An outline of North American Deglaciation with emphasis on central and northern Canada

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Introduction

The Late Wisconsinan North American ice sheet complex consisted of three major ice sheets: the Laurentide Ice Sheet, which was centreed on the Canadian Shield but also expanded across the Interior Plains to the west and south; the Cordilleran Ice Sheet, which inundated the western mountain belt between the northernmost co-terminus United States and Beringia (unglaciated Yukon Territory and Alaska); and the Innuitian Ice Sheet, which covered most of the Canadian Arctic Archipelago north of about 75°N latitude. The ice cover over Newfoundland and the Maritime Provinces of Canada is usually referred to as the Appalachian Ice Complex, because ice flowed out from local centres rather than from the Canadian Shield. All of the peripheral ice sheets were confluent at the Last Glacial Maximum (LGM) with the Laurentide Ice Sheet, and the Greenland Ice Sheet was confluent with the Innuitian Ice Sheet. The nucleus of this complex, the Laurentide, comprised three major sectors, the Labrador Sector, the Keewatin Sector, and the Baffin Sector, named for areas of inception and probable areas of outflow at LGM and located respectively east, west and north of Hudson Bay (Dyke *et al.*, 1989). Previous continental scale syntheses of ice recession are given by Prest (1969), Bryson *et al*. (1969), Denton & Hughes (1981) and Dyke & Prest (1987a-c).

This paper presents revised maps of North American deglaciation at 500-year and finer resolution (Dyke *et al.*, 2002b, c; 2003). These maps represent an updating of a series prepared nearly two decades ago for the INQUA 1987 Congress (Dyke & Prest, 1987a-c). A revised reconstruction for the LGM was presented as part of the EPILOG review (Dyke *et al.*, 2002a) and is retained here with minor changes. The timing and detail of the current revision is a response to the need for more accurate and more detailed continental ice-margin histories to support the next generation of efforts in glacial isostatic adjustment modeling (Thorleifson, 2001), as well as to support other reconstructions of Late Quaternary Palaeoenvironments continuing at the Geologicalal Survey of Canada and elsewhere.

Significant changes of interpretation of ice margin history have appeared since the map series of Dyke & Prest (1987a-c). The net effect is a young shifting of the ages assigned to ice-marginal features. This change is the result chiefly of improvements in radiocarbon dating with the advent of Accelerator Mass Spectrometry (AMS) dating methods and the consequent 'retirement' of hundreds of earlier conventional radiocarbon dates, discussed further below. Dates of deglaciation are thus now seen as being typically 500-2000 years younger than they were previously in most areas. With respect to the LGM reconstruction, changes in age are even more significant, because large regions within Arctic and Atlantic Canada, where glacial deposits had been assigned a pre-Late Wisconsinan age, are now assigned to the Late Wisconsinan.

Because regional chapters in this volume deal with glacial geology in considerable detail, The discussion in this chapter will be restricted to a broader overview and highlight aspects of continental deglaciation, such as the opening of the corridor between Laurentide and Cordilleran ice, the expression of Younger Dryas cooling in the continental glacial record and the deglaciation of Hudson Bay. Changes in ice-flow patterns and ice-divide migrations are not explicitly dealt with. Stratigraphical terminology (e.g., till names, formation and member names) is largely avoided, the major features of interest being end moraines, glacial lakes and post-glacial seas.

Construction of palaeogeographic maps

The glacial geomorphology of the continent is the fundamental constraint on any interpretation of North American deglaciation. Two specific patterns, those of end moraines and other ice-marginal features and those of ice flow features, are of primary importance. These broad patterns have been known for decades from large-scale compilations such as the Glacial Map of Canada (Falconer *et al.*, 1958; Prest *et al.*, 1968; Dyke & Prest, 1987c for moraines and their names) and the Glacial Map of the United States East of the Rockies (Flint *et al.*, 1959). The fundamental assumption in interpreting these patterns, common to the present and previous efforts (e.g., Prest, 1969; Denton & Hughes, 1981), is that where ice margins are not recorded by mappable features, these margins trended normal to ice flow directions. This assumption limits the range of lateral correlations that may reasonably be considered between marginal features, and thus accounts for the fundamentally similar patterns seen in most reconstructions of Laurentide deglaciation over decades, namely recession back to the three major centres of Keewatin, Québec-Labrador, and Baffin Island.

Fig. 1. Place and feature names referred to in text.

An underlying further assumption is that the ice-flow features, and associated meltwater features such as eskers, were formed in a time transgressive sequence, forming later the further inward during deglaciation. This assumption is not always valid, and hence must be abandoned locally. The assumption is incorrect wherever a phase of warm-based ice flow that generated bedforms, be they striae or drumlins, was followed by a final phase of cold-based ice flow in a different direction. The cold-based flow does not alter the older bedforms, which survive as local to regional, or possibly sub-continental, relict landscapes. Local to regional scale relict landscapes have been recognised in Arctic Canada from cross-cutting relationships (Dyke *et al.*, 1992).

The second major constraint in interpreting the history of deglaciation is the set of relevant numerical age determinations. New 'exposure dating' studies have provided important tests of models of terrain age in the 'weathering zones' along the mountainous eastern seaboard (e.g., Marsella *et al.*, 1999; Gosse *et al.*, 1995; Steig *et al.*, 1989; Clark *et al.*, 2003) and have established the Late Wisconsinan age of final coalescence of the Laurentide and Cordilleran ice sheets in Alberta (Jackson *et al.*, 1997). However, only radiocarbon age determinations are sufficiently precise at this time to allow mapping of the deglaciation sequence at centennial (or sub-millennial) resolution. Locally, the chronological database is supplemented by occurrences of early post-glacial tephra of known radiocarbon age and by 'varve dates' tied into the deglaciation sequence by radiocarbon dates. In the latter case, the varve dates are converted herein to the radiocarbon time scale. The radiocarbon time scale is used throughout this paper in the *kilo anni* (ka) BP notation. Where it is necessary to refer to calendar time, the expression 'cal ka BP' is used.

All known age determinations that are potentially useful in constraining the chronology of deglaciation, including those on readvances, were assembled in a geo-referenced database and are plotted on the palaeogeographical maps. The web-based versions of these maps will be interactive with the database and digital versions of both the maps and database are available (Dyke *et al.*, 2003a). Unfortunately, radiocarbon dates are reported in various ways (uncorrected or corrected for isotopic fractionation; 'corrected' (i.e., normalised) for fractionation to either –25 ‰ $\delta^{13}C_{PDB}$ or to 0 ‰ $\delta^{13}C_{PDB}$; corrected or not corrected for reservoir effects), but it is not always clear which of these types of date is reported, or this may be inconsistently stated for a given date from report to report. In the database, therefore, all dates are expressed as conventionally-normalised ages; that is, the ages are normalised for carbon isotope fractionation to the terrestrial standard of –25 ‰ $\delta^{13}C_{PDB}$. In the many cases where assumptions had to be made as to the form of the reported date on marine shells, these assumptions are stated under 'comments'. Where assumptions are incorrect, the normalised dates are in error by about 400 years.

Regional reservoir corrections were then applied to marine shell dates. These corrections range from –800 years for the Pacific coast and for the Champlain Sea (a former sea that occupied the Ottawa-St Lawrence Lowland) to –450 years for the south-western Gulf of St Lawrence. Except for the Champlain Sea, the corrections are based on a new set of 270 age determinations on live-harvested molluscs collected prior to nuclear bomb contamination (Dyke *et al.*, unpublished data). The Champlain Sea correction is based on extrapolation inland of the gradient of reservoir ages in the Gulf of St Lawrence from –450 years in the outer gulf to –610 years in the St Lawrence Estuary and on the expectation that the carbonate rock substrate of the Champlain Sea would have augmented the local reservoir effect. Note that previous interpretations of North American deglaciation have, as a whole, applied a uniform –400-year reservoir correction to marine shell ages or have not explicitly considered the reservoir effect.

Numerous radiocarbon dates, including many used in previous reconstructions, are now considered spuriously old. Rejected dates are compiled in a separate spreadsheet within the database, with a brief reason for 'retirement' given for most. Dates on the total organic carbon fraction of marine sediments are universally rejected, because they are thousands of years older than mollusc shell or foraminifera dates from the same sediments, but they are not here compiled. Dates in the Middle Wisconsinan age range (>25 ka BP) on these materials from the south-eastern Canadian Continental Shelf had largely formed the basis for assigning the uppermost till of that region to the early Middle or Early Wisconsinan (e.g., Piper *et al.*, 1990). Similarly, comparisons of AMS dates on terrestrial macrofossils from lake sediment cores with dates, either conventional or AMS, on bulk lake sediment have repeatedly shown the bulk dates to be much too old. Differences are largest in the organic-poor basal parts of sediments, particularly, but not exclusively, where sediments are calcareous (e.g., Richard *et al.*, 1997). Furthermore, both AMS and conventional dates on aquatic moss from lake sediments have been shown to yield spuriously old results (e.g., MacDonald *et al.*, 1987; Gajewski, 1995). Hence, lake sediment dates constitute the largest group of rejected ages in this database. Some of those dates that are still retained will probably be found to be too old in future studies, but we cannot at present afford to reject all dates in this class.

Another large group of problematically old dates are on marine shells, after the reservoir corrections mentioned above are applied. These dates are of two sorts: (1) bulk dates on samples that contained shells of a wide range of ages, typically post-glacial and interstadial, have yielded 'blended ages' that are problematically early, or are at least atypical, for their locations; and (2) deposit feeding molluscs yield ages that are substantially older (ca 0.1-2 ka) than do suspension feeding molluscs from the same sites, with the difference being largest on the more calcareous substrates and negligible on carbonate-free substrates (Dyke *et al.*, 2003b; England *et al.*, 2003; Dyke *et al.*, unpublished data). This deposit feeding effect is presumably due to the mollusc's use of pore water, which contains some portion of dissolved carbonate from the sediment, in assimilating its shell, rather than bicarbonate in freely circulating seawater. Unfortunately, among these deposit feeders is the genus, *Portlandia*, which is able to colonise rapidly accumulating ice-proximal mud. They have, therefore, been numerously and preferentially dated because of this environmental affinity. The genus, *Macoma*, is also in this group and it has been even more commonly dated. All suspected 'blended' shell ages are retired, and all evident cases of at least strong deposit feeder effects have been removed from the database. Some that are retained may be shown to be problematic in future work.

In summary, the maps presented here and in Dyke *et al*. (2003) were constructed in the following steps: (1) the digital palaeogeography (ice margins and shorelines) from Dyke & Prest (1987b), as amended by Dyke (1996) and Dyke *et al*. (1996, 2002a), were plotted together with the culled radiocarbon database discussed above as applicable for each time slice, with dates pertaining to ice advance in a different colour; (2) amendments to ice margins and shorelines were drawn as required by recent regional interpretations, by new radiocarbon dates, or by the rejection of dates previously used; (3) the revised maps and the radiocarbon database were distributed for comment to individuals and groups in Canada and the Unites States, and revisions were made on the basis of reviews returned; and (4) additional time slices were added to provide a 500-year resolution.

Extent and quality of age control

The radiocarbon database provides direct and minimum-limiting dates on ice recession as well as maximum-limiting dates on readvances, including the advance to the Late Wisconsinan glacial limit. At time of writing, the database contains 4035 entries, 385 of which pertain to readvances (Figs 2 and 3). Because of the number of original sources, it is impossible to cite most of them here. The database may be consulted for this purpose (Dyke *et al.*, 2003a). Dating control on ice margins older than 14 ka BP (Fig 2) is meager, but this mainly reflects the fact that only small

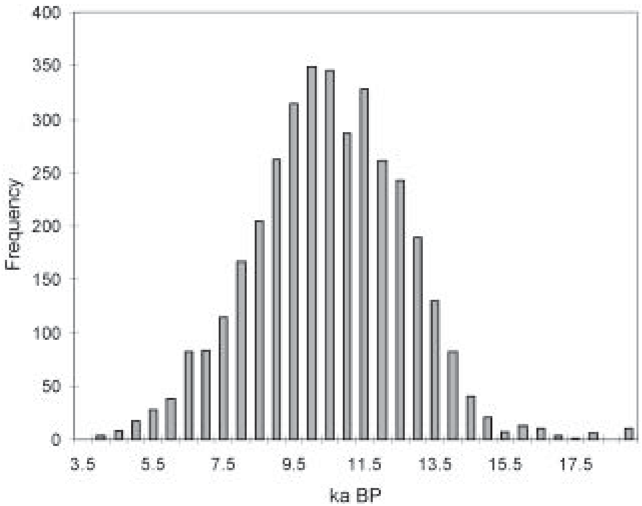


Fig. 2. Histogram of direct and minimum-limiting radiocarbon dates on ice recession for North America, including Greenland.

areas became ice free prior to that time. The increase in frequency of dates from 14 to 10 ka BP reflects the gradual increase in deglaciated area (Fig. 5). However, deglaciated area continued to increase at a nearly constant rate until 6 ka BP. The sharp drop in frequency of dates between 10 and 6 ka BP reflects the relatively small effort that has been applied to establishing deglacial chronologies in remote boreal and inland tundra regions of Canada.

About 40% of the radiocarbon dates are on marine molluscs. Indeed, the great majority of direct dates on ice-marginal positions are from sites where moraines or meltwater features have been followed into correlative shell-bearing marine deposits, such as commonly occur in ice-contact, marine-limit deltas. Opportunities to directly date ice-marginal features, landward of the limit of post-glacial marine incursion, are exceedingly rare, being limited to those few localities where ice readvanced across living or recently dead vegetation or where the varve chronologies of proglacial lakes can be tied into the radiocarbon time scale. The key varve-dated sites are in the basins of glacial Lake Hitchcock in New England (Antevs, 1922; Ridge & Larsen, 1990; Ridge *et al.*, 1999, 2001; Miller & Spear, 1999; Rittenour *et al.*, 2000; Brigham-Grette *et al.*, 2000), glacial lakes Barlow and Ojibway in Québec and Ontario (Antevs, 1925; Hughes, 1965; Hardy, 1977, 1982; Veillette, 1994; Paulen, 2001), Lake Superior (Mothershill, 1984; Thorleifson & Kristjansson, 1993), glacial Lake Agassiz (Elson, 1967; Thorleifson, 1996; Anderson *et al.*, 2002) and glacial lakes Columbia and Missoula in Washington (Atwater, 1986). The age dichotomy between the New England varve chronology and the Champlain Sea shell chronology (see below) is essentially resolved by application of the marine reservoir correction mentioned above and by rejection of dates on deposit feeding molluscs.

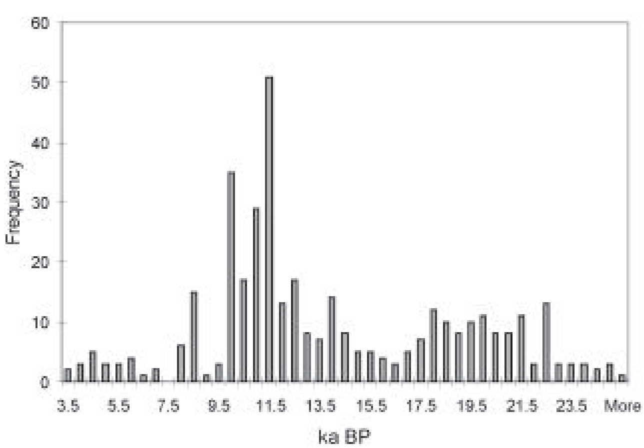


Fig. 3. Histogram of maximum-limiting radiocarbon dates on the advance to the Late Wisconsinan glacial limit and on readvances during general deglaciation in North America.

The remaining age control is entirely in the form of minimum-limiting dates on lake sediments, peat, wood, plant macrofossils, foraminifera and mammal bones, listed in decreasing order of abundance. Reservoir corrections are applied to dates on foraminifera as they are to marine molluscs. Appropriate reservoir corrections for dates on marine mammal bones, most of which herein are from wide ranging whales, are difficult to assess. All are currently corrected by –400 years, pending further assessment, but this seems adequate in light of bone-wood comparisons in the Canadian Arctic (Dyke *et al.*, 1996). The chronological database contains much redundancy, perhaps as much as 25%, because many sites have minimum-limiting ages that are younger than others nearby.

Well-dated readvances, discussed further below, form modes in Fig. 3. These events are under-represented, because only dates on overridden or glacially-transported materials are included and many readvance moraines are only relatively-dated. The prominent well-dated events are as follows: (1) readvance of the Des Moines Lobe to the Bemis Moraine in Iowa about 14 ka BP (Ruhe, 1969; Clayton & Moran, 1982); (2) the resurgence that deposited Two Rivers Till over the Two Creeks Forest Bed in Wisconsin and above the Cheboygan bryophyte bed in Michigan about 11.8-11.5 ka BP (Black, 1976; McCartney & Mickelson, 1982; Schneider, 1990; Kaiser, 1994; Larsen *et al.*, 1994); (3) the Sumas or Sumas I readvance in the Fraser Lowland of British Columbia and Washington about 11.5-11 ka BP (Clague *et al.*, 1997; Kovanen & Easterbrook, 2002) and probable nearby correlative (Mamquam Valley, BC, Friele & Clague, 2002a); (4) a readvance of Laurentide ice into the Champlain Sea near Québec City 11.5-11.0 ka BP (LaSalle & Shilts, 1993); (5) readvances of valley glaciers from the Gaspé Ice Cap in the Québec Appalachians about 11.5-10.8 ka BP (Hétu & Gray, 2000); (6) readvance of the north-westernmost Laurentide Ice Sheet to the Winter Harbour till limit in Viscount Melville Sound 11-10.5 ka BP (Hodgson & Vincent, 1984); (7) the Gold Cove readvance of Labrador ice across the mouth of Frobisher Bay on south-eastern Baffin Island at 10 ka BP (Miller & Kaufman, 1990); (8) the Marquette readvance

Fig. 4 (following pages). Palaeogeographic maps of glaciated North America at 500-year and occasionally finer time steps. Times are indicated on each map in both radiocarbon and calendar years.


