 225 radiocarbon dates on such materials that indicate the general time of advance to the LGM limit and to moraines close to the limit in this region. Other than for the greatly increased sample size, the features of this distribution of ages rising sharply in abundance at $\sim$23 ^{14}C ka BP to a strong single mode at 19–20 ^{14}C ka BP and declining sharply at 18 ^{14}C ka BP are the same as illustrated by Dreimanis and Goldthwait (1973), when this advance was already the best-dated Quaternary event in the world.

Both Lake Erie and Lake Ontario were largely or entirely ice free during the Plum Point Interstade, which is dated between 23 and 28 ka BP (Karrow, 1989), and the advance of the Ontario Lobe to the LGM limit culminated after 22,800 ± 450 (GSC-816), which is a date on wood in interstadial sediments below drift near Niagara Falls, ON, a site about 100 km behind the LGM ice limit (Hobson and Terasmae, 1969). Hansel and Johnson (1992) suggested that the first advance of the Lake Michigan Lobe to its Late Wisconsinan limit occurred at ~26 ^{14}C ka BP. However, the age of that advance, which is based on a sample of organics from fluvial sediment, seems problematically early in light of the large number of radiocarbon dates indicating ice-free conditions at that time extending from the Canadian Cordillera (Clague, 1989) through to the Gulf of St. Lawrence (Fig. 3).

Close scrutiny of the chronology of these lobes subsequent to reaching their maximum limit has led to additional important insights. Hansel and Johnson (1992) suggested that subsequent advances of the Lake Michigan Lobe ~22.5, 18.5, 17.5, and 15.5 ^{14}C ka BP, all occurred in proximity to the LGM limit and the margin remained near the maximum extent until the onset of rapid retreat ~14 ^{14}C ka BP. Like the Lake Michigan Lobe, lobes in the Erie and Ontario basins reached their maximum extent ~23 ^{14}C ka BP (Dreimanis, 1977; Lowell et al., 1990), experienced several subsequent fluctuations near the maximum extent, and retreated rapidly into the lake basins beginning ~14 ^{14}C ka BP. Lowell et al. (1999) and Clark et al. (2001) found that fluctuations of the ice margin in the Great Lakes region and during the last deglaciation can be correlated to the stadials and interstadials in the Greenland ice cores (Stuiver and Groots, 2000), suggesting that the southern ice margin was sensitive to climate changes in the North Atlantic region.

Lowell's analysis also illuminates what may be an important problem in the "established" stratigraphy of this region (Dreimanis and Karrow, 1972). Each recessional interval of the Erie Lobe south of Lake Erie resulted in forest growth across recently deglaciated terrain. However, there is no forest or other dated macrofossil interval that corresponds to the proposed Erie Interstade, the largest of the proposed Late Wisconsinan recessions, when the ice margin is thought to have retreated northward from the maximum limit well into the Lake Erie basin ~16 ^{14}C ka BP (Morner and Dreimanis, 1973). The only organic remains yet found in association with Erie Interstade deposits yielded a radiocarbon date of 37,840 ± 3255 ^{14}C yr BP (St-3438; Morner and Dreimanis, 1973), suggesting either reworked organics or that the interpretation of the stratigraphic context of the unit is incorrect. Circumstantial evidence for the Erie Interstade comes from the fine-grained character of tills deposited by ice readvance in proglacial lakes that would have existed in this region during the interstade (Dreimanis and Goldthwait, 1973). Moreover, there is clear evidence for the Erie Interstade in the Lake Ontario region that, although not precisely dated, supports an age of ~16 ^{14}C ka BP (Ridge, 1997; Licciardi et al., 1999).

2.5. Southeastern Laurentide–Appalachian ice margin

The terrestrial ice margin between western New York state and Cape Cod (Fig. 1) is the conventional LGM limit of earlier reconstructions. The timing of advance to this limit is best constrained by the youngest ice-thrust materials in the end moraine zone on Long Island (21,750 ± 750 ^{14}C yr BP [SI-1590]; Sirkin and Stuckenrath, 1980). Ice recession remains poorly dated in New England. The best evidence for early recession from the terminal moraine is the onset of sedimentation at the south end of Glacial Lake Hitchcock at 15.6 ^{14}C ka BP. This date is derived by adding the number of varve years between that event and deposition of younger AMS-dated plant remains in varves farther north in the lake (Ridge and Larsen, 1990; Rittenour et al., 2000).

Interpretations of LGM ice extent in the Canadian Atlantic Provinces have been more in flux. The interpretation of restricted LGM ice there (Dyke and Prest, 1987; Grant, 1989) had rested on the following lines of argument. Gratton et al. (1984) argued that ice did not override Middle Wisconsinan marine sediments on Anticosti Island in the northern Gulf of St. Lawrence. Only ancient glaciation was recognized on the Magdalen Islands in the western part of the gulf and the presence there of surficial organic deposits dating from the Sangamonian (isotopic stage 5) suggested an absence of Late Wisconsinan ice cover (Prest et al., 1977; Dredge et al., 1992). Limited Laurentide ice was also indicated by a general preclusion of Laurentide (Canadian Shield) erratics from the Gapsé Peninsula of Quebec (Gray, 1987; Veillette and Cloutier, 1993) and from all but the northernmost tip of Newfoundland (Grant 1977, 1992). South and east of the erratic limit there is evidence only of glaciation from multiple Appalachian ice centres (Prest and Grant, 1969; Grant, 1977; Rampton et al., 1984). Furthermore, weathered terrain above fresh glacial trimlines on both sides of the Gulf of St. Lawrence (e.g., Grant, 1989, 1992) indicated that Appalachian ice was too thin to have flowed far into the gulf. On the continental shelf off Nova Scotia, "old" radiocarbon dates from marine sediments above the youngest till indicated deglaciation during the Middle Wisconsinan (e.g., King and Fader, 1986; Piper et al., 1990). In accordance with that view, a Late Wisconsinan till limit was mapped only 5 km off the Nova Scotia headlands and an unconformity was traced from that limit to an LGM relative sea level lowstand at −120 m (Piper et al., 1986). Finally, evidence of submergence from deep LGM or younger RSL lowstands as far "inland" as the Magdalen Shelf was compatible with restricted LGM ice loads (e.g., Quinlan and Beaumont, 1981; Peltier, 1996).

The interpretation of restricted LGM ice extent in Atlantic Canada has been overturned based on two main new lines of evidence. First, new cosmogenic exposure ages indicate that cold-based ice carapaces covered the weathered terrains of western Newfoundland at the LGM (Gosse et al., 1995). Second, many new AMS radiocarbon ages on marine sediment from the continental shelf off Nova Scotia and off southern Newfoundland show that all dated sediment above the youngest till, which extends to the continental shelf break, at least in places, and to the mouth of the Laurentian Channel, are of post-LGM age (Amos and Knoll, 1987; Bonifay and Piper, 1988; Gipp and Piper, 1989; Mosher et al., 1989; Amos and Miller, 1990; Piper et al., 1990; Forbes et al., 1991; Gipp, 1994; King, 1996; Stea et al., 1998; Josenhans and Lehman, 1999; see Dyke et al., 2001 for compilation of dates).

