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Radiocarbon deglaciation chronology of the Thunder Bay, Ontario area and implications for ice sheet retreat patterns

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ABSTRACT

The sensitivity of ice sheets to climate change influences the return of meltwater to the oceans. Here we track the Laurentide Ice Sheet along a ~400 km long transect spanning about 6000 yr of retreat during the major climate oscillations of the lateglacial. Thunder Bay, Ontario is near a major topographic drainage divide, thus terrestrial ablation processes are the primary forcers of ice margin recession in the study area. During deglaciation three major moraine sets were produced, and have been assigned minimum ages of 13.9 ± 0.2, 12.3 ± 0.2–12.1 ± 0.1, and 11.2 ± 0.2 cal ka BP from south to north. These define a slow retreat (~10–50 m/a) prior to major climate oscillations which was then followed by a factor of ~2 increase during the Bölling–Alleröd, and an additional increase during the early Holocene. When compared to retreat rates in other terrestrial settings of the ice sheet, nearly identical patterns emerge. However this becomes problematic because a key control on retreat rates is the surface slope of the ice sheet and this should vary considerably over areas of so-called hard and soft beds. Further these ice margin reconstructions would not allow meltwater sourced in the Hudson Basin to drain into the Atlantic basin until after Younger Dryas time.

1. Introduction

Although alpine glaciers display rapid responses to climate change (e.g. Lowell, 2000; Oerlemans, 2005) the relationship with ice sheets is less clear. This is especially true during the retreating phase. Do ice sheets have sufficient inertia that they remain immune to climate oscillations or are they highly reactive to them? This issue has relevance as we try to understand the return of meltwater into the oceans at the end of the last ice age, or for the future behavior of the Greenland Ice Sheet responding to current warming trends. We can turn to the geological record of ice sheets and climate change. One approach to such comparisons lies in modeling experiments (e.g. Marshall et al., 2000; Tarasov and Peltier, 2004; Marshall and Koutnik, 2006). The lateglacial was a time of rapid climate oscillations as recorded in Greenland ice cores (e.g. Stuiver and Grootes, 2000; Johnsen et al., 2001) in particular there were rapid and large swings in temperature (e.g. Severinghaus et al., 1998; Grachev and Severinghaus, 2005) the magnitudes of which are under discussion. Recently Kelly et al. (2008) reported on independent glacier systems in the Scoresby Sund region of East Greenland. There the glaciers' fluctuations were in phase with the timings of major changes in the ice core record, however, the glacier equilibrium line reconstructions produced required smaller temperature changes than those interpreted from the ice cores. This relationship supports the seasonality hypothesis (Denton et al., 2005), which posits that annual mean temperatures are driven by a much colder winter temperature. If so, then the summer temperature mean, which is the key for glacier ablation, is smaller.

A complementary approach is to document the history of ice sheets from the geological record. Here we examine the behavior of one sector of the relic Laurentide Ice Sheet near Lake Superior, during the time of lateglacial climate oscillations to document its response and compare it with other regions.

2. Regional setting

2.1. Prior work

Tracking the deglaciation patterns of ice sheets presents numerous challenges. Radiocarbon ages from basal organics

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preserved in depressions are a common tool to reconstruct deglaciation and inland of marine deposits they constitute the vast majority of control points on deglaciation of the Laurentide Ice Sheet (Dyke, 2004). However this approach has been criticized because it is of minimum bracket (e.g. Clayton and Moran, 1982). One factor is any lag time between deglaciation and organic deposition. Although modern observations in the high Arctic demonstrate that vegetation can occupy deglaciated landscape in decades, documenting a similar relationship in the past carries unknowns. Björck (1985, 1990) partly overcame this problem by incorporating pollen data into his analysis to identify sites differing from the regional chronology. Varve chronology with radiocarbon dating for an absolute age control overcomes the minimum limitations (Ridge et al., 1999) but this approach has limited geographical applications.