18 C ka BP


17.5 C ka BP


17 C ka BP

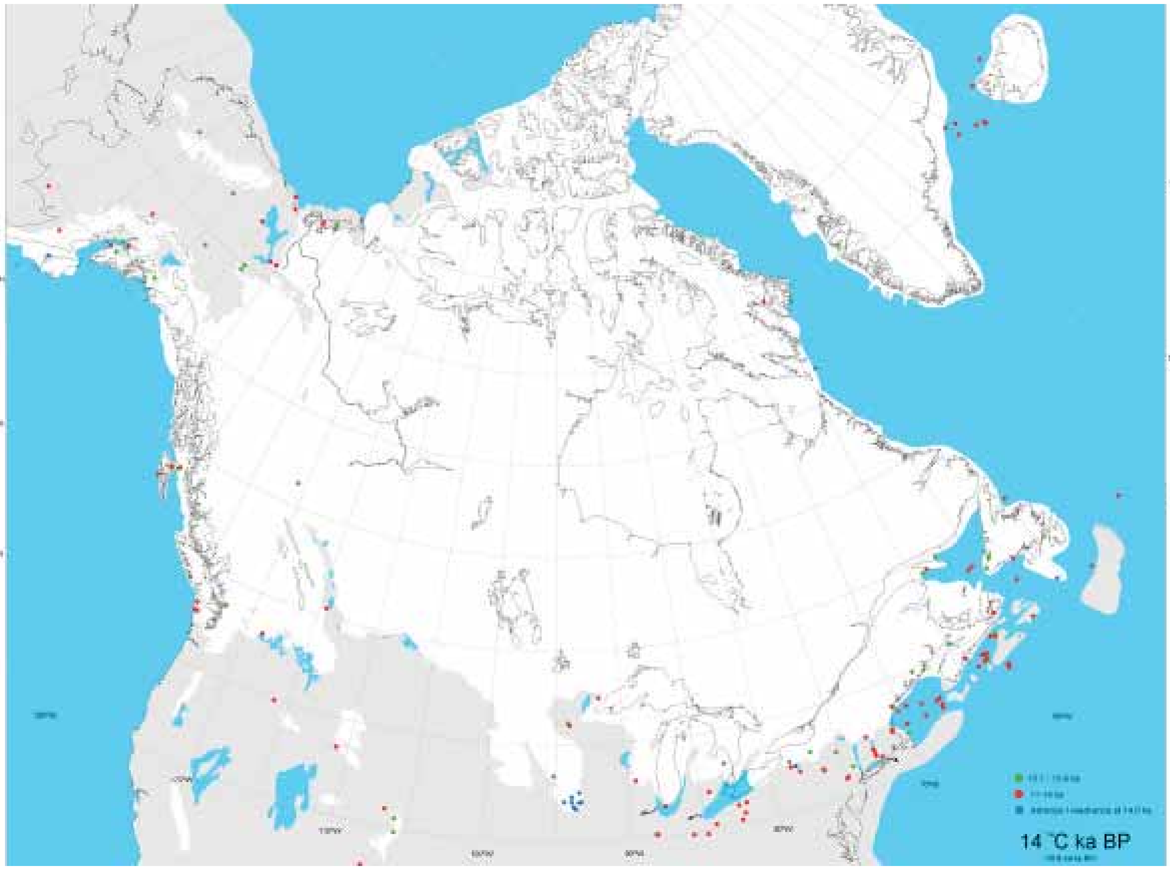

16.5 C ka BP

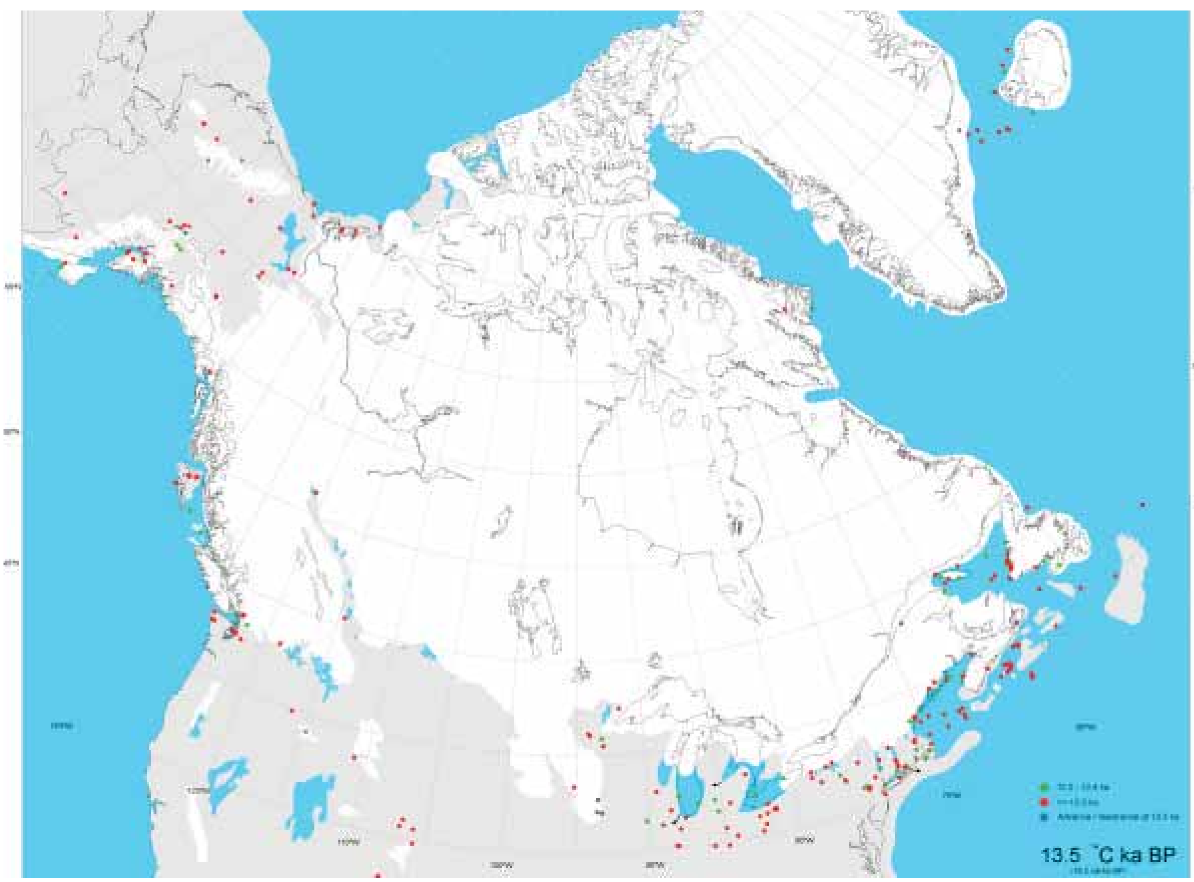

16 ¹⁴C ka BP

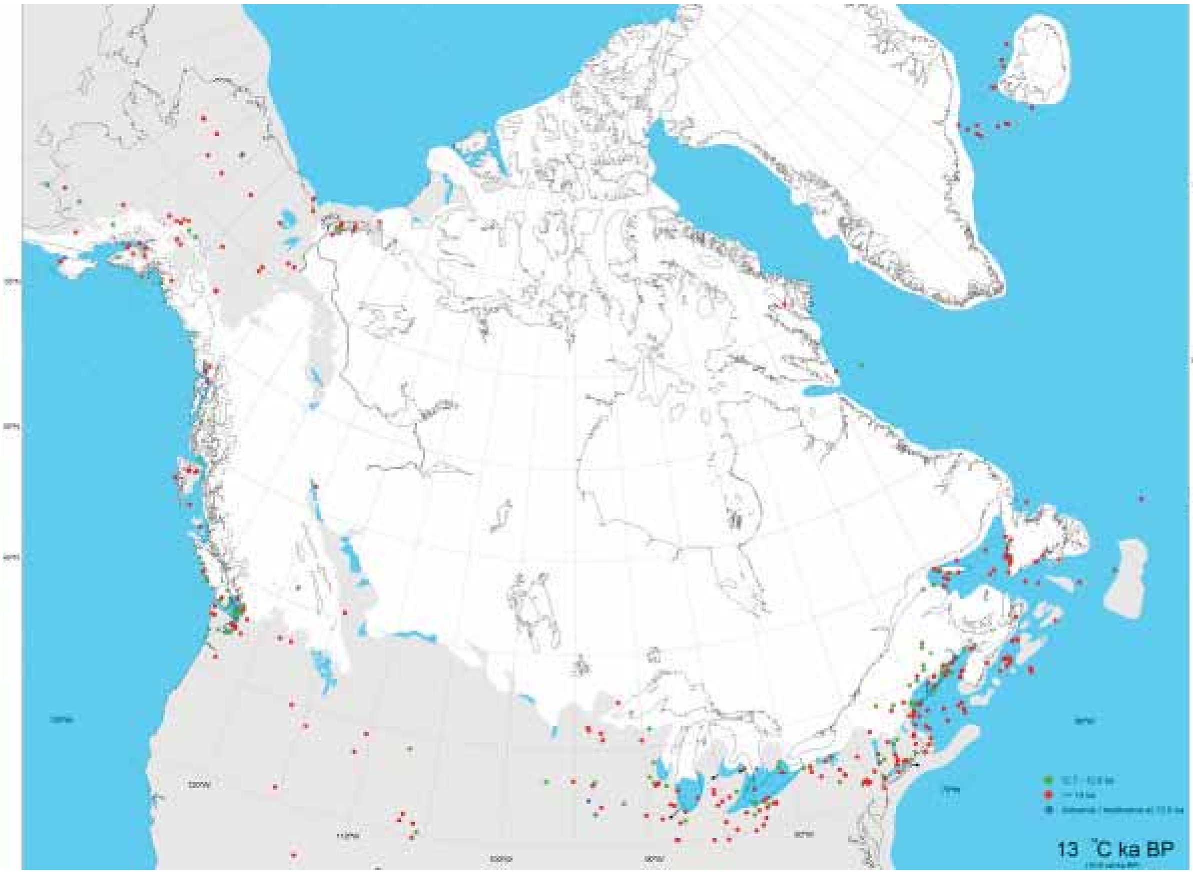

15.5 ¹⁴C ka BP

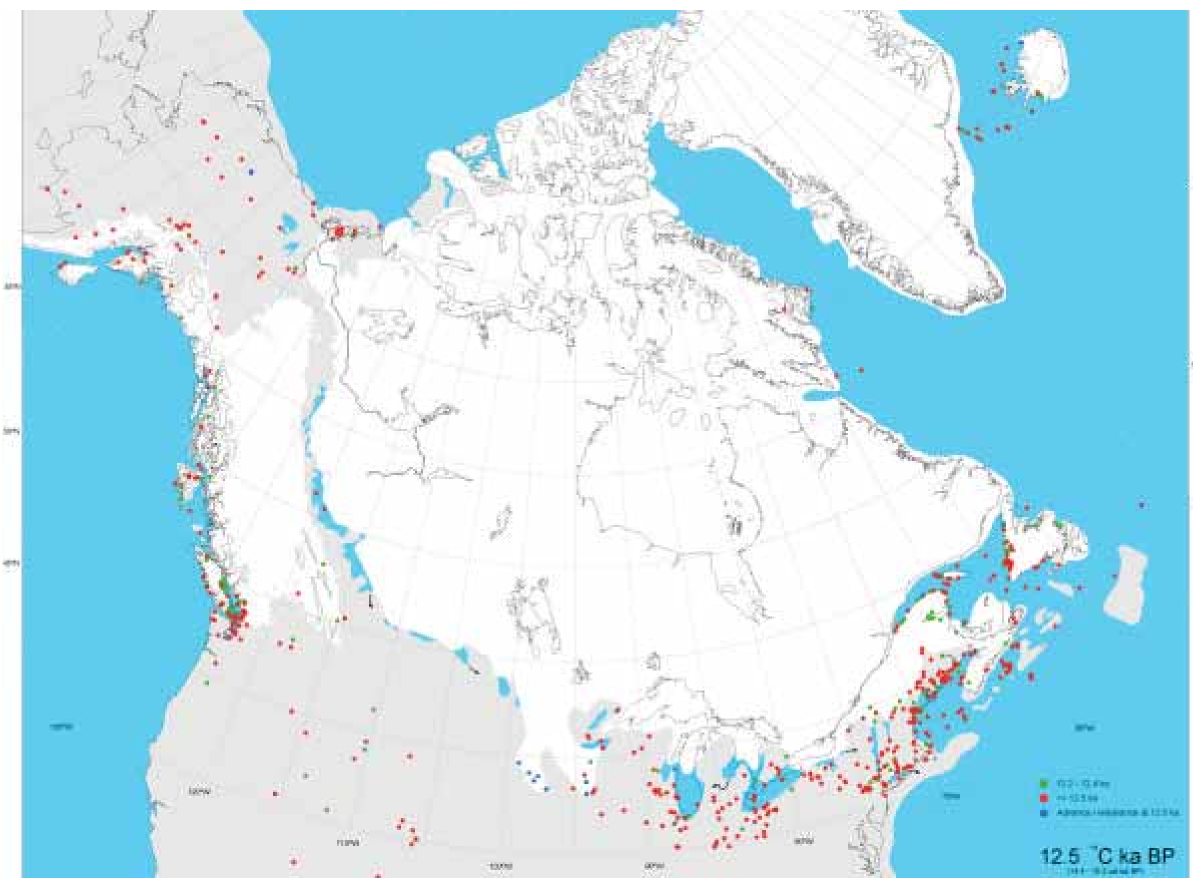

15 ¹⁴C ka BP

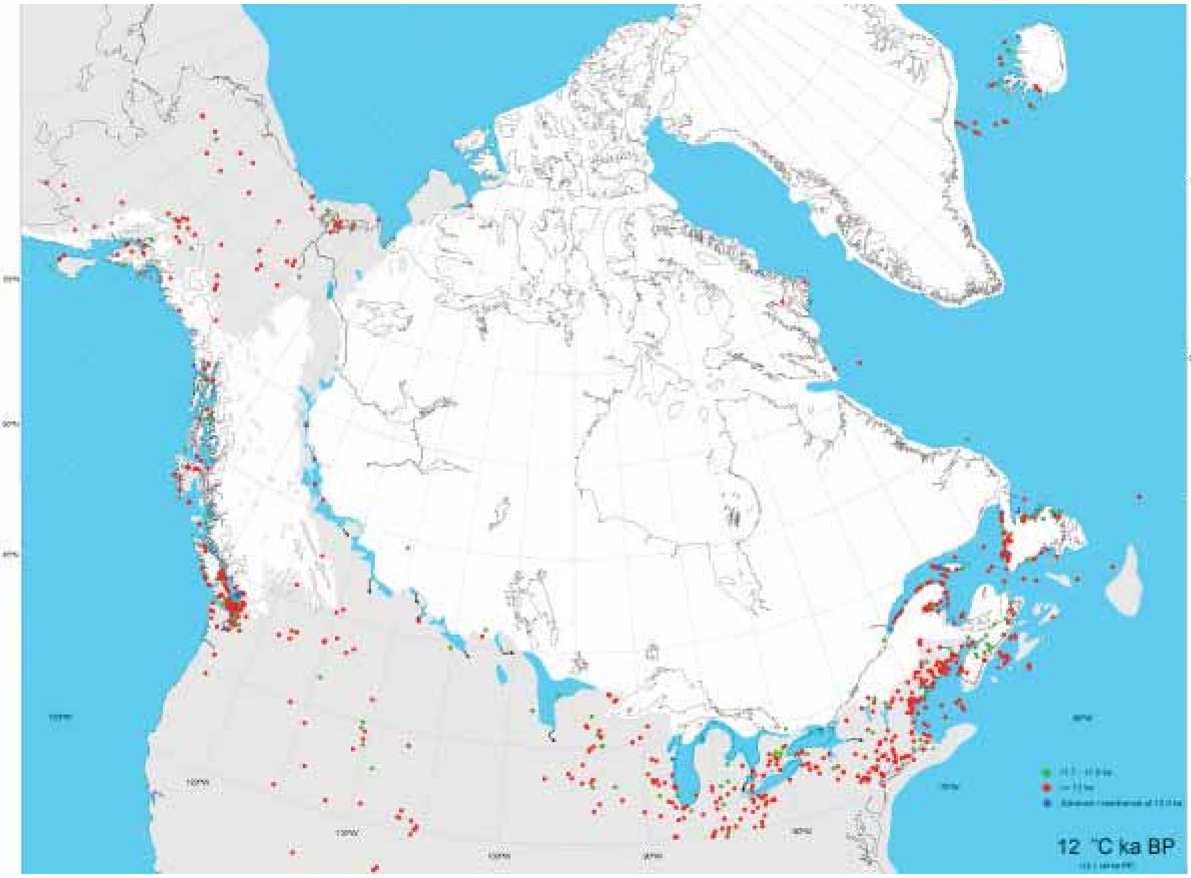

14.5 ¹⁴C ka BP

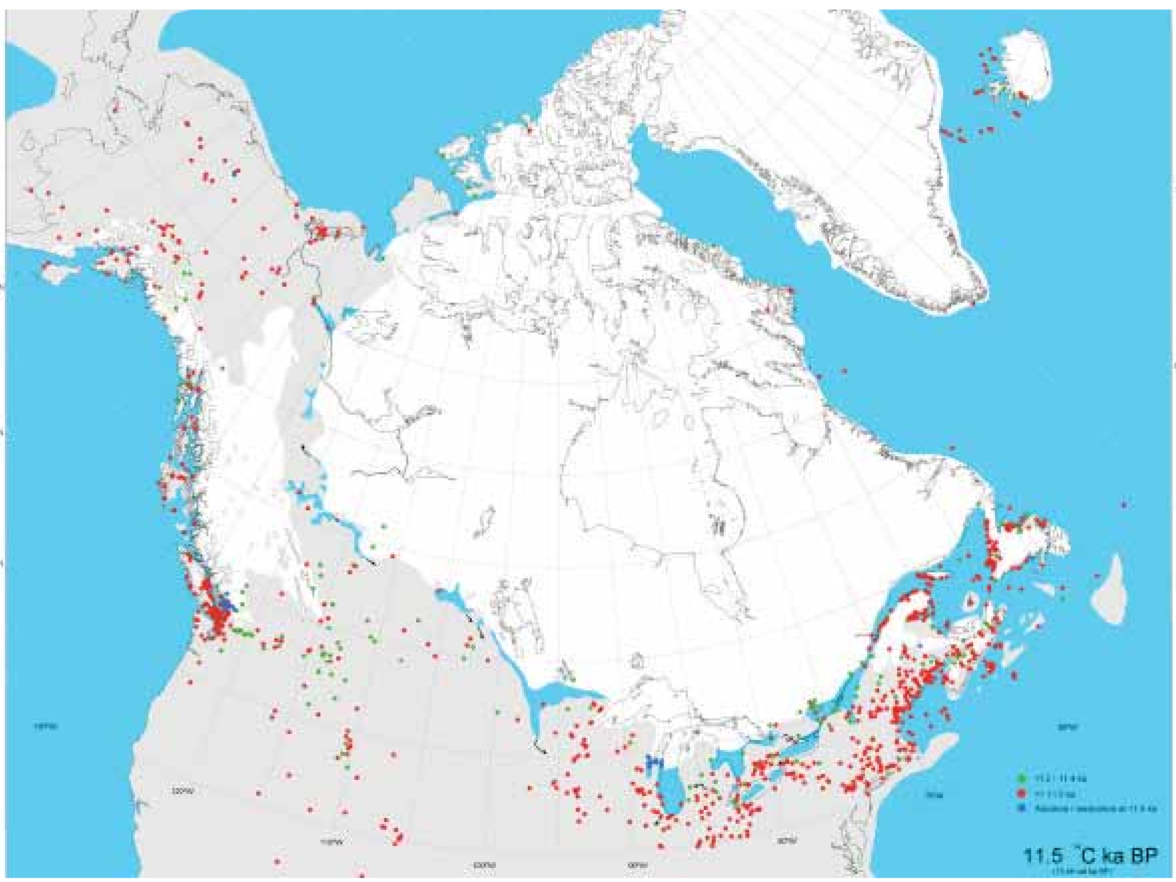

14 C ka BP

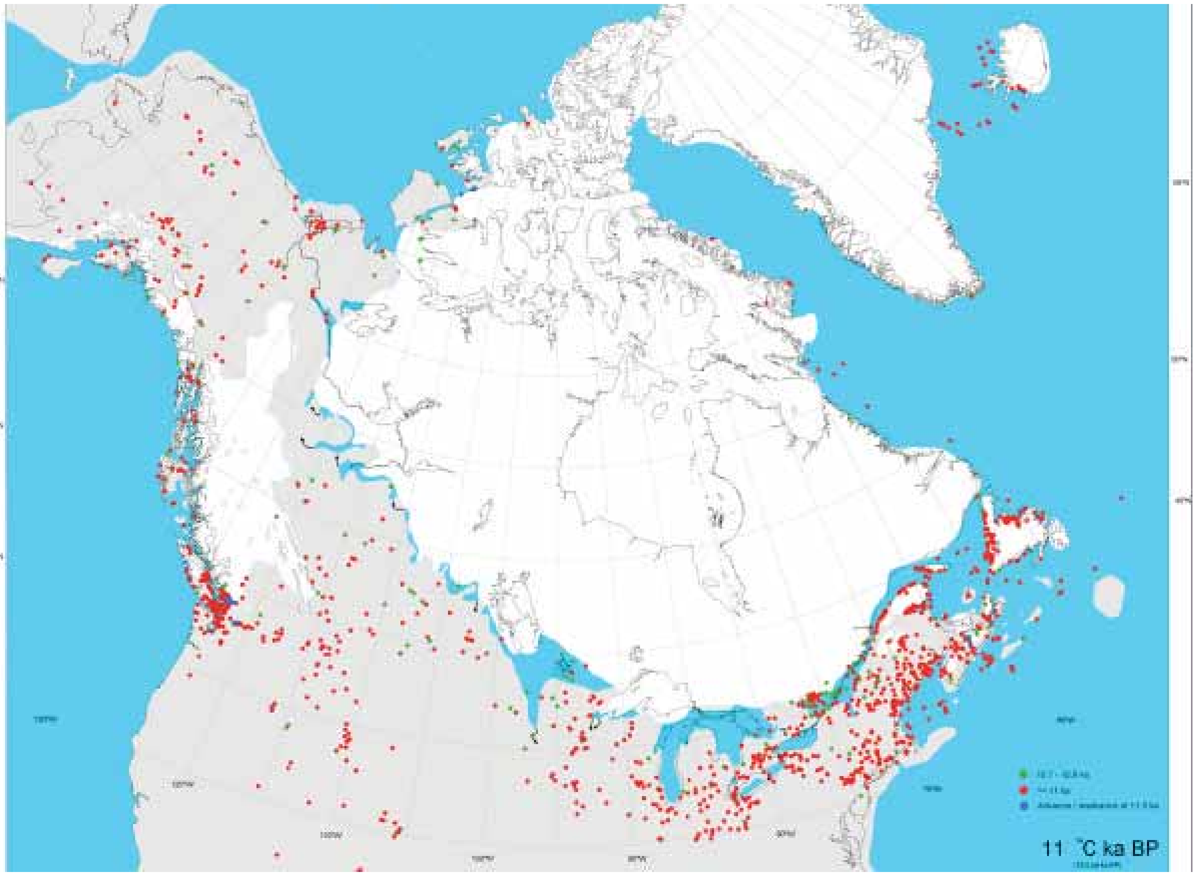

13.5 C ka BP

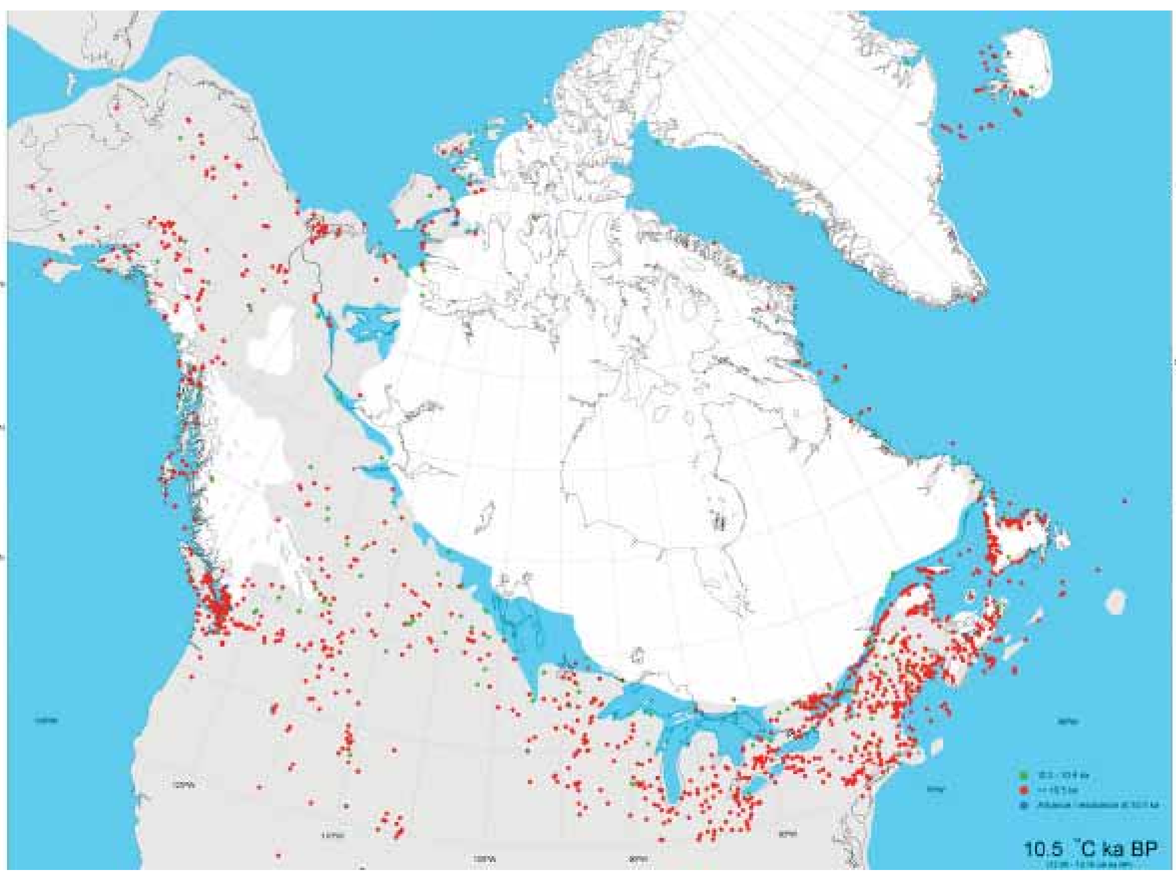

13 ^{14}C ka BP

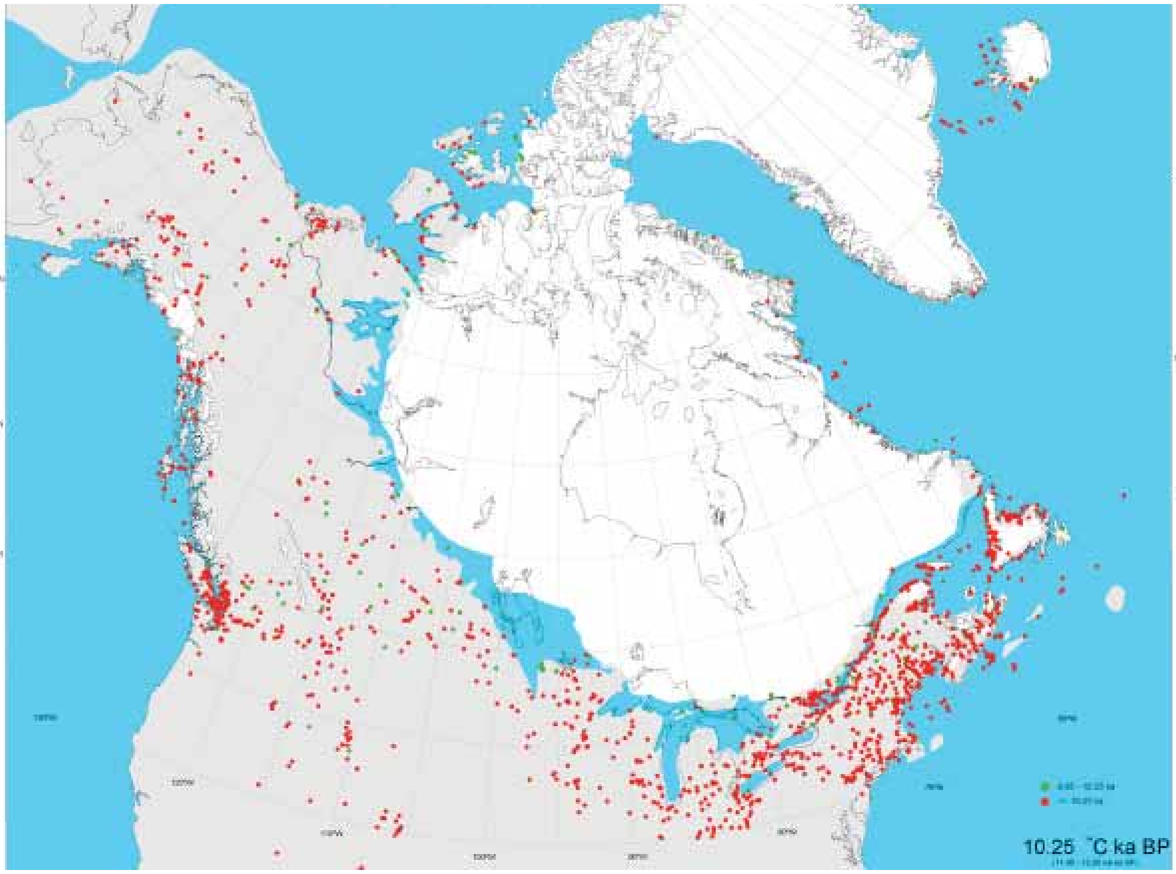

12.5 ^{14}C ka BP

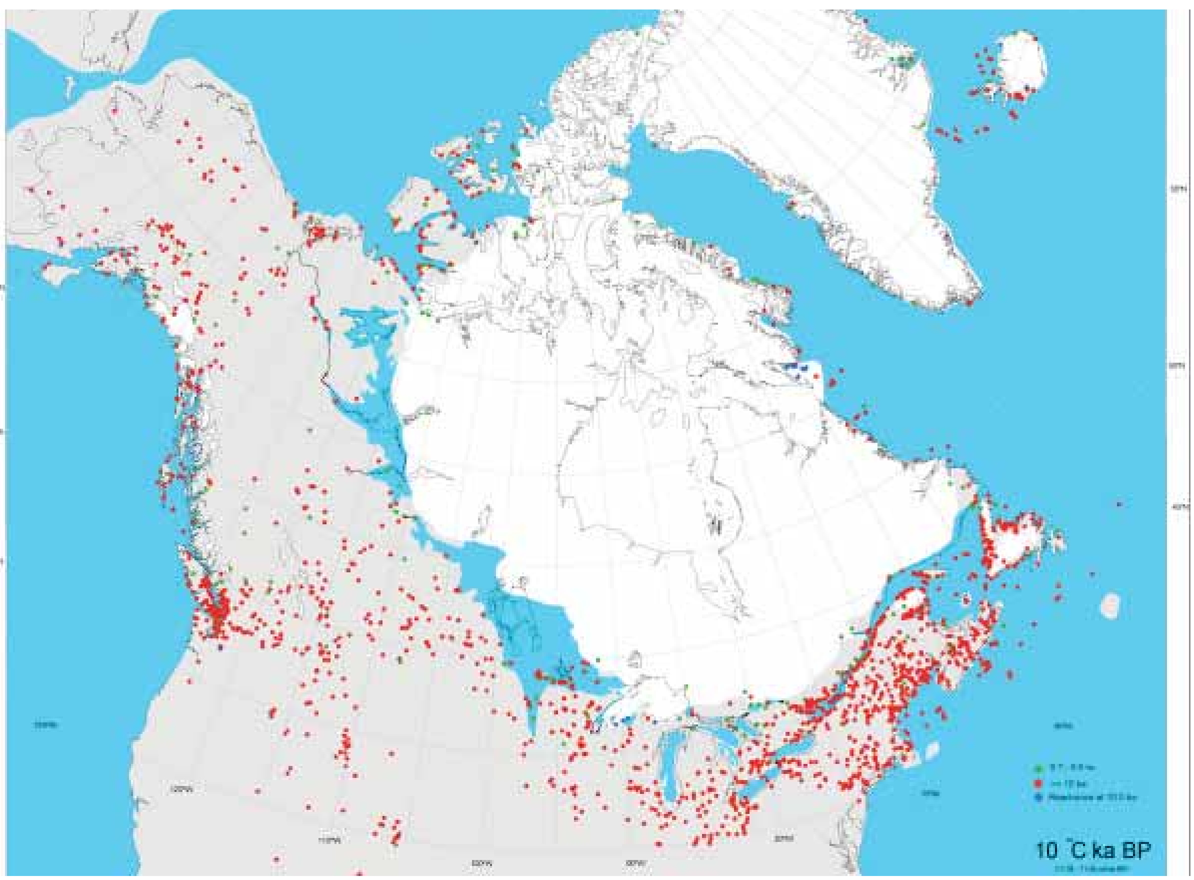

12 ¹⁴C ka BP

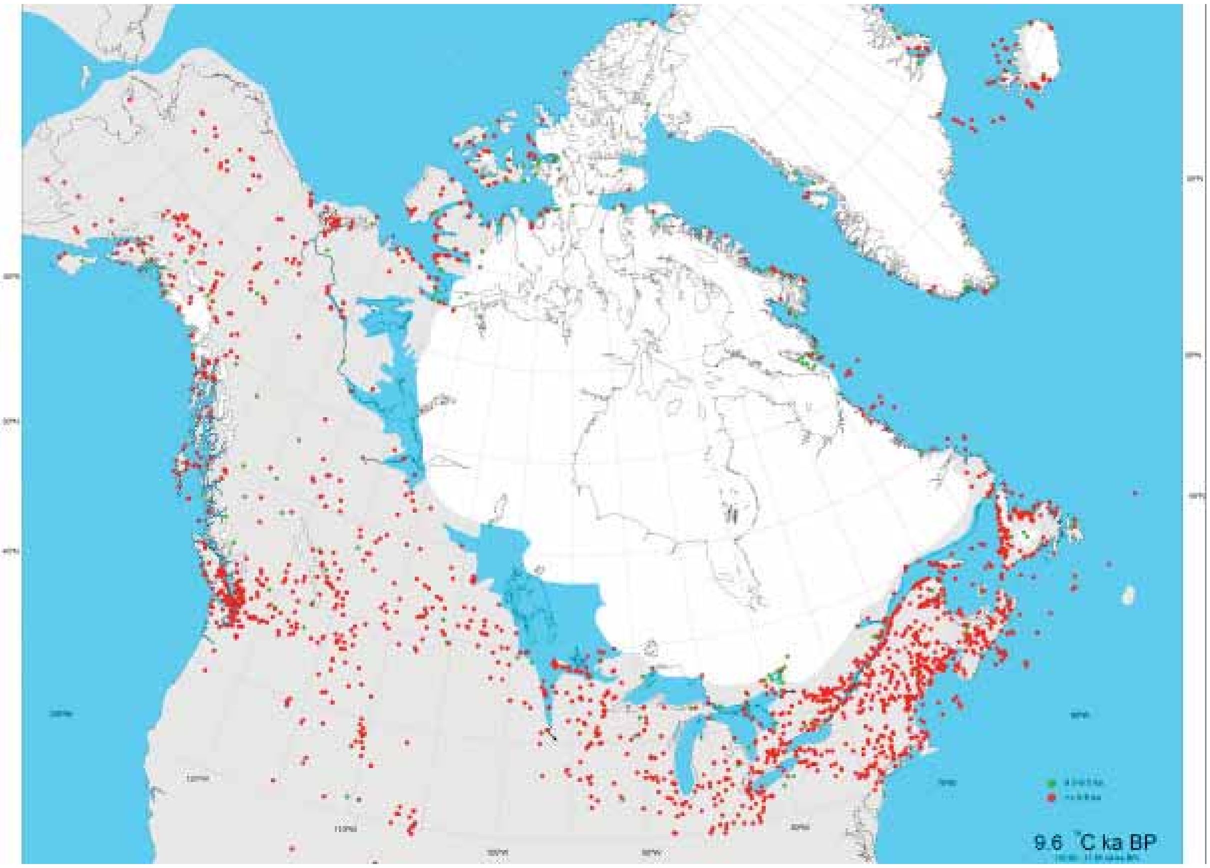

11.5 ¹⁴C ka BP

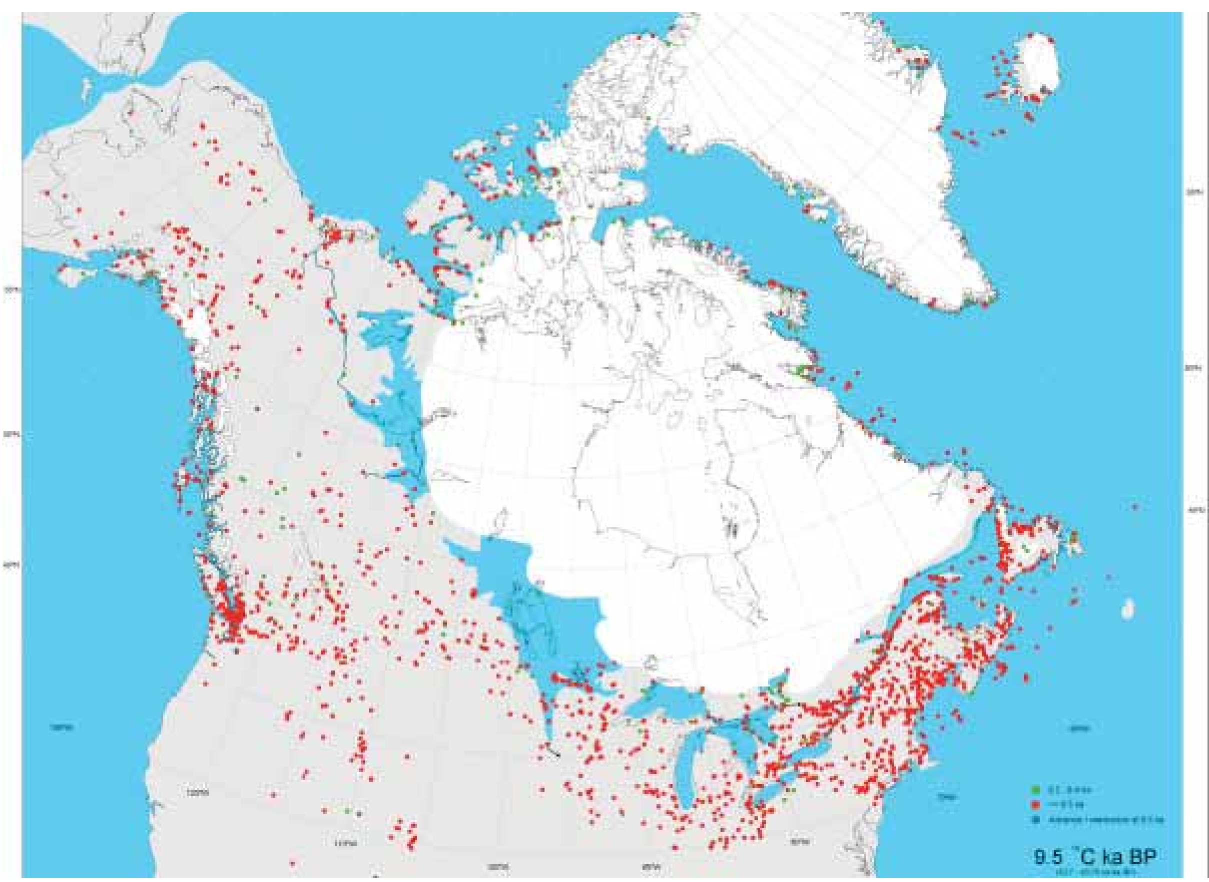

11 ¹⁴C ka BP

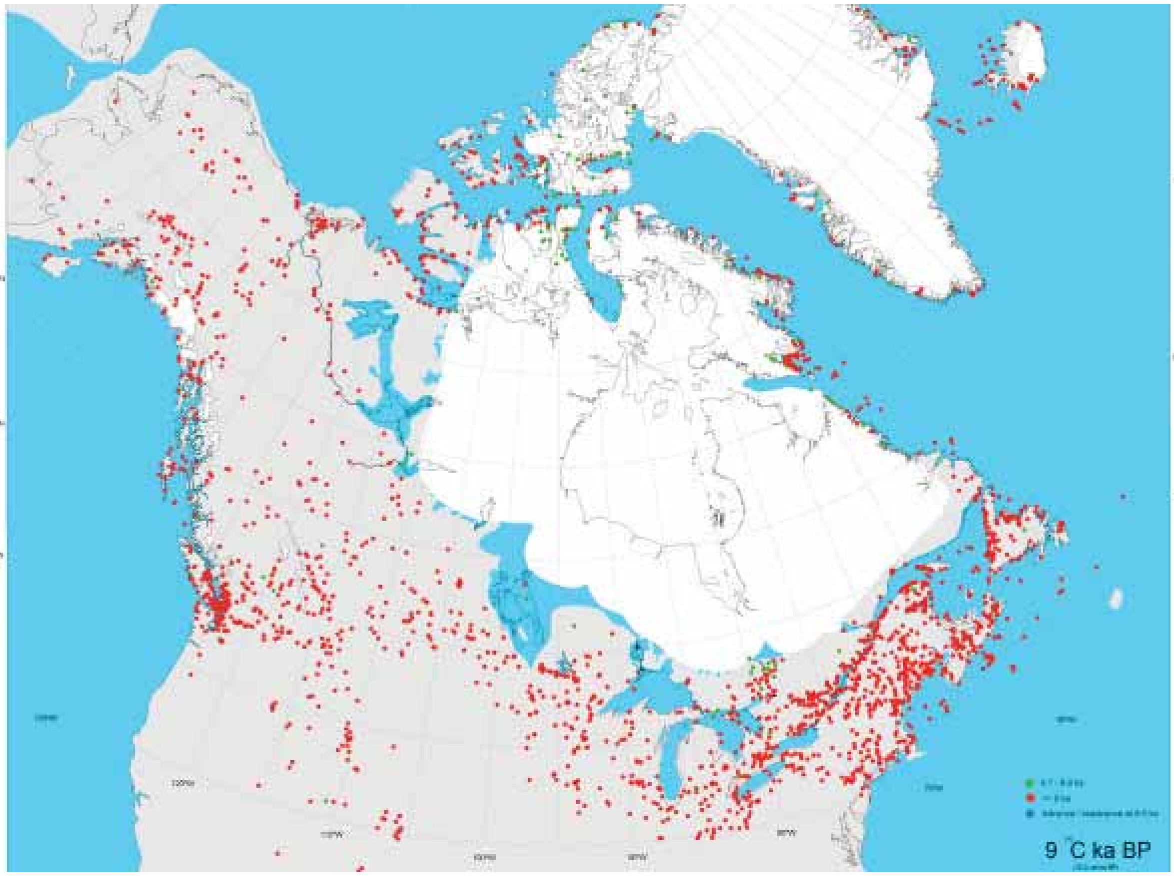

10.5 ¹⁴C ka BP

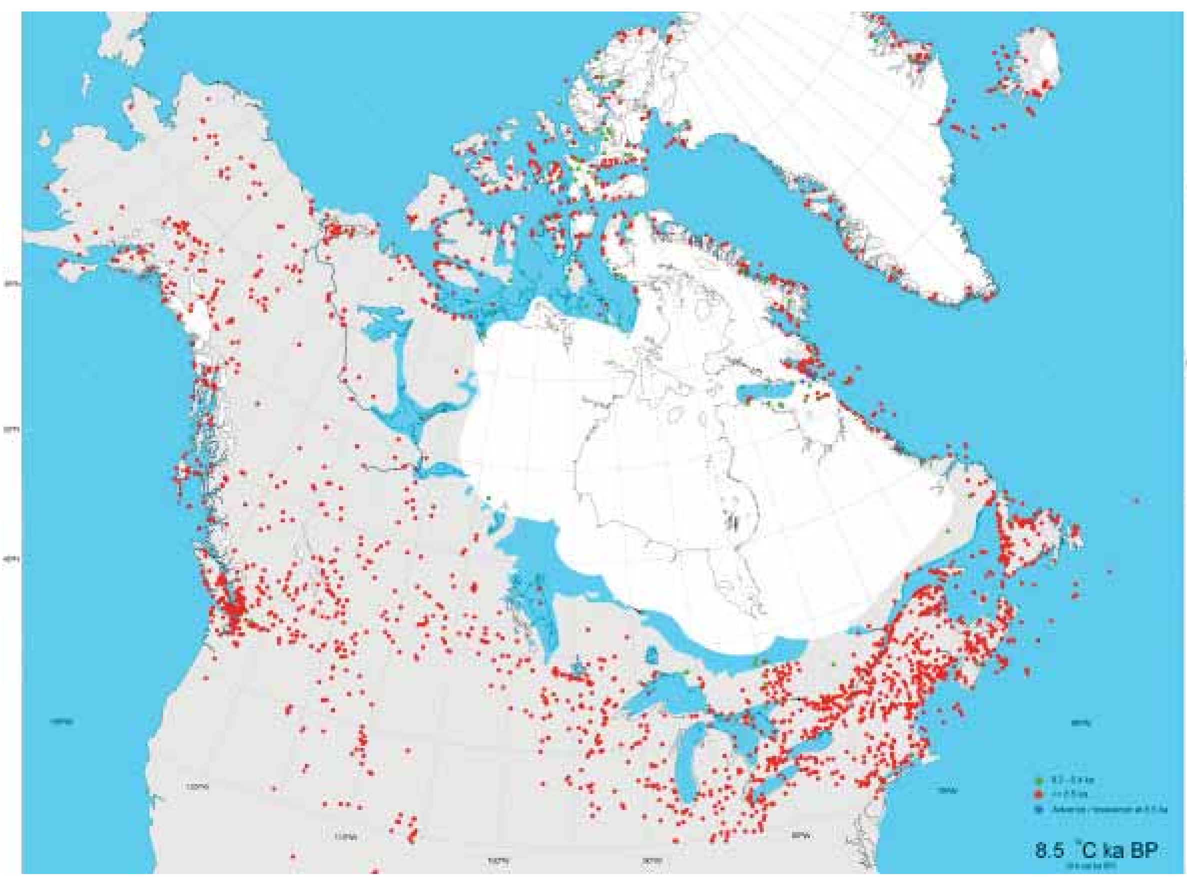

10.25 ^{14}C ka BP

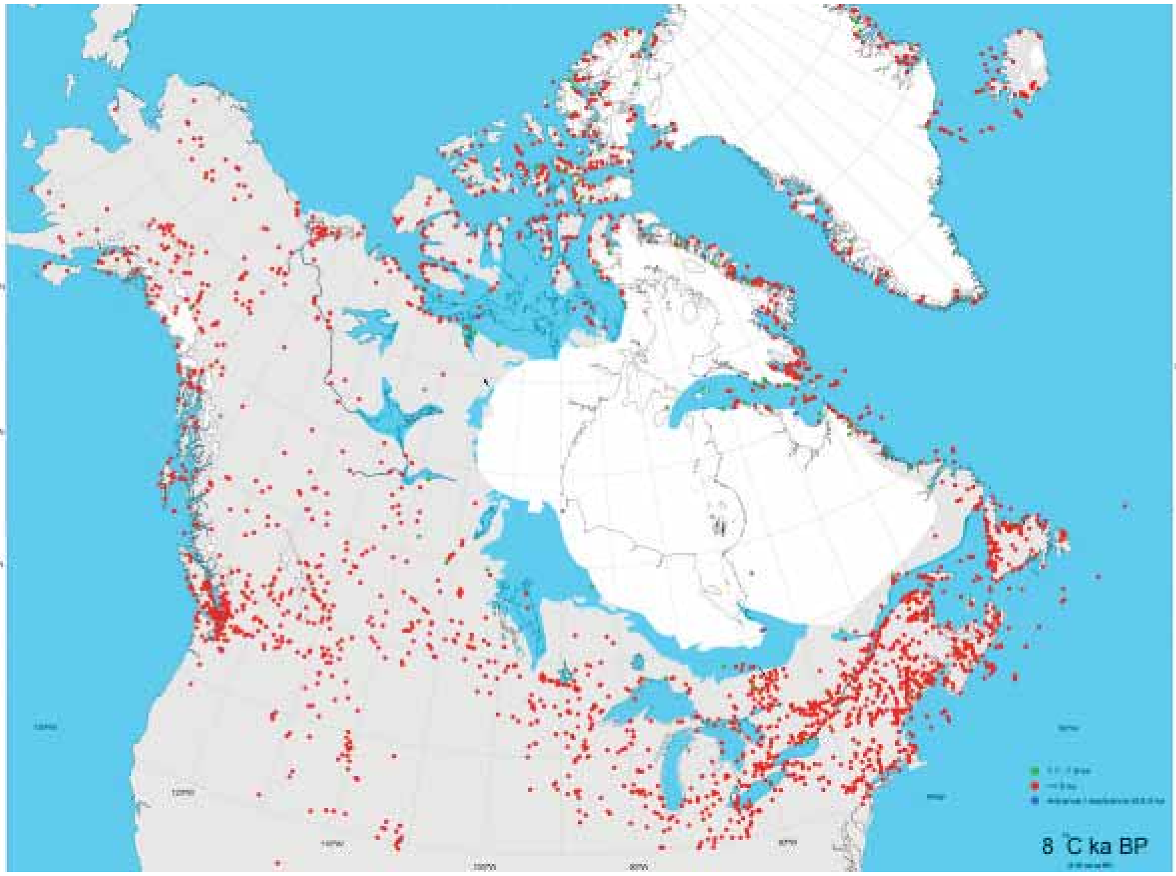

10 ^{14}C ka BP

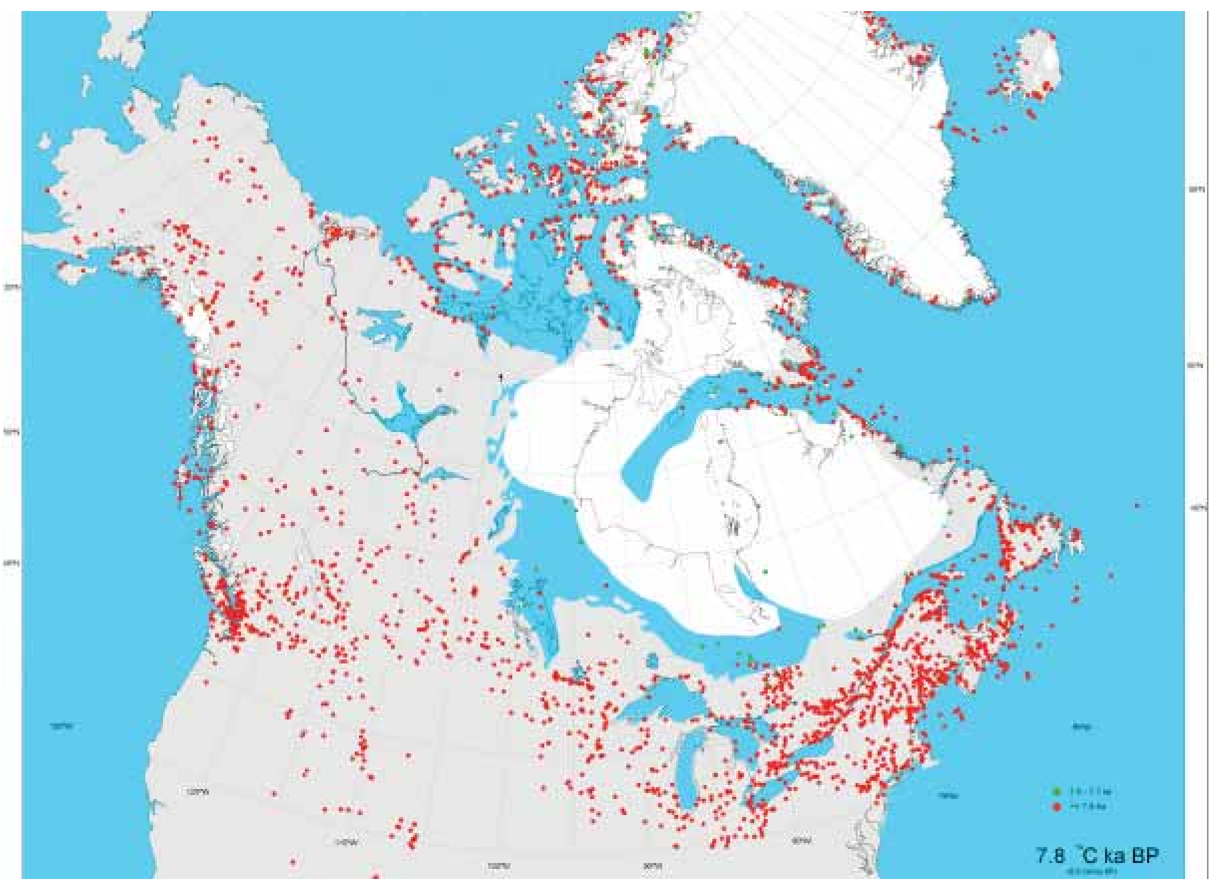

9.6 ^{14}C ka BP

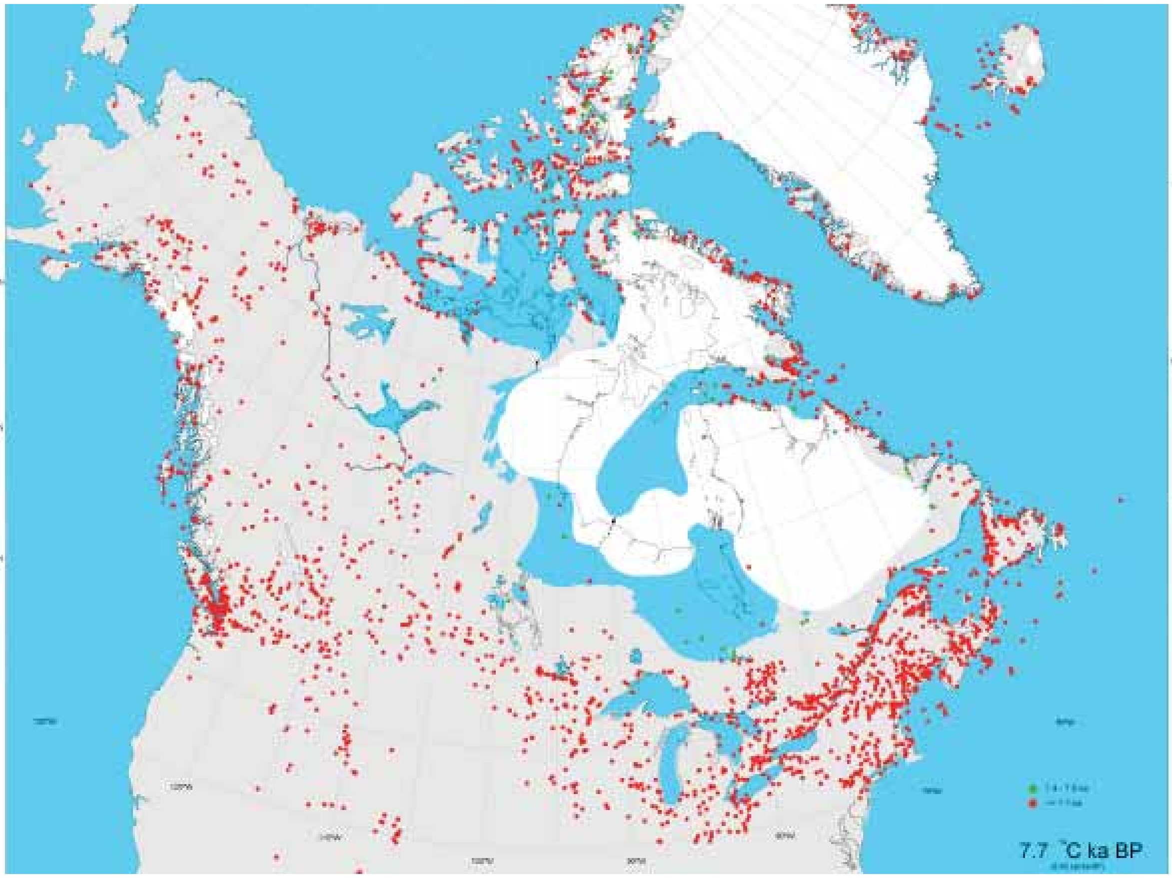

9.5 ^{14}C ka BP

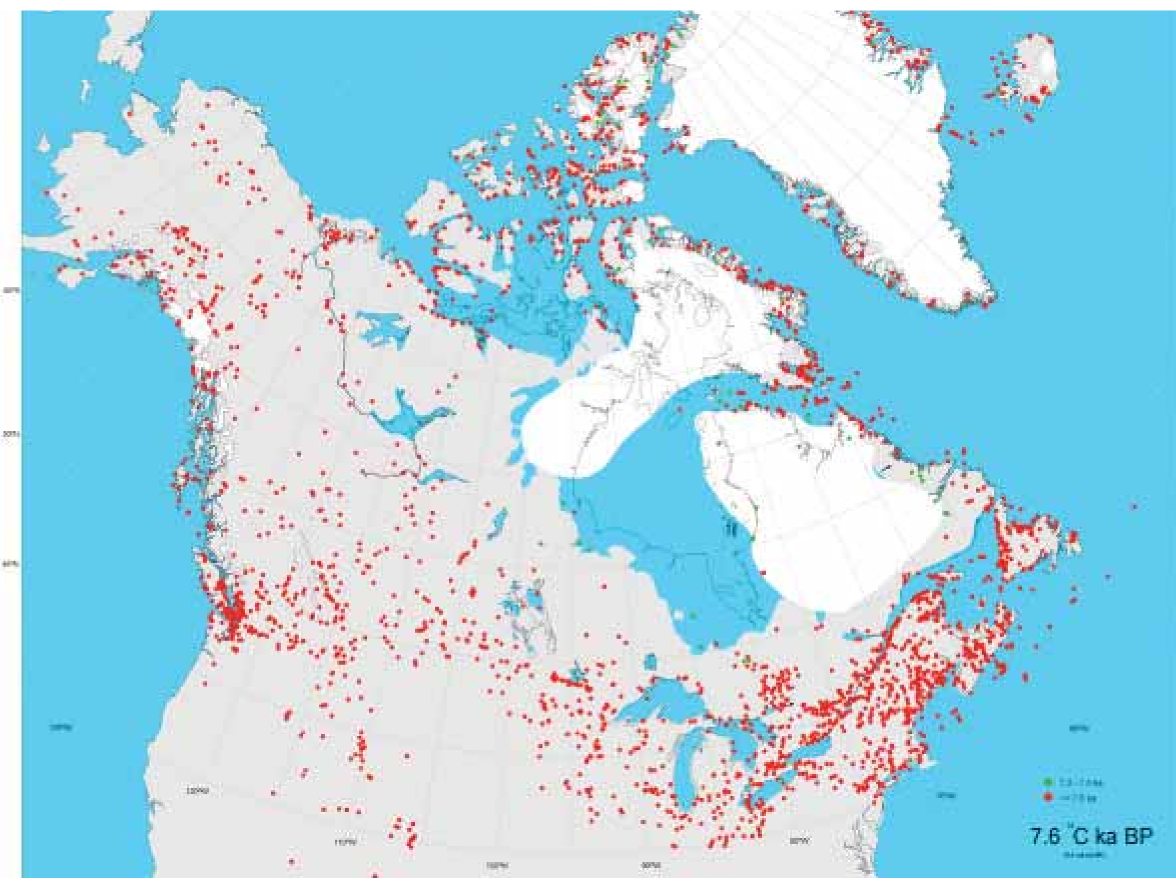

9 C ka BP

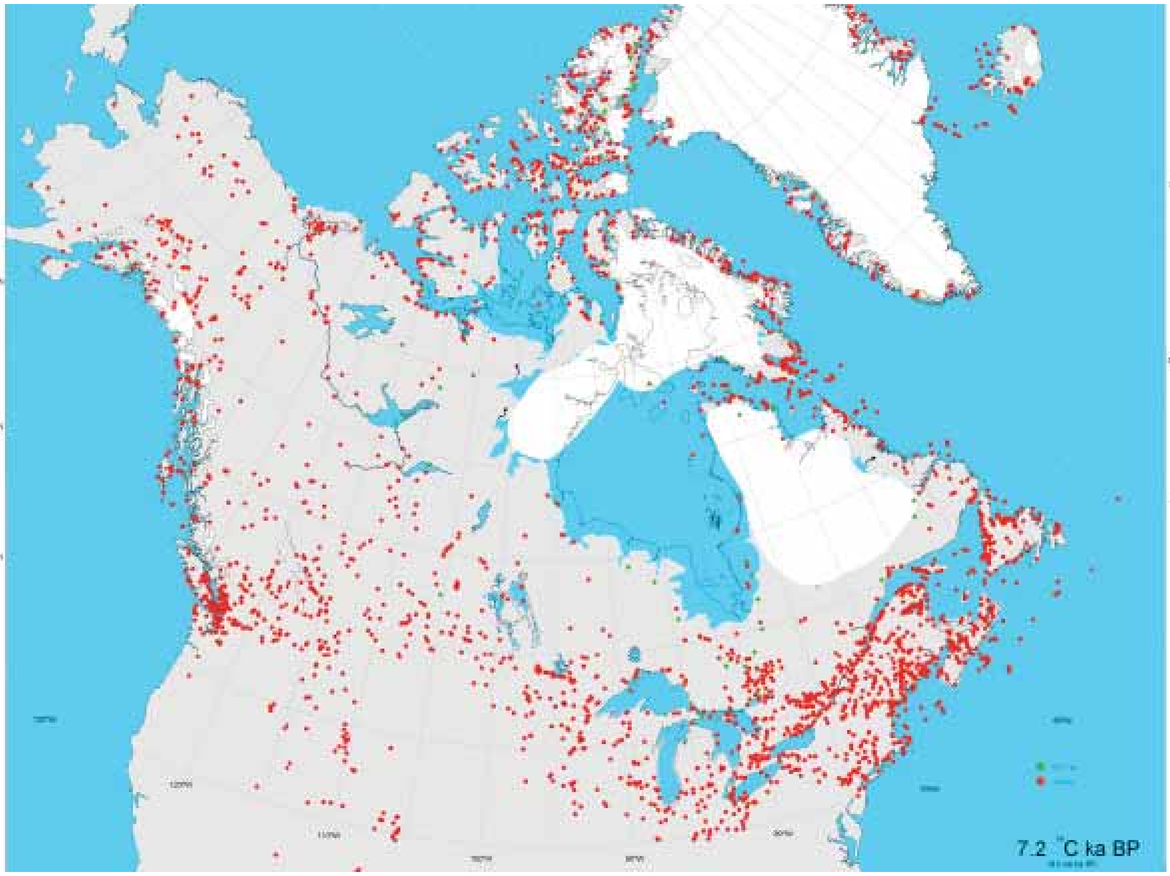

8.5 C ka BP

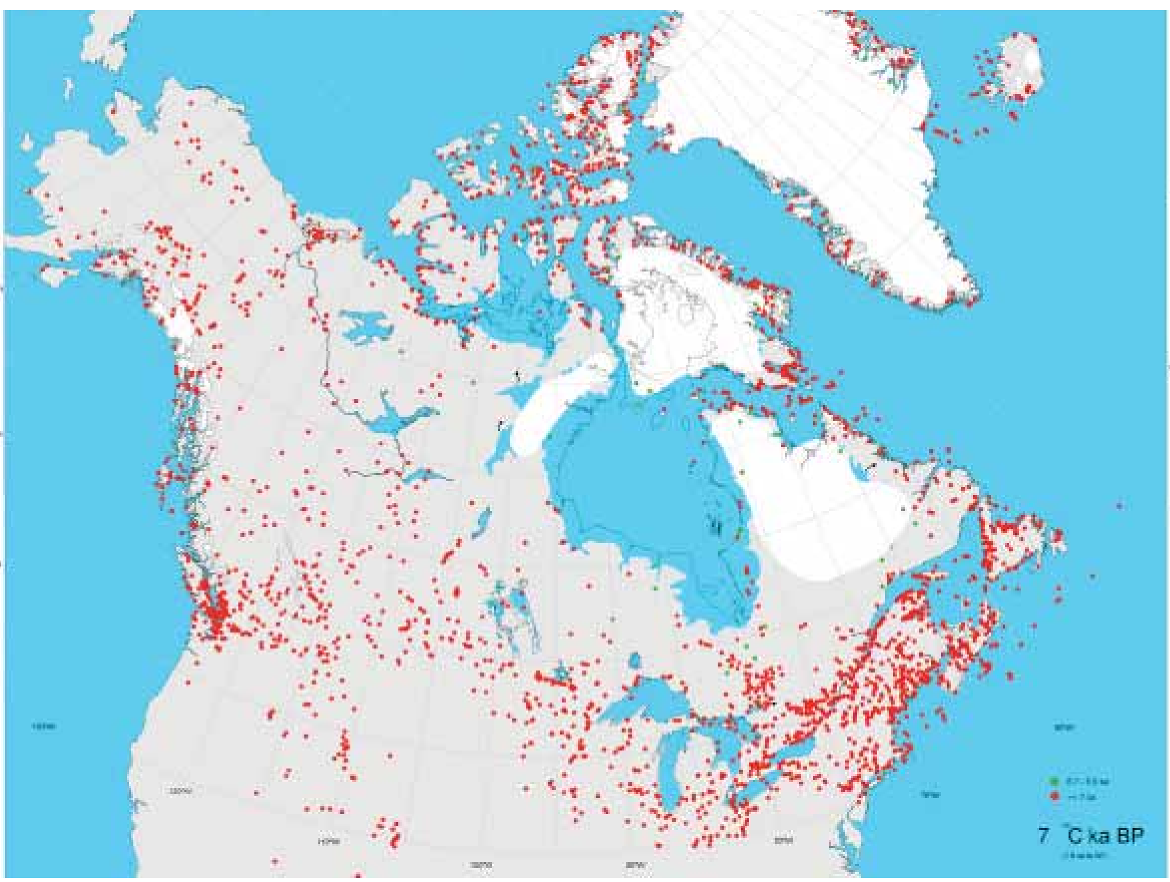

8 ¹⁴C ka BP

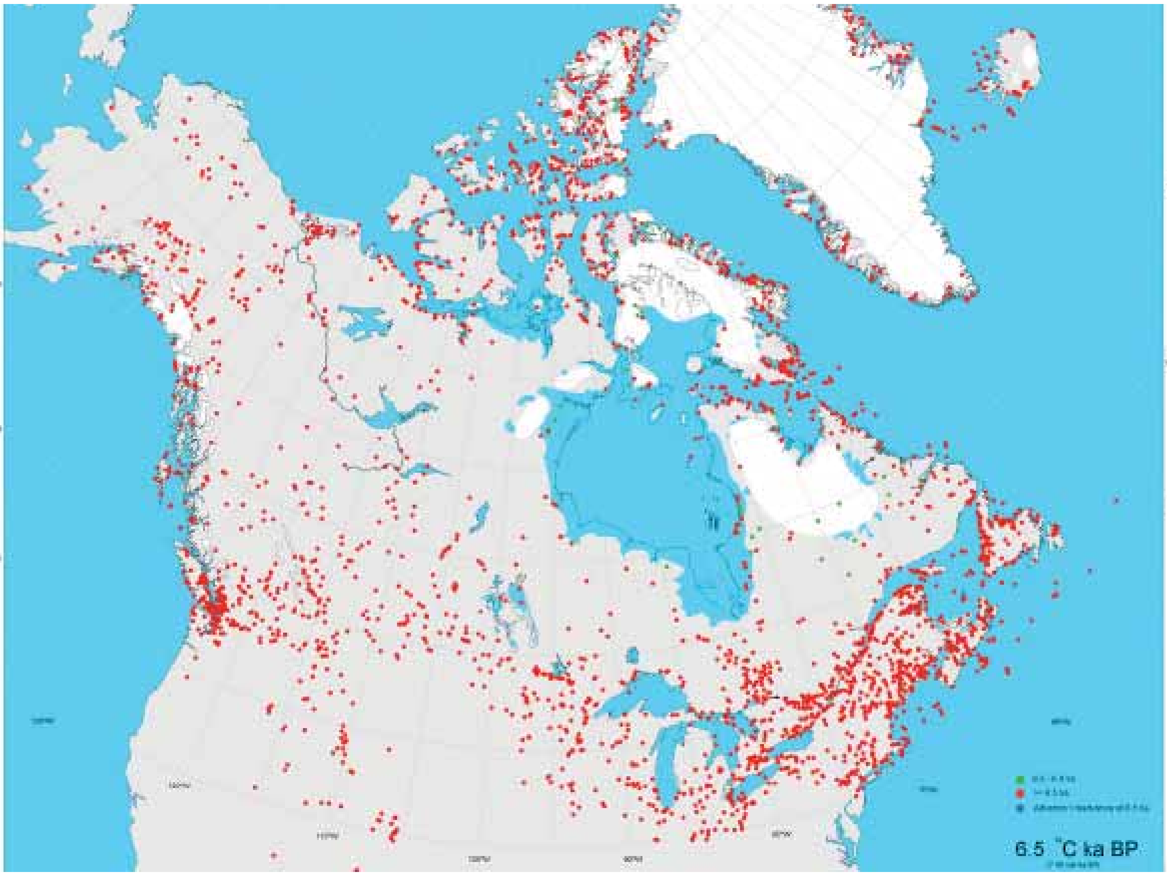

7.8 ¹⁴C ka BP

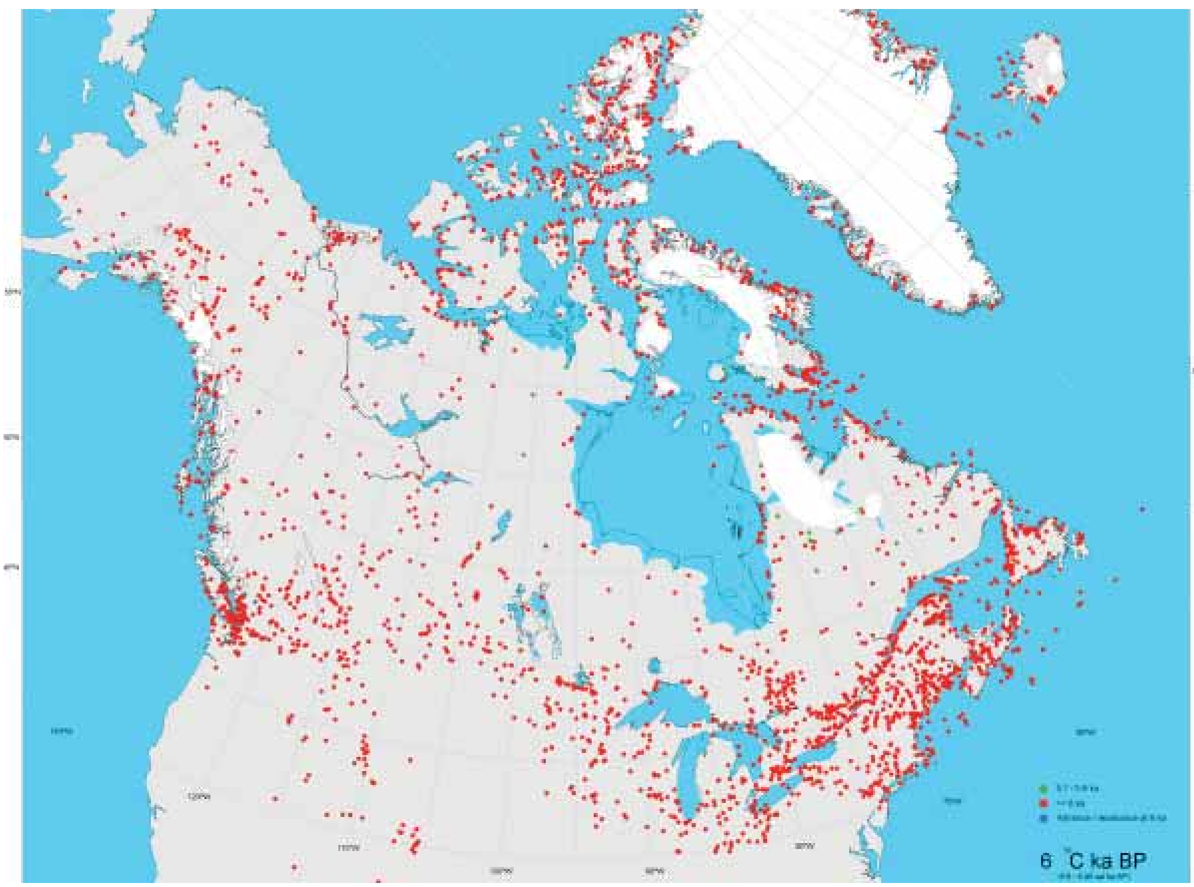

7.7 ^{14}C ka BP

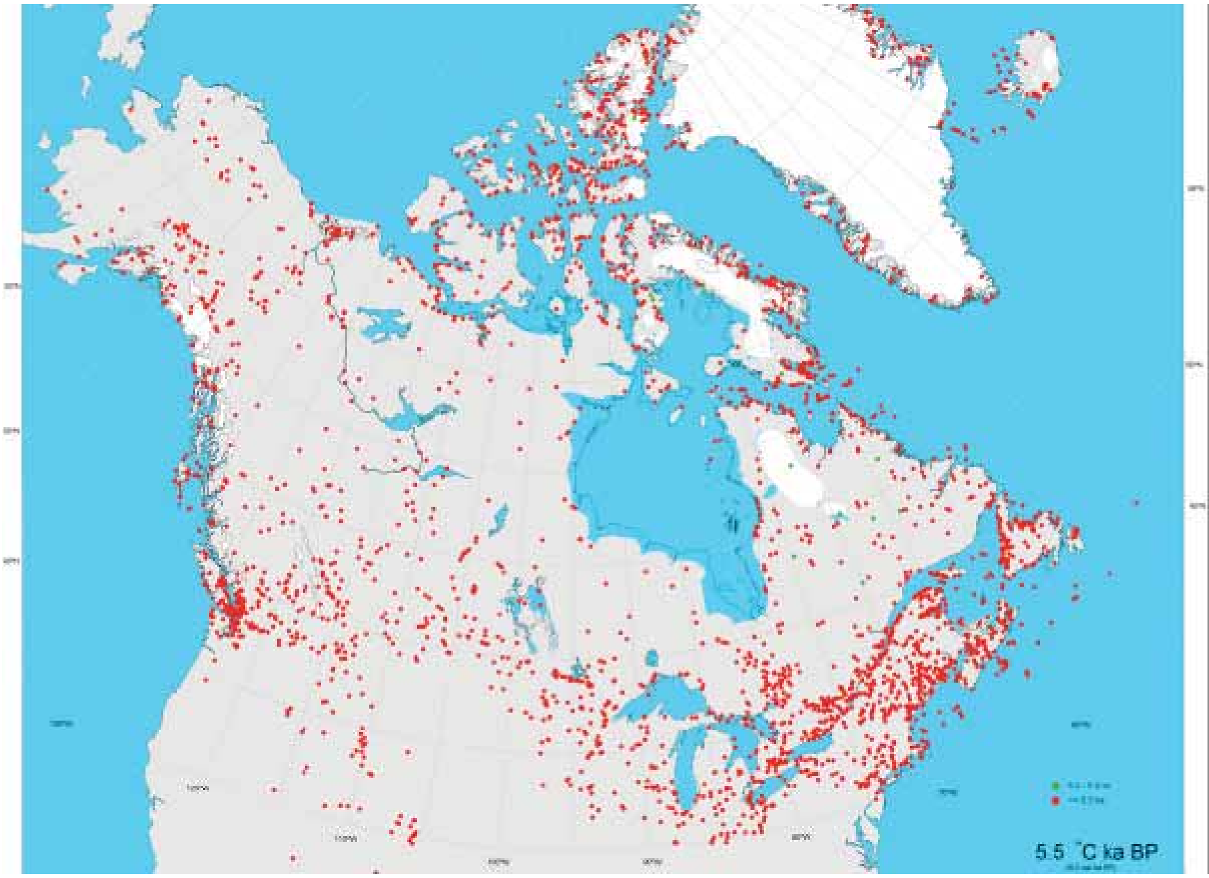

7.6 ^{14}C ka BP

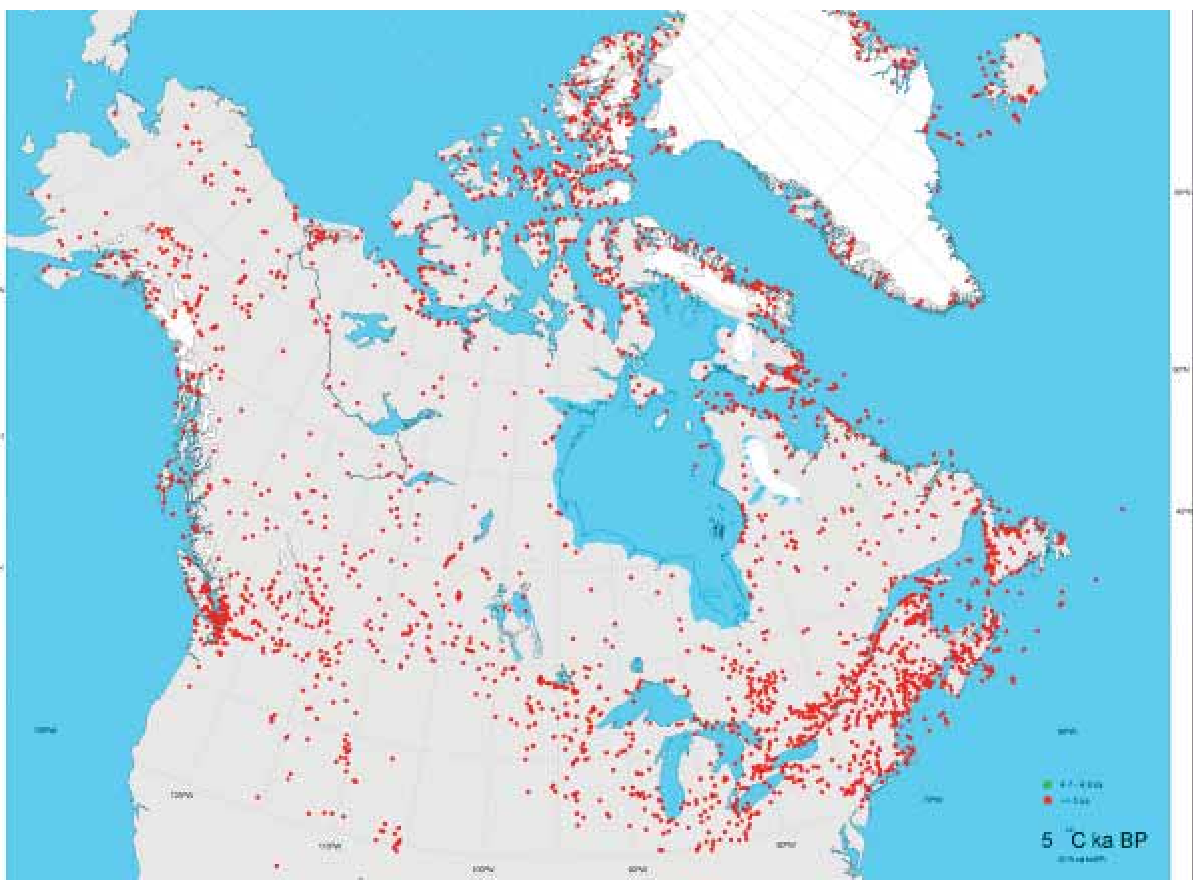

7.2 ¹⁴C ka BP


7 ¹⁴C ka BP


6.5 ^{14}C ka BP


6 ^{14}C ka BP


5.5 ^{14}C ka BP


5 ^{14}C ka BP

across Lake Superior at 10 ka BP (Clayton, 1983; Drexler *et al.*, 1983; Lowell *et al.*, 1999); and (9) the Nobel Inlet readvance of Labrador ice onto the south shore of Baffin Island 8.5-8.4 ka BP (Stravers *et al.*, 1992; Manley & Miller, 2001).

In summary, ages of deglaciation are now reasonably well established in most coastal areas, the west coast of Hudson Bay and eastern Foxe Basin being exceptions. Varve chronologies and dated readvances provide control at key inland sites. Elsewhere, minimum-limiting dates are valuable constraints. However, there is a continuing need to cull the radiocarbon database as better dating proceeds. Furthermore, large regions remain nearly devoid of dates, prominently interior British Columbia and the Canadian Shield region stretching from central Ontario to the Mackenzie Valley and the Arctic coast.