There is abundant evidence of ice-free conditions on the Scotian Shelf during the Middle Wisconsinan in the form of dates on redeposited shells of that age (King, 1996) and sparse evidence of ice-free conditions on land (see Dyke et al., 2001). In the current view, ice reached the continental slope off Nova Scotia before 18.3 ^{14}C ka BP (Mosher et al., 1989). Ice was receding on the central shelf by 17,050$\pm$155 ^{14}C yr BP (Beta-27229; Piper et al., 1990), the Scotian Shelf End Moraine Complex on the inner shelf was forming ~15–16 ^{14}C ka BP (King, 1996; Stea et al., 1998; Stea and Mott, 1998; Todd et al., 1999), and some terrestrial sites were ice-free in Nova Scotia by 14,010$\pm$90 ^{14}C yr BP (TO-2324; Mayle et al., 1993). Laurentide ice still extended to the Laurentian Moraine in the outer Laurentian Channel adjacent to Cape Breton Island until ~14.3 ^{14}C ka BP (Josenhans and Lehman, 1999). A till along the flank of the channel to the west, which has been traced to the moraine, indicates continued glaciation of the northeast Scotian Shelf at that time (King and Fader, 1990), though ice on the shelf presumably had disappeared by 14 ^{14}C ka BP, when shrubs had already colonized parts of Cape Breton Island (TO-2324 date noted above). Laurentide ice then evacuated the Gulf of St. Lawrence almost entirely by 13.9–14 ^{14}C ka (Rodrigues et al., 1993; Josenhans and Lehman, 1999). Whether ice held its position at the mouth of the Laurentian Channel between 18 and 14.3 ^{14}C ka is unknown.

The glacial history of the shelf east of the Laurentian Channel seems broadly similar. The maximum advance has not been dated here, but the ice front had retreated to a prominent end moraine zone off fiord mouths along the south coast of Newfoundland by 14 ^{14}C ka BP (Liverman and Bell, 1996; Shaw et al., 2000). The nearshore chronology renders problematic the evidence of a readvance (interpreted as a surge) onto the continental slope 300 km offshore at the mouth of Halibut Channel at ~11.5 ^{14}C ka BP (reservoir corrected age; Bonifay and Piper, 1988).

The glacial chronology of the Grand Banks and of the northeast Newfoundland shelf is less well known, because many of the regional data are summarized only in abstracts. Nevertheless, till on outer Grand Banks is bracketed by dates of 8–14 ^{14}C ka and 15–19 ^{14}C ka BP (Miller et al., 1998). Presumably, therefore, glaciation of the outer continental shelf not only continued after ice had receded onshore elsewhere in Newfoundland, but the ice was able to readvance in deepening water, which is difficult to reconcile with description of the offshore ice as "thin, buoyant, and short-lived". Hiscott and Aksu (1996) describe debris-flow wedges on the northeast Newfoundland continental slope off Trinity Trough and suggest that they formed at times when sea level was lower and glacier ice reached the shelf edge at 400 m water depth. The youngest glacigenic debris-flow wedge apparently formed during isotope stages 2–4 based on the ^{18}O stratigraphy, but the chronology is otherwise unconstrained. The large Trinity Moraine, mid-way up the trough (King and Fader, 1992) was formed at a terminus of confluent ice that emanated from Trinity, Conception, and Bonavista Bays. Although the moraine is undated, ice had receded onto land in Bonavista Bay by ~13 ^{14}C ka BP (Cumming et al., 1992), as it had along the coast of Notre Dame Bay farther northwest (Scott et al., 1991; Shaw et al., 1999). We show ice to the shelf edge off northeast Newfoundland (Fig. 1), but this remains highly speculative. Indeed, foraminifera from sediment above till in Notre Dame Channel on the mid-shelf produced an age of ~21 ^{14}C ka BP (Scott et al., 1989), and the dated assemblage was of interstadial (warm) rather than proglacial character. Notre Dame Channel is the first trans-shelf channel north of Trinity Trough.

There is now a need to reconsider whether the RSL record of Atlantic Canada is compatible with the revised interpretation of ice load history. An extensive, but thin, LGM ice load of short duration would probably not conflict with the sea level history because most of that ice load would have been replaced by the current water load. However, current interpretations favour ice loads, including those in offshore areas, thick enough to have survived and readvanced to the continental slope as late as 11.5 ^{14}C ka BP (Bonifay and Piper, 1988) or into Younger Dryas time (e.g., King, 1994; Stea and Mott, 1998; Stea et al., 1998). A sustained ice load would have delayed crustal rebound until after much of the eustatic sea level rise had occurred, hence resulting in raised shorelines of Holocene age through much of the region. But instead 110 m of submergence appears to have occurred on the Magdalen Shelf since deglaciation ~10 ^{14}C ka ago (Josenhans and Lehman, 1999).

2.6. Northeastern Laurentide margin

The northeastern Laurentide margin stretches from southern Labrador (52°N) to the head of Baffin Bay (74°N). Most recent efforts along this margin have focused on studies of marine sediment cores from the continental shelf and slope, mapping the northernmost advances of the Labrador ice margin, studies of lake sediment cores collected from weathered terrains beyond the oldest moraines on Cumberland Peninsula of Baffin Island, and on cosmogenic exposure dating of these same terrains and bounding moraines.

Isotopic and geochemical properties of different sediment size fractions have been used to trace the provenance of sediments derived from glacial erosion and transported to the North Atlantic by icebergs and meltwater during Heinrich events (e.g., Hemming et al., 2000a, b; Barber, 2001). In a general sense, these studies show that the non-carbonate component of Heinrich layers had their source in bedrock of the structural Churchill Province of the Canadian Shield, which extends both east and west of Hudson Bay (Barber et al., 1995; Gwiazda et al., 1996a, b; Hemming et al., 1998, 2000a, b; Hall et al., 1999; Barber, 2001). For example, Hemming et al. (2000a, b) used the $^{40}Ar/^{39}Ar$ ages of individual hornblende grains from a core from Orphan Knoll off the northeast Newfoundland Shelf to interpret the history of ice discharge into the Labrador Sea from the Laurentide margin. Just *after* H-2, many grains with ages comparable to that of the Grenville Province, which occupies the southern part of the Quebec–Labrador Peninsula, were deposited at Orphan Knoll. These grains were probably delivered by icebergs calved from ice exiting Hamilton Inlet on the central Labrador coast.

The Saglek moraines are generally thought to mark the vertical limit of Late Wisconsinan ice in the Torngat Mountains, northern Labrador, which suggests that large areas of the mountains remained ice free as nunataks (Ives, 1978; Clark, 1988). However, the age control on these moraines is poor (Clark et al., 1989), leaving open the possibility that the moraines are older or younger than LGM. Clark et al. (2000a, b) recently reported ^{10}Be cosmogenic ages on the boulders from Saglek moraines, including some from the type area. These suggest the moraines were deposited between 10.9 and 17.1 ^{10}Be ka BP, with an uncertainty-weighted mean age of 12.4 ± 1.6 ^{10}Be ka BP, or in latest Wisconsinan time. Further dating is required to evaluate whether thicker ice existed in the mountains earlier in the Late Wisconsinan.