There have been too few field efforts designed specifically to track the chronology of deglaciation of the Laurentide Ice Sheet (Wright, 1971). One exception near the present study area is Björck (1985, 1990). A basal bulk-sediment date of $11,110 \pm 100$ ^{14}C yr BP (WIS-1327) from Rattle Lake (Site H, Fig. 3), 15 km southwest of the Eagle-Finlayson Moraine (Björck, 1985), has been a key date to reconstruct deglaciation in this region although it rests on the minimum assumption. The next limiting radiocarbon date is a basal bulk-sediment age of 9740 ± 100 ^{14}C yr BP (WIS-1329) from Sioux Pond (Site Q, Fig. 3) in a meltwater channel cutting across the Lac Seul Moraine (Björck, 1985). Farther east, and outside of the Thunder Bay lowland there are radiocarbon ages from north of the Kaiashk Moraine (sites O, P, and Q respectively, Fig. 2) (Teller et al., 2005). Likewise there are an additional 17 control points listed in Table 1 and shown in Fig. 3 along 400 km of ice margin retreat. In the study area (Figs. 1 and 2), a further constraint on deglaciation chronology comes from a long, floating varve record at Steep Rock Lake (Antevs, 1951) recording several hundred years of lake sedimentation.

The moraines in the study area (Fig. 2) have been mapped at regional scales (e.g. Zoltai, 1965; Sado et al., 1995), studied for mineral exploration (Minning et al., 1994), and employed for tracking paleoecological change (Björck, 1985). Moraine orientation indicates that ice recession was from the southwest marked by the Vermilion Moraine and continued to the northeast (Fig. 3). The remaining moraines indicate a generally linear ice margin configuration during deposition of the Eagle-Finlayson and Brule Creek Moraines. Farther north, the lobate outline of the Hartman and Dog Lake Moraines, and the Lac Seul and Kaiashk Moraine indicates contributions from two source areas and cross-cutting moraines indicate readvances during the retreat (Fig. 3).

That the lateglacial climate signals propagate into the middle of North America is demonstrated by pollen and isotope studies. Yu and Wright (2001) reviewed oxygen isotope records within the interior of North America and found a close correspondence to the record from the Greenland ice cores; although the exact climatic response may differ considerably depending upon local conditions at individual sites. Closer to the study area, Hu et al. (1997, 1999) used signals in ^{18}O records during the middle of the Younger Dryas to suggest that a recurrence in spruce (*Picea*) was caused by penetration of cooler air into northern Minnesota.

This study also contributes to another problem. The Thunder Bay area has been hypothesized as meltwater passageway for a catastrophic flood. Based on the extension of strandlines from the central Lake Agassiz basin Teller and Thorleifson (1983) suggested a route crossing the continental divide west of Thunder Bay (near site 11, Fig. 3). Later, Broecker et al. (1989) noted the apparent temporal correspondence (at about 11 ka ^{14}C), of a low stand in the Lake Agassiz basin (Moorhead Phase) with an isotopic change in the ^{18}O signature in the Gulf of Mexico. These observations could be explained by Lake Agassiz waters flowing toward Thunder Bay. However, Zoltai (1965) concluded from field mapping that it was unlikely that meltwater from the Agassiz basin crossed this region. Rather Zoltai (1965, his Fig. 4) showed instead meltwater drainage westward into the Lake Agassiz basin but the timing is not well constrained. For the Thunder Bay area both Lowell et al. (2005) and Teller et al. (2005) noted the lack of geomorphic evidence for a catastrophic eastward drainage. Teller et al. (2005) have explained this lack of evidence by invoking a readvance of the ice sheet margin to remove or bury any evidence of the flood. This study will contribute by constraining the age of the moraines, and hence any ice margin obstacle to the meltwater flow in this region.

2.2. Geography

In practice tracking past history of ice sheets is problematic because during the deglacial hemicycle, pro-glacial lakes and marine waters are common along their margins. Such conditions induce changes in glacier dynamics that may be independent of climatic forcing. Thus we sought a field area deglaciated during the lateglacial and where the impact of pro-glacial water bodies was minimal. The Thunder Bay area of northwestern Ontario was chosen as it lies on the sub-continental divide between the Hudson Bay and Atlantic basins and thus between the major lake basins of Lake Agassiz to the west and Lake Superior to the east.