Broad patterns of ice cover change through time

LGM to Erie Interstadial (18-15.5 or 14.5 ka BP)

The Summit ice core record in Greenland indicates progressive warming from LGM to 14.2 ka BP (17 cal ka BP; Johnsen *et al.*, 1992). The LGM is the interval encompassing the last eustatic sea-level minimum centreed on 18 ka BP (21.4 cal ka BP; Mix *et al.*, 2001). Because the North American ice sheets account for more than half, possibly 75%, of Late Pleistocene ice in excess of present volumes, these ice sheets must also have held maximum or near maximum ice volumes at that time. The southern limits of Laurentide ice were attained generally about 22 ka BP (Dyke *et al.*, 2002a), well in advance of the major growth of the southern part of the Cordilleran Ice Sheet (Clague, 1989). By 18 ka BP, ice cover in North America was at or near its maximum everywhere except along the southern Cordilleran ice margin (Fig. 4-18 ka).

Dated wood from below drift at several sites shows that the Fraser Lowland and Strait of Georgia region of British Columbia were not glaciated until after 18 ka BP (Hicock & Lian, 1995; Lian *et al.*, 2001), nor indeed until after 16 ka BP (Clague *et al.*, 1988), when the ice margin was still some 300 km short of its Late Wisconsinan limit in the Puget Lowland of Washington. South-western Cordilleran ice continued to build to its maximum extent until about 14 ka BP. However, the south-eastern Cordilleran ice margin had advanced far enough to retain lakes Missoula and Columbia prior to 15.5 ka BP, by which time the first of the Missoula floods had occurred (Atwater, 1986).

The west-central to north-western margins of the Cordilleran Ice Sheet, however, appear to have responded more similarly to Laurentide margins during this interval. Ice was in recession by 16 ka BP on the Queen Charlotte Islands (Clague *et al.*, 1982; Warner *et al.*, 1982; Blaise *et al.*, 1990), and much of the huge Cook Inlet ice stream in Alaska had disappeared by 15.5 ka BP (Reger *et al.*, 1996). The preceding maximum ice cover, therefore, occurred during the global LGM (Mann & Peteet, 1994).