A specific LGM margin cannot yet be defined along the Labrador Shelf with any certainty. Hall et al. (1999) provide the best age control from the marginal troughs on the inner shelf. These data suggest an interval of ice-free conditions during the Middle Wisconsinan. Marine and glaciomarine sediments of that age are unconformably overlain by deglacial sediments of Younger Dryas age with no intervening till or ice-proximal sediment. Possibly the LGM ice margin extended seaward along the margin of mapped upper till (Josenhans et al., 1986), but no limiting radiocarbon dates have been reported at the till contacts or its limit.

Important new data concern the history of ice streams in Hudson Strait and Cumberland Sound. The largest known Pleistocene northward advance of Labrador ice extended *across* Hudson Strait and the mouth of Frobisher Bay on Baffin Island during the Gold Cove Advance of terminal Younger Dryas age (Miller and Kaufman, 1990). At that time, there was no through flow of ice along eastern Hudson Strait from the Laurentide interior. The Hudson Strait Ice Stream did not then reach the mouth of Hudson Strait, presumably because it had receded at least a third the way up the strait before the advance. We discuss the Gold Cove event further in the next section, after reviewing the earlier history of the ice streams from marine sediment core data. Field evidence for an ice stream within Hudson Strait includes the convergent flow toward the strait in northern Hudson Bay (Aylsworth and Shilts, 1991), striae and drift transport directions on both the south and north shores of the strait (Clark, 1985; Manley, 1996; Bruneau and Gray, 1997), and the provenance of non-carbonate components of Heinrich layers in Labrador Sea (Barber et al., 1995; Hemming et al., 1998; Barber, 2001).

Sedimentological studies and high-resolution ^{14}C AMS dating of marine sediment cores from the SE Baffin and Labrador shelves and adjacent slopes led to the first major reconsideration of LGM ice extent since the synthesis of Dyke and Prest (1987). Reinterpretation began in Cumberland Sound, where Dyke et al. (1982) had concluded that Late Wisconsinan Laurentide ice had reached only the head of the sound. However, Jennings (1993) concluded that Cumberland Sound was filled by an ice stream until ~ 10 ^{14}C ka BP based on the occurrence of a distinctive black diamicton derived from Cretaceous shale on the floor of the sound just below early Holocene glaciomarine sediment. Barber (2001) similarly concluded that an ice stream in the sound reached the shelf break at LGM and subsequently, including during an advance ~ 11 ^{14}C ka BP. Thin black mud layers in sediment cores from the continental rise off Cumberland Sound are thought to have been derived from glacial erosion of bedrock in the sound. They date close to LGM and are taken as evidence of an ice stream filling the sound at that time (Andrews et al., 1998a, b).

Studies of other marine sediment cores have established the direct linkage between the oscillations of the Hudson Strait Ice Stream and deposition of some of the detrital sediment (largely carbonate) layers, termed Heinrich layers, in the Labrador Sea and more distal

North Atlantic (Andrews et al., 1990, 1991, 1994; Andrews and Tedesco, 1992). The youngest Heinrich layers (H-0 through to H-4) are defined and dated (11, 14.5, 20.5, 27, and 34 ^{14}C ka, in turn) in several sediment cores from sites adjacent to Hudson Strait, but they are of unequal thickness. Layer H-4 is regionally prominent and thick, which suggests that a Middle Wisconsinan advance, or its subsequent collapse, may have been more extensive than the LGM advance (Kirby and Andrews, 1999). Layer H-3 is missing directly off Hudson Strait and hence other sectors of the Laurentide Ice Sheet, such as Cumberland Sound, or the Gulf of St. Lawrence, or another ice sheet need to be considered as possible sources. A variety of evidence indicates that during H-3 (late Middle Wisconsinan), if that event had a northeast Laurentide source, the Laurentide Ice Sheet had a flow configuration similar to that during H-0, the Younger Dryas (see above). The lack of a major carbonate pulse in cores on the southeastern Baffin slope coeval with H-3 (Andrews et al., 1998a, b), all of which contain prominent H-4 and H-2 layers, suggests a parallel between ice dynamics in the Hudson Strait region during H-3 and H-0 (Andrews et al., 1995; Kirby and Andrews, 1999; Pfeffer et al., 1997). The H-0 layer is also thin in the northwestern Labrador Sea (Kirby, 1997). It is uncertain how far the Hudson Strait Ice Stream retreated during either H-3 or H-0 (cf. Laymon, 1991), but clearly each substantial recession of the Hudson Strait Ice Stream did not always produce a strong Heinrich layer at the mouth of the strait. Layer H-2 records a major delivery of sediment to Labrador Sea from Hudson Strait just prior to the global LGM. This event suggests that the Hudson Strait Ice Stream had advanced to the shelf break by that time and perhaps that both the adjacent eastern margins and the central parts of the ice sheet over Hudson Bay and Foxe Basin were close to their maximal configurations. The Hudson Strait Ice Stream presumably terminated somewhere within the strait or in Hudson Bay between H-2 and H-1.

Similar detrital carbonate layers have been long known from the Late Quaternary sedimentary record of Baffin Bay. These sediments probably were derived from calving glaciers in Nares Strait, Jones Sound, and Lancaster Sound; that is from the northwest Greenland, southeast Innuitian, and northeastern Laurentide ice sheets, all of which overlay extensive carbonate bedrock. Recent AMS dating and other considerations (Andrews et al., 1998a, b) suggest that the thickest detrital carbonate layer in Baffin Bay is $\sim$50 ka old, indicating recession of Early(?) Wisconsinan ice that approached or entered the head of Baffin Bay. The youngest Baffin Bay detrital carbonate layer dates to $\sim$12.5 ^{14}C ka. Thus, in this record there is no evidence of major iceberg release close to the LGM (neither H-2, nor H-1); if ice extended into the head of Baffin Bay at that time, as we propose in this paper, calving rates were low. Similarly, there is no Younger Dryas (H-0) layer, nor one that clearly correlates with the well-dated deglaciation of the major marine channels in the Canadian Arctic Archipelago (10–9 ka BP). The latter may be recorded in cores from the Baffin Shelf but these cores are not yet well dated (Andrews et al., 1989). A major lesson from these offshore core studies is that the history of fluctuation of the major ice streams is considerably more complex than that suggested by the terrestrial glacial records, and yet not all major events are seen in the deep sea records.

Groundbreaking studies of lake sediment records on Cumberland Peninsula, Baffin Island, began with the investigation of Wolfe (1994; Wolfe and Hartling, 1996), who provided the first records of sedimentation in apparently biologically productive environments extending through LGM or earlier time in glaciated North America east of Beringia. Six lakes from glaciated, but weathered terrain above the Duval Moraines, which mark the limit(s) of ice advance during the last glaciation on southern Cumberland Peninsula, now have sedimentary records dated by ^{14}C AMS and luminescence methods that start in the Sangamonian (stage 5) and extend into the postglacial without intervening tills (Wolfe et al., 2000). Most and possibly all of these lakes have depositional hiatuses that probably relate to complete freezing over of the lakes during LGM (Steig et al., 1998; Miller et al., 2000), but this response is not unexpected at such localities. These records demonstrate that there is terrain of considerable antiquity on eastern Baffin Island that escaped Late Wisconsinan glaciation.