During the same 3-ka interval, parts of the Laurentide Ice Sheet underwent net recession. The Mackenzie Lobe had pulled back from the terminal moraine on Herschel Island by 16.2 ka BP, allowing horses to live there (Harington, 1989), and the Hudson Strait ice stream pulled back to a poorly-defined position within the strait after discharging a pulse of icebergs during the Heinrich-2 (H-2) event, roughly coincident with LGM (Andrews *et al.*, 1998). Substantial net recession occurred along the aquatic south-east Laurentide margin between Cape Cod and Newfoundland starting about 18 ka BP (Gipp & Piper, 1989; Mosher *et al.*, 1989; Piper *et al.*, 1990; King, 1996; Schnitker *et al.*, 2001; Miller *et al.*, 2001), perhaps in response to the initial eustatic sea-level rise, if not to warming. Ice lobes south of the Great Lakes oscillated during slow net recession from 18 to 16 ka BP (Mickelson *et al.*, 1983; Johnson, 1986; Matsch & Schneider, 1986) evidently in tune with the stadial-interstadial fluctuations recorded in the Greenland (GISP-2) ice core (Lowell *et al.*, 1999). These lobes then retracted dramatically by the culmination of the Erie Interstadial, which is estimated to have occurred about 16.5-15.5 ka BP (Barnett, 1992) or 16.0-14.5 ka BP (Ridge, 1997), but is nowhere dated. It is placed here at 15.5 ka, following Barnett (1992) and earlier sources. At that time, the Lake Erie basin is thought to have held glacial Lake Leverett (Morner & Dreimanis, 1973; Barnett, 1992), which is thought to have drained eastwards, probably via the Mohawk Valley in New York. Subaerial erosion and fluvial gravel deposition occurred in the Mohawk Valley prior to deposition of Valley Heads Drift (Ridge *et al.*, 1991; Ridge, 1997). Lake Leverett drained into an early phase of Lake Albany in the Hudson Valley, which in turn drained into a large lake in Long Island Sound (Ridge, 1997). The southern part of Lake Hitchcock had formed in the Connecticut River Valley by this time (Ridge & Larsen, 1990). There is no evidence of large recession of the James and Des Moines lobes during the interval leading up to the peak of the Erie Interstadial. Indeed, wood dated 15.775 ka BP below till in Iowa (Bettis *et al.*, 1996) may be evidence of readvance shortly thereafter. On the other hand, an AMS date of 15.67 ka BP on wood from basal lake sediment in south-western Alberta (Beierle & Smith, 1998) indicates that initial decoupling of Laurentide and Cordilleran ice had begun by that time.

Readvance after Erie Interstadial (15.5 or 14.5 to 14 ka BP)

The readvance of Great Lakes ice lobes during the Port Bruce Stadial (after Erie Interstadial) was massive, evidently producing the only net increase in continental ice area during deglaciation (Fig. 5). The combined Erie and Huron lobes readvanced to the Powell and Union City moraines in Indiana and Ohio by about 14.5 ka BP (Ogden & Hay, 1965; Fullerton, 1980; Mickelson *et al.*, 1983). Features and deposits assigned to the Erie Interstadial in the Mohawk Valley, New York are overlain by Valley Heads

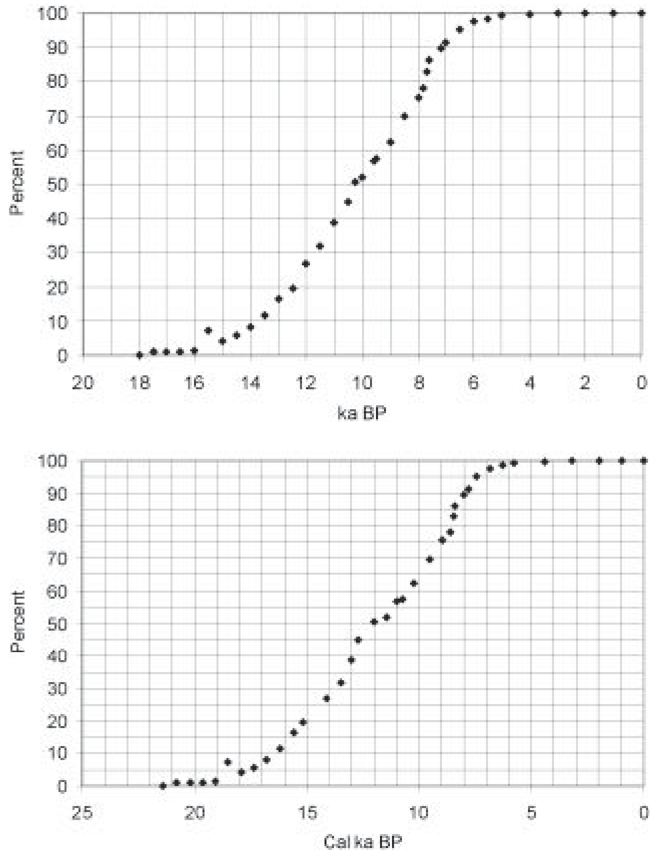


Fig. 5. Percent area deglaciated in North America from Last Glacial Maximum to present, plotted in radiocarbon time scale (a) and calendar time scale (b).

Drift, which terminates at the Valley Heads moraines dating about 14 ka BP (Ridge, 1997). If these moraines mark the limit of the Port Bruce stadial readvance, it would be more convenient to place the Erie Interstadial at 14.5 ka BP than at 15.5 ka BP, and doing so would eliminate the large re-expansion of ice area shown in Figure 5. Because Erie Interstadial herein is placed more conventionally at 15.5 ka BP, however, post-Erie 15 and 14.5 ka BP ice margins are shown south of the Valley Heads position. Alternatively, these ice margins might have been placed between the Erie Interstadial minimum ice margin and the Valley Heads position. Nevertheless, currently there is a lack of chronological evidence for more ice-free terrain than is shown at 15 and 14.5 ka BP in New York.