Cosmogenic exposure dating of the Duval Moraines (Marsella et al., 1999) confirm extensive Late Wisconsinan outlet glaciers in the fiords of Cumberland Peninsula, in agreement with Jennings' (1993) conclusion that Late Wisconsinan ice extended to the mouth of Cumberland Sound. A large series of dates from these moraines cluster into two groups, one at about 24 ^{10}Be ka BP, the other at about 11 ^{10}Be ka BP. Additional cosmogenic dates from the mouth of the sound demonstrate that the outlet glacier extended onto the continental shelf (Kaplan, 1999). It has now been shown that the oldest moraines on Baffin Island between southern Cumberland Peninsula and Hudson Strait are all of Late Wisconsinan age. Although cosmogenic dating of the weathered terrain above the Duval Moraines demonstrates a considerable antiquity (Middle Quaternary), the terrain evidently has been shielded by thin, protective ice for unresolvable portions of the last half million years (Bierman et al., 1999).

In contrast to southern Baffin Island, extensive flights of complex moraine systems occur along eastern Baffin Island from northern Cumberland Peninsula to Lancaster Sound. The oldest moraines on northern Cumberland Peninsula were long thought to date to the Early

Wisconsinan sensu lato (stage 4 to 5d; Pheasant and Andrews, 1973; Dyke et al., 1982). Boulders on parts of this moraine system have produced cosmogenic exposure ages of ~36 ^{10}Be ka BP whereas erratics and glaciated bedrock at a higher elevation were dated to at least 70 ^{10}Be ka BP (Steig et al., 1998). However, moraines just behind these old moraines have produced ^{10}Be ages of 17.7–20.6 ka BP and indicate ice extending to the outer fiords during the LGM. Possibly, therefore, the fiord-mouth moraines on northern Cumberland Peninsula, and perhaps on eastern Baffin Island in general, consist of a complex of Late and earlier Wisconsinan features. A multi-crested lateral moraine system along Sunneshine Fiord on easternmost Cumberland Peninsula may illustrate this case. Boulders from the innermost moraine yield Late Wisconsinan cosmogenic ages. However, boulders from the higher-elevation moraines yield ages older than Late Wisconsinan and paired marine mollusc shells from a delta contacting one of these moraines are >50 ^{14}C ka old (Miller et al., 2000). The general impression arising from these studies, therefore, is that Laurentide ice extended to near the fiord mouths along eastern Baffin Island during both the Early Wisconsinan and the Late Wisconsinan, but the Late Wisconsinan fiord ice was thinner and of lower gradient.

Klassen (1993) placed the Late Wisconsinan Laurentide ice limit on northern Baffin Island to the south of Bylot Island, slightly beyond the position illustrated by Dyke and Prest (1987), and concluded that Late Wisconsinan alpine glaciers on Bylot Island were of similar size to those of the present. He assigned the Eclipse Moraines along the south shore of Lancaster Sound on Bylot Island to the Early Wisconsinan along with younger moraines between the Eclipse position and his Late Wisconsinan limit. That age assignment was based on the non-finite 14C age of marine shells from a raised delta on western Bylot Island, well behind the Eclipse limit and the behind position of a prominent readvance moraine at the mouth of Navy Board Inlet, which separates Bylot Island from Baffin Island. However, recent mapping and extensive ^{14}C dating of the deglaciation of northern Baffin Island (Dyke, 2000; Dyke and Hooper, 2001) indicates that Late Wisconsinan ice filled Navy Board Inlet and that the age of the delta does not, therefore, require a pre-Late Wisconsinan age assignment for the Eclipse Moraines.

These revisions of moraine ages on Cumberland Peninsula and on northern Baffin Island are the most significant reinterpretations of LGM ice extent along the northeastern Laurentide margin. They requires us to consider the fiord-mouth moraines as the most probable LGM features (Fig. 1), rather than the fiord-head Cockburn Moraines (Dyke and Prest, 1987), while acknowledging that parts of that outer moraine system are older. Thus, in this paper we move the LGM margin seaward by the length of a typical fiord, about 100 km. As of yet, however, this age reassignment is not entirely secure. First, there are several natural processes that can render exposure ages too young or too old. Second, the fiord-mouth moraine system between the Duval Moraines in the south and the Eclipse Moraines along Lancaster Sound in the north (Klassen, 1993) are *nowhere* directly associated with ice-contact marine deposits of Late Wisconsinan age, despite the common occurrence of ice-contact raised deltas along them. Furthermore, many raised marine deposits thought in earlier studies to be associated with these moraines have yielded pre-Late Wisconsinan radiocarbon ages (Andrews, 1989; see Dyke et al., 2001 for date list). The Cape Aston (Løken, 1966) and Remote Lake glaciomarine deltas (Ives and Buckley, 1969) remain key sites in this regard. Indeed, nowhere along the Baffin sector of the ice sheet perimeter is there a moraine securely associated with a radiocarbon-dated marine fossil older than 10 ^{14}C ka BP and younger than 25 ^{14}C ka BP.

3. Interior configuration

LGM ice surface elevation can be read directly only where the ice terminated on land, such as on Banks Island (Vincent, 1989), along the mountain front from the Yukon coast to the point of full coalescence with Cordilleran Ice (Prest et al., 1968; Dyke and Prest, 1987), around the lobate margins of the southern perimeter (Mathews, 1974; Clark, 1992), and along certain fiords and channels in the eastern Arctic (e.g. Dyke et al., 1982; Clark, 1988) (Fig. 4). In most places within the Laurentide Ice Sheet limit topographic relief is far short of former ice thickness but a few overtopped mountains provide useful minimum ice thicknesses. All that can be concluded from direct mapping is that the vast interior region of the ice sheet, generally the part that was more than about 1000 km behind the margin, lay more than 2000 m above present sea level (Fig. 4). However, isostatic correction would shift portrayed ice-surface contours inland.

Otherwise, the best estimates of central ice thickness at LGM are provided by geophysical models that fit computed RSL histories to radiocarbon-dated postglacial RSL curves and to warped glacial lake shorelines by a process of fine-tuning the ice load history (e.g., Peltier, 1996). The precision of such calculated ice thicknesses is limited by the separation in time between LGM and the earliest postglacial shorelines in the central glaciated region. This separation is about 8–10 ka, or about 4–5 half lives of the isostatic relaxation process (Dyke and Peltier, 2000). Thus the "memory" in the sea-level data of LGM ice thickness in central Laurentide (Hudson Bay and Foxe Basin) RSL curves is only 6–12%. This amount might be sufficient to give confidence in inverse