The large readvances of Great Lakes ice lobes during the Port Bruce Stadial were presumably a response to climate change. If so, correlative readvances might be expected to have occurred elsewhere. There do not appear to have been correlative readvances around the rest of the ice margin at 15 ka BP. Several dates on sub-till wood of the Des Moines Lobe allow for readvances throughout the interval 15.5-14 ka BP, but these are most simply interpreted as the result of a single readvance to the Bemis moraine culminating about 14 ka BP (Clayton & Moran, 1982; Ruhe, 1983; Bettis *et al.*, 1996; Lowell *et al.*, 1999), approximately correlative with the Valley Heads readvance(s). Apart from the continued advance of the Puget Lobe of Cordilleran ice at 14.5 ka BP (Porter & Swanson, 1998) and readvance(s) of Cordilleran ice on Kodiak Island, Alaska at 14.5-14 ka BP (Mann & Peteet, 1994), the most significant ice-marginal event was the re-expansion of the Hudson Strait ice stream to the sill at the mouth of Hudson Strait at about 14 ka BP. This readvance is dated in proglacial marine sediment; it was presumably the result of general thickening of central Laurentide ice subsequent to H-2 drawdown, thus allowing release of ice bergs culminating in the H-1 event (Andrews *et al.*, 2001). Finally, large regrowth of ice on the continental shelf south of Newfoundland emplaced till above marine sediment dated at 14.3 ka BP (Miller *et al.*, 2001), and substantial end moraines were under construction by the south and west aquatic margins of the Newfoundland Ice Cap at about 14 ka BP (Shaw, 2003). Hence a case can be made for shifting the Erie Interstadial about 1 ka younger than its traditional placement, thus allowing Port Bruce stadial correlatives to be more clearly recognised elsewhere on the continent. This would bring the onset of Port Bruce Stadial into correlation with the sharp temperature decline seen in the Summit record at about 14.2 ka BP (Johnsen *et al.*, 1992). Perhaps contrary to that interpretation, however, substantial ice recession appears to have occurred in the Gulf of St Lawrence and Gulf of Maine regions between 14.5 and 14 ka BP. However, this can be seen as a calving response to rising sea level, and furthermore recession in the Gulf of St Lawrence could be reduced to about half of that shown with retirement of a single date from Anticosti Island (Gratton *et al.*, 1984), the only currently accepted date in the inner gulf that exceeds 14 ka BP.