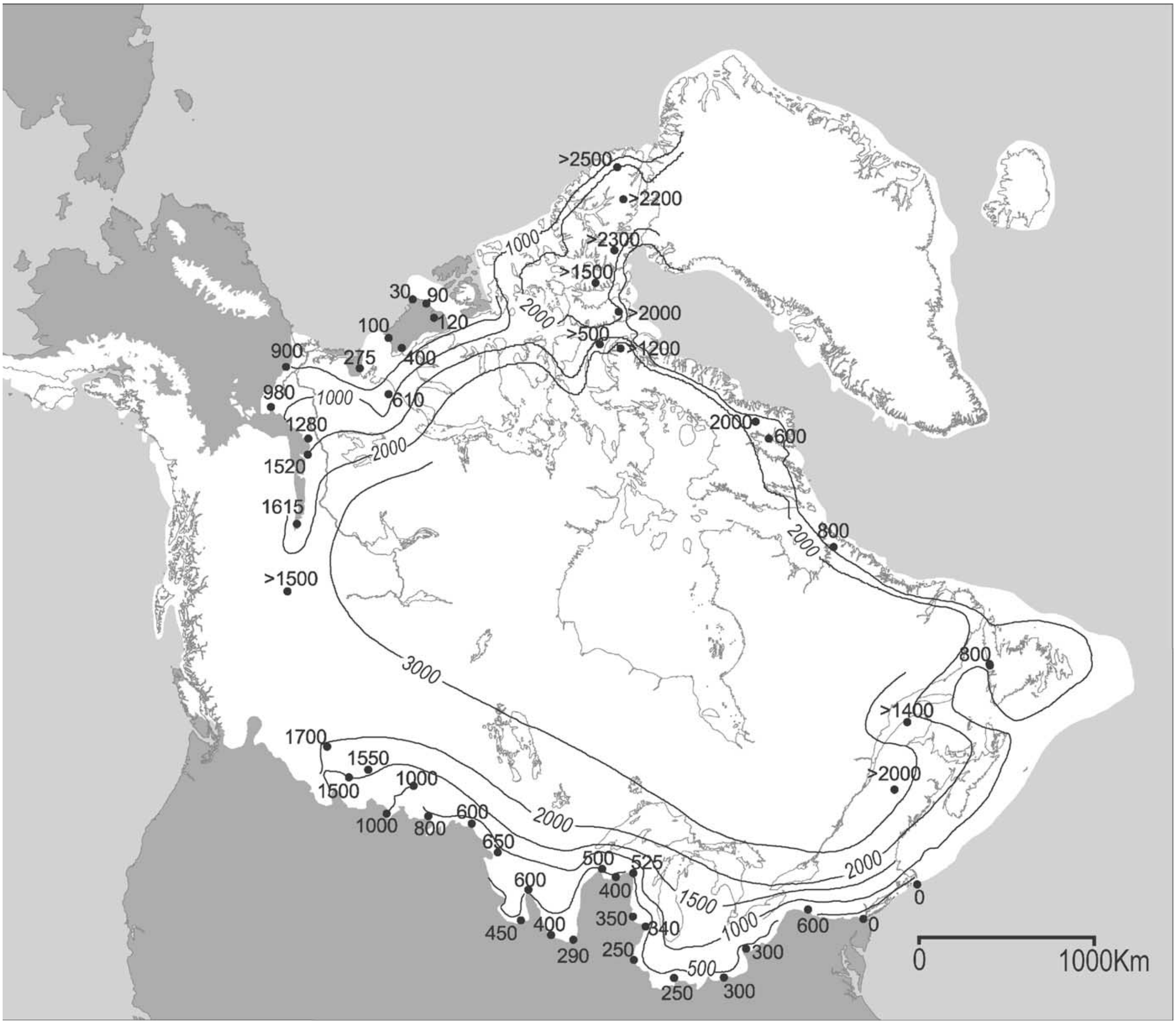

Fig. 4. Ice surface contours based on elevations along the LGM ice margin and topographic high points overridden by ice.

calculations of ice thicknesses if central RSL curves or shoreline deformations were precisely defined. Unfortunately, this is not presently the case (Dyke and Peltier, 2000) and providing the best possible constraints on central RSL history should be a priority for future research if Laurentide ice volume history and its contribution to global sea-level history are to be properly understood. Because of a lack of attention to the geological aspects of this problem over the last 20 years, our ability to model central ice load history is considerably more advanced than our ability to provide good geological constraints on the models. A qualitative ice-thickness history is currently more tractable. We now review that history and see, for the central and eastern Laurentide Ice Sheet, how it might relate to the Heinrich events at the mouth of Hudson Strait.

Relative central ice thicknesses can be assessed from the flow patterns at LGM. However, any reconstructed LGM flow pattern will depend on which set of assumptions are accepted in interpreting and in assigning ages to shifting regional ice flow patterns, that is to sequences of events that are intrinsically undatable. This discussion considers first the central Laurentide regions and then the peripheral Appalachian and Innuitian regions.

3.1. Laurentide region

The Laurentide Ice Sheet comprised three major sectors, each with its system of ice domes and divides that probably reflect the pattern of inception and buildup of the ice sheet during the last glacial cycle

(Dyke and Prest, 1987). Major centres of outflow were located over Quebec–Labrador, Keewatin, and Foxe Basin, thus forming a system surrounding Hudson Bay, where the ice surface was generally lower. Although each of these domes coincides with a postglacial uplift centre (e.g., Dyke, 1996), ice divides probably shifted throughout each glacial cycle. Shifting of divides is required by the strong west-to-east asymmetry of deglaciation, if nothing else (Dyke and Prest, 1987). Boulton and Clark (1990) portrayed dynamically shifting ice divides based on satellite image analysis of glacial bedforms. Similarly Dyke et al. (1992) recognized abrupt changes of flow pattern in the northern Keewatin Sector and showed that some drumlin fields, which are normally assigned to deglaciation, are relict terrains that belong to older events and survived unmodified beneath later cold-based ice. However, the sequence of shifting flows is well documented by field measurements only for the Labrador Sector, which presages the kind of sequences likely to be found in the other sectors.

Within the Labrador Sector the sequence of shifting flows is documented by hundreds of measurements of striae taken from bifaceted or multifaceted rock outcrops, the youngest of which are readily assigned to deglaciation (Veillette et al., 1999, for review). According to Veillette et al., the Labrador Sector ice divide migrated northwestward from an early buildup position over the Quebec Highlands toward Hudson Bay until the main ice dome lay close to or over the postglacial uplift centre near Richmond Gulf. Two sector-wide reorganizations of flow followed prior to regional deglaciation. These reorganizations are the only firm indications available of ice-thickness changes close to and immediately following LGM time. In the first, the main centre of outflow shifted from the vicinity of the uplift centre to a point near Schefferville, 900 km to the east (Fig. 5). This shift records a large lowering of the ice surface over eastern Hudson Bay relative to that over Schefferville. In the second, the headward part of a regional southwesterly outflow from Schefferville was captured by the up-stream propagation of a regional flow that converged northward into Ungava Bay. Veillette et al. suggested that this capture event correlates with the Gold Cove Advance of Labrador Ice across Hudson Strait. This advance is the farthest known Pleistocene northward advance of Labrador ice, evidently a unique event with an ice-flow trajectory that is otherwise unexplained. The Gold Cove Advance culminated $\sim 10\ ^{14}C$ ka BP and, as mentioned above, it must have followed upon a major recession of the Hudson Strait Ice Stream during the Younger Dryas (H-0). If correlation of the Ungava flow with the Gold Cove Advance is correct, the preceding flow reorganization may have resulted from the H-1 event $\sim 14.5\ ^{14}C$ ka BP as seen in the sediment record adjacent to the mouth of Hudson Strait. The large eastward displacement of the outflow centre away from Hudson Bay at that time probably indicates drawdown of ice over the entire central part of the ice sheet by hundreds of metres.