The Late-glacial Interstadial (13.5-11 ka BP)

Following Lowe *et al*. (1994), this interval is referred to informally as the Late-glacial Interstadial, which encompasses the Bølling and Allerød warm intervals and intervening, brief cold event, the Older Dryas. The interval shows the first clear, widespread signals of climate warming in Europe and eastern North America and it correlates with a northwards placement of the oceanic polar front in the North Atlantic (Ruddiman & McIntyre, 1981). This was an interval of progressive net reduction of ice area in North America, while the ice sheets remained nevertheless very large. The interval is divided into locally defined stadials and interstadials in the Great Lakes region, as discussed below. The now-classic glacial lake sequences evolved in the Great Lakes basins and Cordilleran ice became detached from the Laurentide Ice Sheet during this interval. However, only minimal recession of Arctic ice margins occurred.

The ice-free corridor

Perhaps no topic has garnered more interest in discussions of North American glacial history than the chronology of openings and closings of the transient ice-free corridor between the Cordilleran and Laurentide ice sheets. This interest has entirely to do with debates about the early or first peopling of the Western Hemisphere and about what routes might have been available to these people. At times, the geologicalal history of the corridor seems to be viewed by archaeologists as being firmly enough known, or at least the timing of its last opening is so known, that categorical conclusions are warranted. For example, Dixon (2001: 277) stated, “Research dating the late Pleistocene deglaciation indicates that terrestrial connections between eastern Beringia and areas south of the North American continental glaciers were not reestablished until about 11,000 BP (Jackson *et al.*, 1997). This precludes a mid-continent route for human entry until ca. 11,000 BP.” Dixon (1999) made similar assertions citing Jackson & Duk-Rodkin (1996). If this history were firmly established, it could indeed be concluded that people must have arrived in southern North America and in South America by some other route, because it is generally admitted that Clovis people had been in these areas since 11.5 ka BP and possibly, though arguably, pre-Clovis people had been there for 1-1.5 ka before that time.

Dyke & Prest (1987a-c) showed fully coalescent Cordilleran and Laurentide ice sheets at LGM, then a controversial issue (cf. Stalker, 1977). They showed the ice-free corridor first opening at 13 ka BP and placed the Laurentide ice front 200-600 km east of the Cordilleran mountain front by 11 ka BP. Age control was admittedly weak, especially for the initial opening, being based on long-distance correlations with better-dated margins to the north and south. Ice margins internal to the corridor region were based on glacial lake and lake-spillway sequences (e.g., St-Onge, 1972; Mathews, 1980). What, however, is the newer evidence for keeping the two ice sheets coalescent until 11 ka BP? Jackson *et al.* (1997), cited above, presented eight ^{36}Cl dates on boulders of the Foothills Erratics Train, which was formed by coalescent Laurentide and Cordilleran ice flow. These ages range from 17.6 ± 4.5 ^{36}Cl ka BP to 12.0 ± 0.6 ^{36}Cl ka BP, with no significant trend of age along flowline. Jackson *et al.* (1997) used these data only to conclude that the ice sheets had coalesced during the Late Wisconsinan and hence that people could not have traversed an ice-free corridor during time of coalescence. Furthermore, Jackson & Duk-Rodkin (1996), in explicitly considering the chronology of the ice-free corridor, stated “Their [Dyke & Prest (1987)] 12 000 BP ice-marginal position is independently corroborated through archaeological evidence as the eastern and northern limits of known occurrences of fluted points (*circa* 11 500 BP; Haynes, 1980) occurs tightly within it (Wilson, 1983).” Clearly then, the cited sources do not contain positive evidence of continued coalescence until 11 ka BP. To allow such coalescence until 11 ka BP, several 10s of currently-accepted minimum-limiting radiocarbon dates on deglaciation would need to be rejected.

Unfortunately, the initial opening of the ice-free corridor remains only tenuously dated. The initial opening at the south end probably started about 15 ka BP, based on an AMS date of 15.67 ka on wood, mentioned above (Beierle & Smith, 1998) and exposure dates on the Laurentide terminal moraine in the south-western Alberta Foothills (Jackson *et al.*, 1999). By 13.5 ka BP, the southern half of the corridor seems to have opened, because wood (evidently small wood, probably arctic willow; P. Bobrowsky, personal communication, 2002) from basal sediment of glacial Lake Peace yielded an AMS date of 13.97 ka BP (Catto *et al.*, 1996). It is possible that the corridor was entirely open by 13.5 ka BP, because the dated site is located midway within it. However, in the absence of similarly old dates from the northern part of the corridor, and assuming slower ice ablation further north, initial joining of the southern and northern approaches of the corridor may not have occurred until 12.5 ka BP or possibly even 12 ka BP. It seems exceedingly unlikely, however, that ice coalescence could have continued until 11.5 ka BP, for the Mackenzie Lobe of Laurentide ice had by that time receded halfway up the Mackenzie Valley (Mackay & Mathews, 1973; Smith, 1992). Furthermore, south-eastern Cordilleran ice had by then receded into the alpine zone (Reasoner *at al.*, 1994), and Cordilleran ice distribution in Alaska was close to the present one. In summary, the known history of the ice-free corridor, although imprecise, does not preclude the possibility of pre-Clovis people using this route and its availability to early Clovis people is almost certain.

Recession of ice in the Alberta portion of the ice-free corridor during the Late-glacial Interstadial was not accompanied by significant end moraine construction. However, recession of the Mackenzie Lobe was interrupted by readvances to the Tutsieta Lake moraine, its correlative the Sitidigi Lake moraine, and then the Kelly Lake moraine (Hughes, 1987). These moraines are not dated directly. They are here placed in the deglaciation sequence at 12.5 ka and 12 ka BP, respectively, based on their correlation with moraines of the Amundsen Lobe along the Arctic coast and a minimum-limiting age of 11.53 ka for the younger moraine (Smith, 1992). Further recession of the Mackenzie Lobe did not involve moraine construction until Younger Dryas time. Lake Mackenzie formed above The Ramparts (a canyon cut through a bedrock ridge during lake drainage) of the Mackenzie River about 12 ka BP and lengthened as the ice receded (Smith, 1992; Lemmen *et al.*, 1994).

Great Lakes-Champlain Sea

The sequence of ever-shifting glacial lakes in the Great Lakes basins started just prior to 14 ka BP. Detailed sequences of lakes at various levels, related to various outlets during intervals of recession and readvance have been carefully worked out over decades of detailed study. The full sequence can only be shown in time steps of 100 years or finer, but the dating control scarcely warrants such resolution. The 500-year steps shown here thus capture only parts of the sequence, and the reader is referred to Karrow

& Calkin (1985) and to Barnett (1992) for fuller details. The Erie-Huron basin lake sequence is particularly complex because at times of greater ice recession water was able to flow eastward to the Lake Ontario basin, whereas at times of readvance eastern outlets were closed and lake levels rose back to older shorelines and outlets. For example, readvance of the Ontario and Huron lobes during the Port Huron Stadial about 13 ka BP terminated eastward drainage of low-level Lake Ypsilanti (<13.5 ka BP) in the Erie basin and re-established westward draining Lake Whittlesey (see 13 ka BP map). The next major change was establishment of Early Lake Algonquin in the Michigan and Huron basins during the Two Creeks Interstadial, about 12 ka BP. This lake drained southward at Port Huron into Early Lake Erie, which in turn drained into Lake Iroquois in the Ontario basin, and thence to the Hudson River in New York. Subsequent readvance of the Michigan Lobe across the Two Creeks forest bed in Wisconsin (Schneider, 1990) and its age-equivalent bryophyte bed at Cheboygan, Michigan (Larsen *et al.*, 1994) at 11.8-11.5 ka BP severed the water connection between Michigan and Huron basins, temporarily returning the former to Mississippi drainage. The large readvance of the Michigan Lobe implies at least a minor readvance of the adjacent Huron Lobe. There is no evidence, however, of contemporaneous readvance of the Ontario Lobe. Instead, the Ontario Lobe waned and retreated into the Ottawa Valley by 11.5 ka BP, initially allowing Lake Vermont, successor to Lake Iroquois, to extend far to the north-east, and then to be suddenly replaced by the Champlain Sea.

The Two Rivers advance of the Michigan Lobe across the Two Creeks forest bed has been interpreted as a surge event resulting from dynamic feedback within the ice sheet, because the event falls within a warm interval prior to the Younger Dryas (Wright, 1971). The Sumas readvance of Cordilleran ice in the Fraser Lowland, underway by about the same time, has been similarly interpreted (Clague *et al.*, 1997). However, the Sumas event has now been divided into several readvances extending through a longer interval (Kovanen, 2002; Kovanen & Easterbrook, 2002; see below). Furthermore, Lowell et al. (1999) suggest that the Two Rivers advance may correlate with the short Older Dryas cooling trend seen in the GISP2 record.

The Champlain Sea chronology has been controversial over the years, especially when molluscs dating as old as 12.8 ka BP (reservoir corrected by –400 years) came to light from the Ottawa Valley (Richard, 1975; 1978). This chronology was rendered all the more problematic when the New England varve chronology was securely dated by radiocarbon, because an implication of that chronology is that Laurentide ice still retained glacial Lake Vermont, which necessarily preceded the Champlain Sea, until 11.4-11.7 ka BP (Ridge *et al.*, 1999). The new marine reservoir corrections applied in the current re-analysis, and especially the rejection of numerous Champlain Sea dates on deposit feeding molluscs, bring the Champlain Sea chronology into acceptable agreement with the New England varve chronology. The most extended phase of Lake Vermont, that which reached Ottawa and was figured by Dyke & Prest (1987b) at 12 ka BP (also referred to as Lake Candona because of its ostracod assemblage and Lake St Lawrence (Rodrigues, 1992)), can thus now be assigned an age of 11.6-11.7 ka BP.

The Younger Dryas Chronozone (11-10 ka BP)

Perhaps the most significant reinterpretation of Laurentide deglaciation history involves reassignment of the ages of important moraines to the Younger Dryas interval. For example, the largest end moraine belt along the north-western margin of the Laurentide Ice Sheet, including the Bluenose Lake moraine system on the Arctic Mainland and its correlative on Victoria Island, earlier thought to date 12.5-11 ka BP, is now firmly dated 11-9.6 ka BP (Dyke & Savelle, 2000; Dyke *et al.*, 2003). Moraine construction occurred here throughout the Younger Dryas Chronozone and into the subsequent Preboreal cold event, typically as the result of lobe to sublobe scale readvances. Further north, the Viscount Melville Sound Lobe readvanced to form the distinctive Winter Harbour Till, which is thought to have been laid down around the grounded fringe of a 60 000 km^2 ice shelf (Hodgson & Vincent, 1984; Hodgson *et al.*, 1984). Shells underlying that till or reworked into it date 11.46-10.66 ka BP using a reservoir correction of –760 years. An ice stream in Cumberland Sound, Baffin Island, is thought to have readvanced early in Younger Dryas time to the sill at the mouth of the Sound, sending distinctive black mud down the continental slope where it is dated in sediment cores (Jennings *et al*, 1996; Andrews *et al.*, 1998). Furthermore, the massive Gold Cove readvance of Labrador ice across the mouth of Hudson Strait and outer Frobisher Bay on Baffin Island culminated at the end of Younger Dryas time or very shortly thereafter (Miller & Kaufman, 1990). Readvances of Laurentide ice into the Champlain Sea (LaSalle & Shilts, 1993) have maximum-limiting dates of 11-10.7 ka BP, using a –800-year reservoir correction. These readvances occurred before construction of the St-Narcisse moraine (about 10.5 ka BP), a major readvance feature initially also interpreted as a response to Younger Dryas cooling (LaSalle & Elson, 1975), but later argued to represent a dynamic response to opening of the Champlain Sea (Hillaire-Marcel *et al.*, 1981). The interpretation of climate forcing is now favoured again, and the Mars-Batiscan moraine, just behind the St-Narcisse, is assigned a terminal Younger Dryas age (Bhiry *et al.*, 2001).

Many other middle deglaciation moraines are currently assigned to the Younger Dryas interval herein, though most are not closely dated. These include readvance moraines at the mouth of Navy Board Inlet on Baffin Island, with a minimum age of 9.7k ka BP (Dyke & Hooper, 2001); the Brador moraine (minimum 10.4 ka BP) and Paradise moraine (minimum 9.81 ka BP) of Labrador (Fulton & Hodgson, 1979; Grant, 1992); some of the moraines along the Québec North Shore (Dubois & Dionne, 1985); the Cartier and McConnell moraine sequences of central

Ontario; moraines marking the culmination of the Marquette readvance along the south shore of Lake Superior, well-dated at 10 ka BP (Lowell *et al.*, 1999); the Eagle-Finlayson (about 11 ka BP), Hartman, Lac Seul, and Sioux Lookout (minimum 9.74 ka BP) moraine sequence of north-western Ontario; the Pas moraine in Manitoba, and the Cree Lake, High Rock Lake, and Slave moraines in Saskatchewan and Alberta (Bednarski, 1999). Ice was probably at the Cree Lake moraine by 10 ka BP, because the Clearwater spillway in front of it was transporting wood by 10.3 ka BP (Smith & Fisher, 1993) has a drying up about 9.91 ka BP (Fisher & Souch, 1998). On the other hand, Thorleifson (1996) argued that the 10 ka BP ice margin in Manitoba had to reach Riding Mountain to impound Lake Agassiz at the level indicated by the extent of red lacustrine clay in Ontario, deposited pursuant to the Marquette advance. This interpretation would make the Pas moraine younger than 10 ka BP and the Cree Lake moraine younger still.

Other glaciers also responded to Younger Dryas cooling. The Newfoundland Ice Cap readvanced to form the Ten Mile Lake moraine about 10.5 ka BP (Grant, 1992, 1994). In Nova Scotia, several widespread occurrences of Allerød peats are buried by till, glaciofluvial and glaciolacustrine deposits related to re-expansion of ice caps in that province and on Prince Edward Island starting at, or shortly after, 11 ka BP and culminating about 10.5 ka BP (Mott *et al.*, 1986; Stea & Mott, 1989; Mott & Stea, 1993; Stea, 2001). The pollen stratigraphical record in Nova Scotia shows an unambiguous climate reversal during the Younger Dryas, with Allerød shrub tundra reverting to herb tundra and with boreal woodland reverting to shrub tundra (Mott & Stea, 1993); the clearest such record in Canada. Readvances of outlet valley glaciers of the Gaspé Ice Cap south of the St Lawrence estuary incorporated shells dating 11.59-10.8 ka BP, using a reservoir correction of –610 years (Hétu & Gray, 2000). Finally, several of the Sumas readvances of Cordilleran ice in the Fraser Lowland of British Columbia and Washington are placed in the Younger Dryas by Kovanen and Easterbrook (2001), as is a readvance of Cordilleran ice somewhat further north in Howe Sound (Friele & Clague, 2002b). Alpine glaciers also evidently advanced in the Olympic Mountains of Washington during the Younger Dryas Chronozone (Porter, 1978; Begét, 1981; Davis, 1988; Kovanen & Easterbrook, 2001), as they did in the southern Alberta Rockies (Reasoner *et al.*, 1994).

In summary, a substantially different view of the North American glacial record of Younger Dryas cooling has largely emerged in the last 15 years. This event scarcely warranted comment in the last comprehensive treatment of North American deglaciation (Fulton, 1989). It now appears that it left an end moraine and readvance record in North America that rivals the one long recognised in Scandinavia.

The changing discharge routes of glacial Lake Agassiz are repeatedly implicated in discussions of possible mechanisms of Younger Dryas cooling (Broecker *et al.*, 1989). Retraction of the Superior Lobe of Laurentide ice after 11 ka BP allowed glacial Lake Agassiz discharge to switch from the Mississippi River to the St Lawrence River, the switch being complete by 10.81 ka BP (Bajc *et al.*, 2000). The theory is that this sudden discharge of freshwater into the North Atlantic caused density stratification of the water, shutting down thermohaline circulation (bottom-water production). Eastward drainage continued until it was blocked by ice during the Marquette readvance, which culminated at 10 ka BP. By that time the isostatically depressed Clearwater spillway was deglaciated and briefly carried Agassiz water northward to the Arctic Ocean via Lake McConnell, which was by this time nearly as large as Lake Agassiz (Smith & Fisher, 1993, Fisher & Smith, 1994; Lemmen *et al.*, 1994; Fisher *et al.*, 2002). Uplift of the Clearwater spillway or ice readvance then returned Agassiz discharge to the Mississippi until recession of ice north of Lake Superior opened lower outlets near Lake Nipigon (Zoltai, 1967; Teller & Thorleifson, 1983). Detailed calculations have been made of the volumes of freshwater passing through these various routes (Teller, 1990; Leverington *et al.*, 2000; Teller *et al.*, 2002) and numerical modelers have simulated the collapse of thermohaline circulation with these forcings (Rutter *et al.*, 2000). The eastward diversion of Agassiz discharge just after 11 ka BP suddenly released about 9500 km^3 of water.

Post-Younger Dryas recession

Great Lakes, lakes Barlow and Ojibway, and later phases of Lake Agassiz

Lake Algonquin, which occupied the upper Great Lakes basins and drained by a series of outlets, first at Fenelon Falls (Trent River) and then in the Algonquin Park-North Bay region in Ontario, came to an end about 10 ka BP with deglaciation of the Mattawa Valley. Nine shorelines, each related to a different outlet, followed the main phase of Lake Algonquin (also referred to as post-Algonquin lakes). Tracing of shorelines to moraines is key to understanding the detailed pattern of ice recession (Barnett, 1992). The low elevation of the isostatically-depressed North Bay outlet to the Mattawa and of other sills upstream to the west allowed lakes to form in the Huron and Michigan basins that were entirely below present lake levels. These lakes then experienced transgression as the outlets uplifted. Additional fluctuations of level were superimposed by hydraulic back-flooding from meltwater draining along the Ottawa River, causing a high-level Lake Mattawa to inundate the Huron and Michigan basins to a common level during most of the interval 9.6-8.2 ka BP (Lewis *et al.*, 1994).