If H-0 and H-1 events can thus be recognized in the ice-flow sequence, should earlier Heinrich events not also be evident? H-2 may simply record advance of the Hudson Strait Ice Stream beyond the sill at the mouth of the strait. Its correlative in the ice flow record would thus be the culmination of the shift in the ice divide from the Quebec Highlands to the uplift centre, just as H-2 correlates generally with the advance to LGM along other segments of the margin (Fig. 6) (Clark, 1994). Perhaps the ice configuration just prior to H-2 was a single dome over Hudson Bay and H-2 caused a draw down of this dome, thus producing the three-dome configuration (Licciardi et al., 1998). However, that possibility is not currently supported by the striation record.

H-3 and H-4 are not evident at present in the circum-Hudson Bay ice flow record. This fact may reflect inadequate documentation of the ice-flow record, because the preservation of each older set of striae is diminished by about an order of magnitude. We suggest instead that the H-3 sediment event, if it has a Laurentide source, may relate to the Middle Wisconsinan opening of the Gulf of St. Lawrence, which is well documented on Anticosti Island (Figs. 3 and 6; Gratton et al., 1986; St-Pierre et al., 1987). If Hudson Bay maintained a thick ice cover during H-3 and the Hudson Strait Ice Stream played little or no role in the H-3 sediment pulse, H-3 would not have a correlative in the circum-Hudson ice flow sequence. The alternative possibility that the buildup phase of the current striation record may start with culmination of H-3 is less probable. That interpretation would require that H-3 effected a draw down of ice over Hudson Bay that was even larger than the H-1 draw down. Such a scenario is not in agreement with the fact that H-3 is missing off Hudson Strait.

Kleman et al. (1994) and Clark et al. (2000a, b) have proposed alternative reconstructions for the evolution of the Labrador Sector. The main difference between these reconstructions and that of Veillette et al. (1999) revolves around the origin and age of the sharp intersection between the general northward flow pattern toward Ungava Bay and the westward to southward radial flow pattern covering a large part of Quebec. This intersection is commonly referred to as the horseshoe-shaped Labrador ice divide and is the feature attributed by Veillette et al. (1999) to the Gold Cove capture event (Fig. 5).

Kleman et al. concluded from airphoto interpretation that the intersection marks the limit of migration up the radial flowlines of the subglacial melting boundary

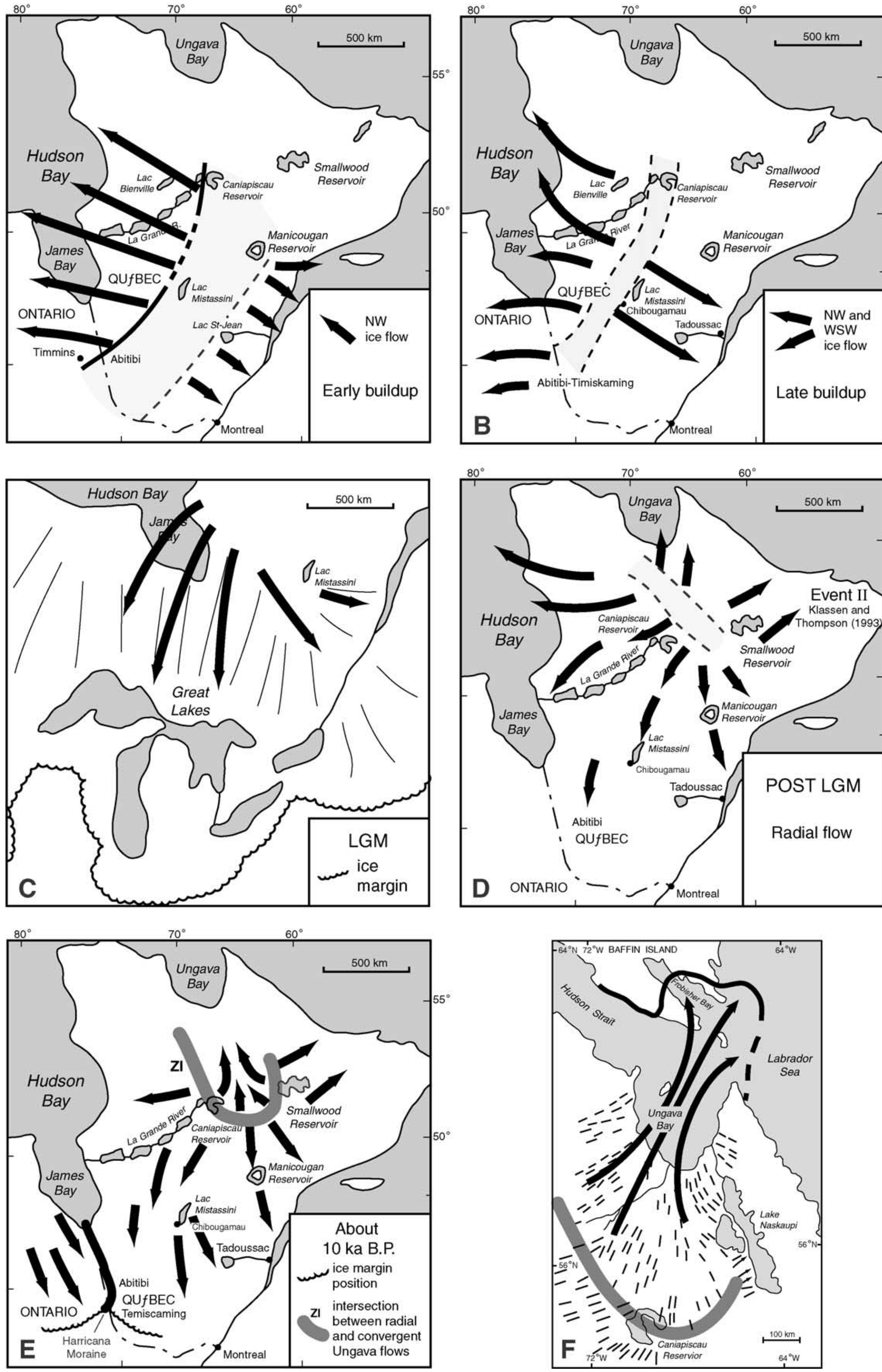

Fig. 5. Ice flow evolution of the Labrador Sector of the Laurentide Ice Sheet based primarily on striation measurements (modified from Veillette et al., 1999).