Recession of ice following the Marquette readvance had cleared all but the north shore of Lake Superior by 9.6 ka BP (Teller & Mahnic, 1988). The Superior basin was then occupied by Lake Minong, which drained eastward to the Huron basin. Further ice recession had cleared Lake Nipigon by 9 ka BP, initiating the final eastward discharge of Lake Agassiz via a series of spillways in that area. A

1600-year varve section at Dorion, on the north shore of Lake Superior, which is tied into the Lake Superior varve chronology of Mothershill (1984, 1988), contains two intervals of thicker varves (the earlier prominent, the other subtle) that have been correlated with glacial readvances to the Nipigon and Nakina moraines at 9.2 and 8.2 ka BP, respectively (Teller & Mahnic, 1988; Thorleifson & Kristjansson, 1993; note that these ages are in radiocarbon years; in the original sources the top of the sequence is fixed in radiocarbon years, but varve years are used to date events below that). The earlier of these advances, which refilled the Nipigon basin with ice, is implicitly accommodated in the sequence of maps presented herein; the other seems problematically young for an ice margin that far south but is not impossible.

Lake Barlow formed in the upper Ottawa River drainage due to isostatic reversal of the river gradient (Vincent & Hardy, 1979; Veillette, 1994) and perhaps due to drift obstruction as well (Lewis *et al.*, 1994; Veillette, 1994). An early phase of the lake is here placed at 9.6 ka BP, somewhat younger than previous age assignments, which relied on basal lake sediment dates that probably contain hard-water effects. The ice margin then stood at the Roulier moraine in Québec and at correlative moraines in northern Ontario. Lake Ojibway, northward successor to Lake Barlow but with its spillway at the Hudson Bay-Ottawa drainage divide, is shown here as starting about 9 ka BP, based on currently accepted minimum-limiting dates.

The westward expansion of Lake Ojibway across Ontario is poorly understood and poorly dated. The most useful age control currently available is the approximate varve chronology at Sachigo Lake (Satterly, 1937; Elson, 1967), which requires deglaciation and sedimentation in north-eastern Lake Agassiz (54°N, 94°W) by 7.9 ka BP. This date suggests a minimum age for the Big Beaver House moraine, constructed by ice that covered Sachigo Lake, of about 8 ka BP, the age applied here. This is the youngest moraine in northern Ontario. The eastward flow of Agassiz water into Lake Ojibway, figured here as starting at 8.5 ka BP, may be too early by as much as 500 years if the 8.2 ka BP age for the Nakina moraine near Lake Nipigon, discussed above, is correct.

The Cochrane ice surges, named Cochrane I, Rupert, and Cochrane II in Québec, (Hardy, 1982), into Lake Ojibway from a source over Hudson Bay are tied into the Barlow-Ojibway varve sequence. The interpretation of this event in Ontario is less clear, though ice streaming into Lake Ojibway from Hudson Bay at this time is accepted (Thorleifson *et al.*, 1993) and the conventional stratigraphy of Cochrane Till intercalated in Ojibway varves is upheld by Paulen (2001). Cochrane I started 400 years before drainage of the lake upon deglaciation of Hudson Bay. The latter event (see below) is now reasonably well fixed at 7.6 ka BP. Hence, Cochrane I is placed at 8 ka BP. By that time, a prong of Lake Ojibway must have extended northwards between Hudson and Québec ice masses to allow for the south-eastward and eastward-directed surges. This prong presumably elongated from that time until formation of the Sakami moraine (see below), because lacustrine sediments have been identified as far north as La Grande Rivière (Hardy, 1977) and possible lacustrine sediments are mentioned further north below marine sediment in Hudson Bay (Gonthier *et al.*, 1993). The Cochrane I margin is here correlated with the Pinard moraine to the west in north-eastern Ontario and with the Big Beaver House moraine.

The Barlow-Ojibway varve sequence contains 2110 known varves. Setting the drainage of the lake at 7.6 ^{14}C ka BP, varve 1 of Lake Barlow would date about 9.3 ^{14}C ka BP (i.e., 7.6 ^{14}C ka BP = 8.4 cal ka BP; 8.4 + 2.11 = 10.53 cal ka BP = 9.3 ^{14}C ka BP). Hence, the date of 9.6 ka BP for start of Lake Barlow is reasonable, either assuming some missing varves or considering the uncertainty in the radiocarbon dating.

The evolution shown for lakes Agassiz and Ojibway after 8 ka BP is necessarily more speculative. However, final northwards drainage of these lakes into Tyrrell Sea (Hudson Bay) has been long accepted (e.g., Boissonneau, 1966; Skinner, 1973) and there now is a better understanding of the deglaciation of Hudson Bay itself thanks to the geophysical surveys and mapping by Josenhans & Zevenhuizen (1990). They mapped glacial bedforms, subglacial drainage channels of extraordinary length, and fields of large-scale, arcuate, iceberg scours, all of which yield a coherent model of deglaciation and final lake-drainage events. This model is incorporated here in the maps for 7.8 and 7.7 ka BP, the final centuries before incursion of the Tyrrell Sea. Of key importance are the arcuate to spiral iceberg scours. These features are limited to a corridor in western Hudson Bay, which is not determined by seabed topography. Josenhans & Zevenhuizen inferred that the sides of the corridor represent the ice-sheet margin at the time of scour formation, hence defining a deep calving bay. They further concluded that the scours might represent the large-scale swirling motion of bergs entrained in a subglacial outburst of Lake Agassiz or swirling due to tidal motion in the calving bay. The swirls clearly translate down-bay through time, and because tides should have moved bergs inwards as well as outwards, the former mechanism seems a better explanation. This interpretation of a calving bay is reasonable, not only because of the unique iceberg scour patterns, but because the latest ice flow features are compatible with the assumed ice margins. Furthermore, large meltwater channels located below marine limit, therefore probably subglacial, near the Ontario-Manitoba border (Prest *et al.*, 1968) are likely northward discharge routes of glacial lake water (e.g., Dredge, 1983; Klassen, 1983) into this calving bay. The initial drainage of combined lakes Agassiz-Ojibway may thus have been into western Hudson Bay, rather than into eastern Hudson Bay, as usually assumed.

Furthermore, a case can be made for more than one northward flushing, or partial flushings, of these gigantic lakes. Klassen (1983) mapped the Fiddler beach of Lake Agassiz, which is too low to correlate with the youngest phase of Lake Ojibway that drained into the Ottawa River.

Thorleifson (1996) portrayed the Fiddler phase of Agassiz draining ice-marginally into Tyrrell Sea in eastern Hudson Bay, and Teller *et al*. (2001) similarly indicated that two bayward drainages were necessary to account for Fiddler beach. Meanwhile, Veillette *et al*. (1999, 2003) have mapped beaches of Lake Ojibway near modern Lake Abitibi and younger beaches north of Lake Abitibi, which also are too low to relate to an Ottawa River outlet, and must therefore represent a regressive pause or a transgression after the initial northwards drainage. In the same region, Veillette has also identified a thick silt layer within upper Ojibway varves, which may represent a drainage event (drainage varve), followed by a lacustrine transgression that deposited the final metre of clay. The maps for 7.8 ka and 7.7 ka BP, therefore, postulate initial northward drainage (or lowering) of Agassiz-Ojibway immediately after 7.8 ka BP, stabilisation of the lakes at the Fiddler-Abitibi level, or refilling of the lakes to this level, at 7.7 ka BP, and final drainage immediately thereafter. The route shown for final drainage is based on the Ontario-Manitoba border channels mentioned above. However, large subglacial channels in Hudson Bay off the Winisk River (Josenhans & Zevenhuizen, 1990) are equally probable routes, or routes of closely subsequent flushes.

The new marine shell reservoir correction for Hudson Bay places the incursion of Tyrrell Sea, hence final and catastrophic drainage of Agassiz-Ojibway at 7.6 ± 0.1 ka BP. Inundation of southern Hudson basin was essentially instantaneous, as indicated by indistinguishable earliest marine shell ages from east of James Bay, south-west of James Bay, and in northern Manitoba. At time of Tyrrell Sea incursion, the western margin of the Québec-Labrador ice cap stabilised at the 800-km long Sakami moraine because of the great water depth reduction at the ice front that accompanied drainage of Lake Ojibway (Hardy, 1982). The contemporaneous ice margin in north-eastern Hudson Bay at that time still lay west of the Ottawa Islands, where well-dated marine limits are younger than 7.6 ka BP (Andrews & Falconer, 1969). Within the next century, and while ice still held at the Sakami position, the central Hudson margin of Québec-Labrador ice retreated to the Nastapoka Hills, a cuesta-like arrangement of hills along the shoreline (Lajeunnesse & Allard, 2003). No similar stabilisation upon lake drainage has been recognised along the south margin of Keewatin ice, but this is perhaps due to its shorter aquatic margin in final Lake Agassiz.

Current dating of Tyrrell Sea incursion brings this event into correlation with the 8.4 cal ka BP beginning of the so-called '8.2 ka cold event' seen in Greenland ice cores (Alley *et al.,* 1997), in North Atlantic marine sediment cores, in European lake sediment records and in North American palaeoclimate records. On this basis, Barber *et al.* (1999) proposed that the sudden flushing of Agassiz-Ojibway water into the North Atlantic via Hudson Strait was the causal mechanism of the cold event. Similarly, the two-pronged structure of this event (Baldini et al., 2002) may be explained by the double-flush sequence outlined above, in which case the two flushes were separated in time by only about a decade. The 8.2 ka event was only one-third the duration and the temperature amplitude of the Younger Dryas cooling, despite the fact that the volume of freshwater involved exceeded by an order of magnitude the amount involved in initial Agassiz diversion to the North Atlantic at the beginning of the Younger Dryas (163 000 km^3 versus 9500 km^3 according to Leverington *et al.,* 2000, 2002), although sustained Agassiz discharge throughout the Younger Dryas may have been essential in prolonging the cold event. Nevertheless, this difference may indicate that the response of North Atlantic thermohaline circulation to freshwater input is substantially sensitive to input location, with direct injection to the Gulf Stream having larger impact than input to the already cold and less saline Labrador Current.

Arctic Archipelago and Keewatin

The northern tier of the Arctic Islands, north of 75°N, was largely covered by the Innuitian Ice Sheet at the LGM as well as by independent ice caps on Melville and Prince Patrick islands (Fig 4-18 ka; Blake, 1970; Hodgson, 1992; Hodgson *et al.,* 1984; Bednarski, 1998; Dyke, 1999; England, 1999). The wider southern tier of islands, from Banks Island in the west to Baffin Island (*cf.* 11 ka map) in the east were largely covered by the Laurentide Ice Sheet; ice of the Keewatin Sector spread over the western part and ice of the Baffin Sector over the eastern part (Dyke & Prest, 1987a, b). Throughout the Arctic Islands, evidence of ice recession prior to the Younger Dryas interval is limited to the glaciated fringe, with Allerød-age marine shells and plant remains known only from north-western Victoria Island and from Melville, Prince Patrick and Ellef Ringnes islands in the west and from the east coast of Baffin Island. The earliest definitive evidence of ice recession is in the form of moss remains dated to 12.6 ka BP in Donard Lake on the eastern tip of Baffin Island (Moore *et al.,* 2001). Similarly, only a single site in coastal Greenland can be shown to have been ice-free by 11 ka BP (Björck *et al.,* 2002). As discussed above, recession during the Younger Dryas in the western Arctic Islands was slow, punctuated by readvances, and characterised by moraine construction. On eastern Baffin Island, recession during the Younger Dryas was slow to negligible. Moraines that approximate the LGM limit contain ridges that evidently date broadly within the interval 20-10 ka BP (e.g., Marsella *et al.,* 1999).

Most of the Arctic Archipelago, therefore, was deglaciated after the Younger Dryas Chronozone. In many places, the pattern of ice-marginal recession can be reconstructed in fine temporal resolution, because many lateral meltwater channels formed along the margins of cold-based glaciers, perhaps annually (e.g., Dyke, 1999; O'Cofaigh *et al,* 2000). The rate of ice recession appears to have greatly accelerated about 10 ka BP. Innuitian and Laurentide ice were separated by 9 ka BP, but Innuitian ice remained confluent with the Greenland Ice Sheet until about 7.75 ka BP (England, 1999). By 8.5 ka BP, Innuitian ice had fragmented into island-centreed ice caps, restricted to the eastern half of the archipelago, and by 8 ka BP ice in many places had receded to modern ice margins. During the recession of Innuitian ice, pauses occurred in many places, typically at fjord heads, as indicated by the accumulation of thick glacio-marine drift and associated till, collectively referred to as the 'drift belt' on western Ellesmere Island (Hodgson, 1985). These pauses are explained as end-of-calving events and hence have no particular palaeoclimatic significance. The rapid disintegration of the Innuitian Ice Sheet occurred during the Holocene thermal maximum (10-8 ka BP), as identified in the melt-layer record of the ice core from Agassiz Ice Cap on Ellesmere Island (Fisher *et al.*, 1995).

The Keewatin Sector of Laurentide ice similarly receded rapidly in the archipelago after 10 ka BP. It had cleared the region by 8 ka BP. On the Arctic mainland, Keewatin ice constructed the MacAlpine moraine, a large unnamed successor, and the Chantrey moraine (see Dyke & Prest, 1987c for moraine names). These moraines are not directly dated but are here assigned to the interval 8.2-8 ka BP. Being located well inland of marine limit, these moraines represent readvances rather than end-of-calving stabilisations (Dyke, 1984). They are the youngest known end moraines of Keewatin ice, the subsequent recession of which is essentially undated.

The Baffin Sector of the Laurentide Ice Sheet, which spread radially out of Foxe Basin, still held a near-maximum configuration at 10 ka BP. Retreat of outlet glaciers up the east and north coast fjords occurred slowly between 10 and 8.5 ka BP. Extensive sets of end moraines were then constructed around much of the Baffin Sector ice margin between about 8.5 ka and 7 ka BP, including moraines of Cockburn age on Baffin Island (Andrews & Ives, 1978; Dyke & Hooper, 2001) and the impressive Melville moraine along the west side of Melville Peninsula (Dredge, 1990). It would appear that as long as the Foxe Basin ice centre withstood marine incursion from Hudson Bay, the mass balance of the Baffin Sector fluctuated from positive to only slightly negative, thus enabling moraine construction.

Final breakup of the Foxe ice dome remains inadequately dated in detail, but appears to have involved the northwards progression of a calving bay from Hudson Bay during the interval 7-6 ka BP. This break-up left residual ice caps on Baffin Island, Southampton Island, and Melville Peninsula that, along with an ice-cap in Arctic Québec, were the only remnants of the Laurentide Ice Sheet. The Baffin Ice Cap was still a substantial and dynamically active glacier throughout the middle Holocene, capable of inscribing a succession of shifting ice flow patterns and constructing substantial end moraines at intervals throughout (Andrews, 1989; Dyke & Hooper, 2001).

Discussion

The overall pattern of deglaciation presented here remains similar to previous reconstructions. Some novel reinterpretations of the fundamental pattern of ice recession simply violate the chronological data (e.g., Kleman et al., 1995; Jansson *et al.*, 2002). For example, Ungava Bay, proposed as a recession centre by Jansson *et al*., was occupied by the sea at 7 ka BP, as shown by dated molluscs in its south-eastern and south-western parts, when ice still covered most of arctic and subarctic Québec. It could not, therefore, have been the recession centre for that ice sheet.

Improved age control and more detailed mapping of deglacial patterns have had the net result of bringing the North American deglaciation sequence into more evident correlation with the major climatic events recognised in the North Atlantic region and in the Greenland ice cores. Thus the acceleration of retreat at 14 ka BP corresponds to the sudden warming evident in the Summit, Greenland record. More significantly, the Younger Dryas now emerges as an important control of ice-marginal behavior of the North American ice sheets, similar to that shown by the Scandinavian Ice Sheet, a result commensurate with growing recognition of this event in North American pollen records. This correlation hopefully will be greatly tested and clarified in the coming decade as recession chronologies are more firmly established. The need for additional dating in this regard is greatest in the western Arctic mainland region and in the Cordilleran interior, where any recession before 10 ka BP, if such occurred, is entirely undated. Similarly, the prominent 8.2 cal ka BP cold event is now firmly correlated with the deglaciation of Hudson Bay and drainage of lakes Agassiz and Ojibway.

The world's largest ice sheet complex lost <10% of its area prior to 14 ka BP. It then retreated nearly linearly until 7 ka BP, by which time only 10% of the area remained more glaciated than today. This linear reduction of area, as currently understood, was interrupted by two events: a reduced rate of recession during the later half of the Younger Dryas, and an increased rate as ice was clearing from Hudson Bay (Fig. 5). These events are clearer when plotted on the calendar time scale, because the radiocarbon time scale abbreviates the duration of the impact of the Younger Dryas effect in North America (Fig. 5b).

The deglacial chronology accords better than previously with the record of global sea-level rise (Fairbanks, 1989), which features two meltwater pulses separated by reduced melting during the Younger Dryas interval. During meltwater pulse IA (13-11 ka BP), the North American ice sheet complex decreased from about $14.9{\cdot}10^6$ km^2 to $11.5{\cdot}10^6$ km^2. During meltwater pulse IB (10.5-8.5 ka BP) ice area decreased from $10.5{\cdot}10^6$ km^2 to $4.2{\cdot}10^6$ km^2. Paterson (1972) showed that the logarithm of ice area and the logarithm of ice volume are linearly related. This relationship allows a calculation of volume loss for the interval 13-11 ka BP of $10.8{\cdot}10^6$ km^3 (5400 km^3 a^{-1}) and for the interval 10.5-8.5 ka BP of $11.7{\cdot}10^6$ km^3 (5600 km^3 a^{-1}). These rates account for 40% of meltwater pulse IA and 60% of meltwater pulse IB. These estimates are crude in that the entire ice complex is treated as one ice sheet. Nevertheless, the ice complex during the earlier interval was scarcely more complex than the current Antarctic Ice

Sheet (the largest ice sheet constraining the area-volume relationship), with its wide variation in basal topography and complex of ice divides, and during the latter interval, the volume loss was concentrated in the Laurentide Ice Sheet. The relatively early demise of the Eurasian ice sheets accounts for their greater contribution to meltwater pulse IA.

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