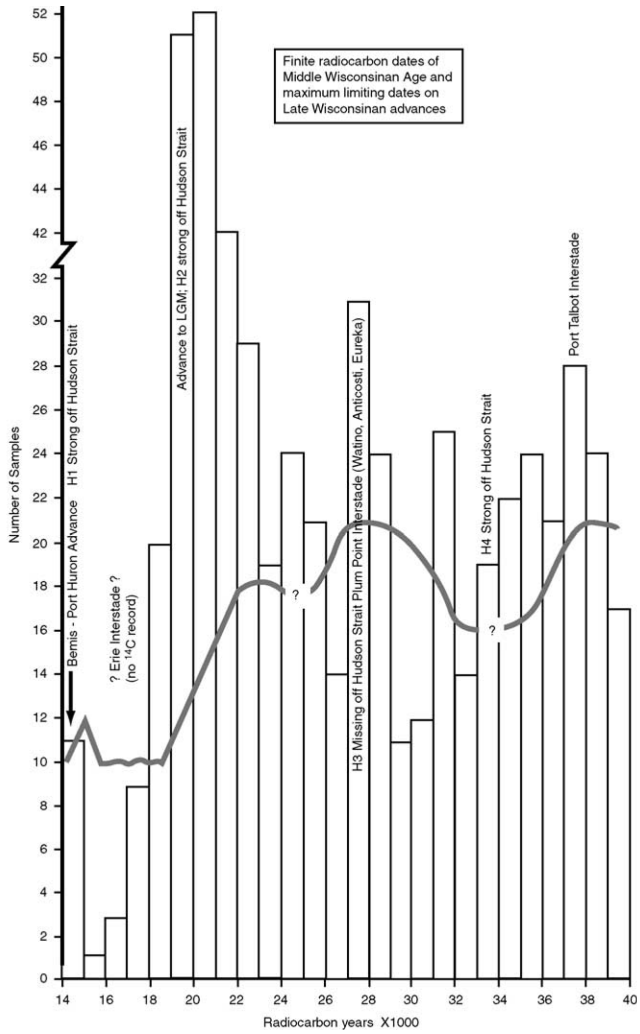

Fig. 6. Distribution of all available dates limiting the chronology of Late Wisconsinan ice advance and recession in relation to the major interstadial stratigraphic units and Heinrich events. The solid line approximates the movement of the southern Laurentide margin.

during deglaciation. They further proposed that northward from the intersection, entirely cold-based ice retreated toward a centre in Ungava Bay; that is across an older, or relict, Ungava Bay drumlin field and associated eskers, which they assigned to a pre-Wisconsinan deglaciation. However, intersecting drift lineaments and the striation record north of the intersection clearly show that the Ungava flow is overprinted on the radial flow (Veillette et al., 1999), as was already concluded by Hughes (1964) and Klassen and Thompson (1993).

Clark et al. (2000a, b) presented a reconstruction of a large part of the Labrador Sector based on interpretation of ice flow sets from radar imagery. A full discussion of their interpretation is beyond the limitations of this paper, but the following of their conclusions are most relevant to our topic. They recognized a flow set which is parallel to the buildup flow of Veillette et al. (Fig. 5, panel B) and suggested that it required that the 18 ^{14}C ka BP ice divide of Dyke and Prest (1987) was located too close to Hudson Bay. Their interpretation, therefore, invokes a substantially thinner central Laurentide ice cover that might be inferred from the Dyke and Prest or Veillette et al. reconstructions. Like Kleman et al., they concluded that the radial flow is younger than the Ungava flow, an interpretation already contradicted by the striation record, but they placed the Ungava flow within the Late Wisconsinan flow sequence rather than in a previous deglaciation. Like Veillette et al., they correlate the Gold Cove Advance with a northward flow into Ungava Bay. However, instead of the prominent flow pattern that originates at the horseshoe-shaped intersection, they correlate the advance with a much more extensive northward flow, one not previously recognized and originating south of the limit of their mapping at 52°N. If their interpretation is correct, the northward flow to the Gold Cove limit (63°N) was >1200 km long. Because the southern ice margin had retreated to near the Quebec north shore of the Gulf of St. Lawrence by about 14 ^{14}C ka BP (see above) and must have been at or near the Quebec North Shore Moraine system (50.5°N) at 10 ^{14}C ka BP (Dyke and Prest, 1987), the putative northward flow originated from an ice divide that was <160 km from the contemporaneous southern margin of the ice sheet. There is no observed field evidence that such a flow actually occurred. If it did, it must have occurred far earlier than 10 ^{14}C ka BP and in all likelihood before deglaciation of the Gulf of St. Lawrence.

The Wisconsinan ice flow sequence of the Keewatin Sector is not as well documented as that of the Labrador Sector. It had been argued that the great length of the Dubawnt dispersal train, which extends about 1000 km from central Keewatin into Hudson Bay and corresponds to the youngest ice-flow pattern, required a stable flow pattern and hence a stable position of the Keewatin Ice Divide throughout the last glacial stage (Shilts et al., 1979). New striation mapping and till stratigraphy (McMartin and Henderson, 1999), however, show that the Keewatin Ice Divide or divide system was mobile during the course of the last glaciation, shifting position by at least 200 km, which is in general accordance with the suggestions of Boulton and Clark (1990).

3.2. Appalachian region

The interpretation that the LGM ice extent in Atlantic Canada was as extensive as any previous glaciation in the region re-opens the old question about the extent of Laurentide versus Appalachian ice at that time (Prest and Grant, 1969; Brookes, 1970; Prest, 1970, 1984). The most recent portrayals of LGM ice configuration (Stea et al., 1998) indicate that, although Laurentide ice extended through the Gulf of St. Lawrence to the mouth of the Laurentian Channel, the Canadian sector west of the gulf was covered by an Appalachian ice complex spreading from local centres on land and on the shallower part of the continental shelf. This interpretation is based largely on lithic composition of Late Wisconsinan Scotian Shelf till, which indicates derivation from the Magdalen Shelf redbeds. Stea et al. suggest that this material was delivered from a local ice centre on the Magdalen Shelf, though it is not clear why a Laurentide flow path would be less probable.

3.3. Innuitian region

In the current view, the probable flow pattern of the Innuitian Ice Sheet at the LGM, is entirely compatible with the pattern of postglacial uplift. The ice sheet was comprised of an alpine sector in the northeast and a lowland sector elsewhere. In the former, ice flowed outward from ice divides atop the mountains of Axel Heiberg and Ellesmere Islands, which are heavily covered today by large remnants of Innuitian ice. This ice filled the large inter-montane basin that is bisected by Eureka and Nansen sounds to an estimated depth of about 1000 m. The central ice evidently did not thicken sufficiently to flow radially outward across the surrounding mountains, probably because of the efficient drainage that occurred along the two sounds (Bednarski, 1998; O'Cofaigh et al., 2000). Sustained outflow along Baumann Fiord and along Massey Channel is indicated by a long dispersal train of granite erratics extending from southeast Ellesmere Island toward the Arctic Ocean (Lamoureux and England, 2000; O'Cofaigh et al., 2000). Maximal Innuitian ice thickness may have been maintained for only a brief period, for as discussed above, pre-LGM plant remains and ice-transported marine shells in both the Eureka Sound and Nares Strait regions date to 25–20 ^{14}C ka BP (Fig. 3). Outflow from the central part of the Innuitian Ice Sheet, which lay over the now partly submerged lowlands, set up a large ice stream in Wellington Channel, which in our present reconstruction extended to the Eclipse Moraine limit in Baffin Bay via the Lancaster Sound Ice Stream. The lowland part of the Innuitian Ice Sheet probably included local ice divides and domes over the present islands separating ice streams in each of the large marine channels (e.g., Dyke, 1999). Judging by the extant regional ice caps, island-centred domes were probably about 1000 m thick and intervening ice streams somewhat thicker but slightly lower.

4. Summary and overview

The reconstruction that emerges from this review of the LGM of the Laurentide and Innuitian ice sheets is surprisingly simple. It is surprising in that much of the research over the last 30 years focused on hypotheses that essentially called for idiosyncratic regional responses of various segments of the ice sheet margin at the scale of the isotopic stages. On that view, extensive Arctic glaciations were confined to the early stages of a glacial cycle (e.g., stage 4 or 5d), when moisture could enter polar regions with ease. In contrast, there was minimal LGM ice in the Canadian Arctic, because it was in the precipitation shadow of the Laurentide Ice Sheet. Arctic glacier responses were thus seen to be generally out-of-phase with responses along the southern margin (e.g. Miller and Dyke, 1974; England, 1976). Although elements of these hypotheses that pertain to earlier events of isotopic stages 3–5 may yet prove to be valid, the weight of evidence regarding the LGM period is that ice built to its maximum everywhere east of the Cordillera at about the same time, some 24–21 ^{14}C ka BP. A parallel shift of interpretation can be seen in studies of the European Arctic (e.g., Boulton, 1979; Mangerud et al., 1998).

The last Laurentide ice buildup started from an interstadial minimum ice sheet extent at $\sim$27–30 ^{14}C ka BP. The outline of the interstadial ice approximately followed the margin of the Canadian Shield. At the same time, ice in the Cordillera (Clague, 1989) and probably in the High Arctic islands was little more extensive than present. Taking the current suite of apparently reliable radiocarbon dates at face value, the Laurentide ice margin may then have reached its maximum as early as 24 ^{14}C ka BP in the Mackenzie Lobe and 23 ^{14}C ka BP in the Great Lakes region, but was still building to its southwestern limit until $\sim$20 ^{14}C ka BP. Similarly the Innuitian ice buildup seems to have culminated in the east after 20 ^{14}C ka BP. Thus, in contrast to the prevalent view of a decade ago, the entire ice sheet system east of the Cordillera seems to have responded to uniformly applied changes of climate. In contrast, the Cordilleran Ice Sheet only began to grow at $\sim$20 ^{14}C ka BP and did not reach its maximum extent until 14.5–15.5 ^{14}C ka BP (Clague and James, this volume). Thus, the Laurentide Ice Sheet reached its maximum extent before both the last minimum in Northern Hemisphere insolation at $\sim$18 ^{14}C ka BP (21 cal ka BP) and the LGM, whereas the Cordilleran Ice Sheet reached its maximum well after the insolation minimum and the LGM, suggesting additional controls on the history of these ice sheets during the late Wisconsinan.

Dyke and Prest (1987) suggested that the out-of-phase behaviour of Laurentide and Cordilleran ice sheets could be explained by the feedback effects of the latter on regional moisture transport. Thus, the Laurentide ice margin advanced across the Interior Plains just prior to 20 ^{14}C ka BP when an ice-free Cordillera allowed easy passage of Pacific moisture to the ice sheet, which was advancing as insolation decreased. Conversely, the full development of the Cordilleran Ice Sheet by 14 ^{14}C ka BP may have increased aridity over the western Laurentide Ice Sheet and promoted the onset of general recession at about that time. Pollard and Thompson (1997) advanced a similar hypothesis with the elaboration that final growth of the Laurentide Ice Sheet changed atmospheric circulation so as to cause additional moisture delivery to the region of the Cordilleran Ice Sheet, thus promoting its growth. Further model simulations will address these issues (S. Hostetler, pers. comm., January 2001).

The newly demonstrated synchrony between millennial-scale Late Wisconsinan fluctuations of the well-dated Laurentide ice margin in the eastern Great Lakes region and temperature changes in the North Atlantic region (Lowell et al., 1999; Clark et al., 2001) suggests that this margin may have responded to climate changes in the North Atlantic. Results from one climate model simulation, for example, show that an increase (decrease) in moisture balance occurred along the southern margin, but not along the rest of the margin, in response to cold (warm) events in the North Atlantic (Hostetler et al., 1999). These results suggest that strong regional climate feedbacks may have led to regionally distinctive glacial histories on millennial timescales. Unfortunately, however, fluctuations of other segments of the ice margin are not yet well dated to allow us to identify if and when such responses might have occurred.

Except in the Atlantic Provinces to Lake Michigan basin, there is little firm radiocarbon evidence of regional recession prior to 14 ^{14}C ka BP. Dated recession prior to 14 ka BP occurred mainly in deep water or involved only the fringe of terrestrial ice. If the correlation drawn by Veillette et al. (1999) between H-1 at 14.5 ^{14}C ka BP and the largest mapped flow reorganization of the central Laurentide Ice Sheet is valid, it is likely that the drawdown of central ice surface that accompanied this event would have promoted general Laurentide deglaciation by decreasing flow to the marginal zone thereafter.

Our review thus demonstrates that the period of maximum ice extent in North America generally encompasses the interval from $\sim$24/21 to 14 ^{14}C ka BP, or considerably longer than the duration of the LGM as defined by EPILOG (Mix et al., 2001). The LGM is defined as occurring during a period of low global sea level as well as during a time of relative climate stability as suggested by regional climate records. The interval of advance of much of the Laurentide Ice Sheet to its maximum extent (between $\sim$27 ^{14}C ka BP and $\sim$24 ^{14}C ka BP), however, does coincide with a suggested interval of rapid fall in global sea level to near LGM levels (Lambeck et al., submitted). Insofar as the Laurentide Ice Sheet was the largest contributor to sea-level change during the last glaciation, this correlation indicates that an increase of ice volume must have accompanied the ice margin advance to its LGM limit. That is, ice over Hudson Bay experienced no appreciable thinning and instead probably thickened during that interval. The long interval of generally maximal configuration and the already substantial size of the interstadial ice sheet suggests that the ice sheet may have attained isostatic equilibrium throughout most of its extent and that it may have had a surface geometry and mass balance nearly in equilibrium with the LGM climate.

Records from continental shelves off Australia (Yokoyama et al., 2000) and Vietnam (Hanebuth et al., 2000) suggest that global sea level began to rise at 16.5 ^{14}C ka BP, which is 2.5 ^{14}C ka before significant recession of the North American ice margins. If the Laurentide Ice Sheet was contributing to initial sea-level rise, then the period 16.5–14 ^{14}C ka BP was one wherein central ice thinning dominated over marginal recession. The southern ice margin may have remained near its limit during initial deglaciation in response to the regional climate feedbacks from cold events in the North Atlantic that maintained an extended ice margin during a time of rising insolation. Moreover, the drawdown of the central part of the ice sheet that caused H-1 may have caused the coeval advance along the southern margin referred to as the Port Bruce Stadial (Dreimanis and Karrow, 1972; Clark, 1994). Subsequent sea-level rise continued at a relatively constant rate until an abrupt jump in sea level at $\sim$12 ^{14}C ka BP (Fairbanks, 1989). There is no known geological record indicating that this event was derived from the Laurentide Ice Sheet (Clark et al., 1996).

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