The study area extends from the Gunflint Range of northern Minnesota to the latitude of the northern shore of Lake Nipigon, Ontario. Elevations range from the level of Lake Superior at about 183 masl to just over 600 masl with a majority of the area lying between 400 and 500 masl. The lowest elevation along the drainage divide connecting the Hudson Bay and Atlantic basins is 350 masl. Bedrock is Archean granites, granidiorites, diorites, and gabbros in places intruded and overlain by Proterozoic mafic volcanics (Wheeler et al., 1996). Overall the surficial drift is sandy and thin, and bedrock structure dominates the topography.

3. Materials and methods

The contribution here is to conduct a directed effort to track the age of deglaciation. We consider that moraines mark major ice margin positions of the Laurentide Ice Sheet during its retreat, and generated chronological brackets for the moraines with new radiocarbon ages. Thus a preliminary geomorphic interpretation of the significant glacial landforms was generated across both the U.S.A. and Canadian portions of the region. This guided our sampling strategy and provided a context for the results. This was achieved with raw elevation data from the SRTM 90m database (http://seamless.usgs.gov) that was processed into a variety of two- and three-dimensional representations. These representations allowed the major moraines to be delineated in a consistent scale and format. A simplified version of the result is shown in Fig. 2.

We undertook an initial reconnaissance trip in May 2003 and subsequent sampling trips in July 2004, January 2005, and February and March 2006. A modified square-rod Livingston corer was used both from lake ice in wintertime, and canoe raft towed by a zodiac boat in summertime. A hydraulic assist permits core recovery directly from the lowest organics layer determined. Generally, the uppermost sediment from each core site was not recovered. In practice at each site we recovered one long core to establish the site stratigraphy, and then 2–4 additional cores across critical contacts to provide sufficient material for ^{14}C analysis. In most cases the core sites were small lakes in bedrock basins or channels. In the individual basins core sites were located in deep water near the edge of the basin to improve chances of recovering terrestrial macrofossils for analysis and dating. In

Table 1
Radiocarbon ages constraining deglaciation.

	Name	Lab number	^{14}C age ^{14}C yr BP	δ^{13}C	Material sampled	Calibrates age range (2σ)	Probability	Reference
Prior work								
A	Kylen Lake	UW-1234	13,900 ± 300	–	Aquatic moss	15,676–17,567	1.00	Birks, 1981; Lund and Banerjee, 1985
B	Weber Lake	W-1763	14,690 ± 390	–	Aquatic moss	16,826–18,654	1.00	Florin and Wright, 1969; Wright and Watts, 1969
C	Mariska Mine	W-1140	11,100 ± 400	–	Wood	11,968–13,843	0.99	Wright and Watts, 1969, from till?
D	Sabin 2	Lu-2507	10,320 ± 170	−29.1	Lake sediment	11,592–12,710	0.95	Björck, 1990
E	Heikkilla	Lu-2556	12,100 ± 150	−22.8	Lake sediment	13,639–14,577	1.00	Björck, 1990
F	Lempia 1	Lu-2555	12,050 ± 240	−25.4	Lake sediment	13,393–14801	1.00	Björck, 1990
G	Lake of the Clouds	Y-2183	9630 ± 110	−27.9	Organic fraction of lake sediments	10,678–11,235	1.00	Stuiver, 1975
H	Rattle Lake	WIS-1327	11,100 ± 100	–	Lake sediment	12,872–13,187	1.00	Björck, 1985
I	Flatrock	TO-11688	8600 ± 80	–	Gyttja	9456–9788	0.98	Teller et al., 2005
J	Head Lake	TO-11695	9300 ± 140	–	Fine vegetation	10,202–10,826	0.94	Teller et al., 2005
K	Flower Lake	TO-11692	10,010 ± 80	–	Gyttja	11,243–11,823	0.99	Teller et al., 2005
L	Rosslyn Brick	GSC-287	9380 ± 150	–	Wood chips	10,248–11,094	1.00	Zoltai, 1965
M	Dorion	Beta-9113	9690 ± 290	–	Fine vegetation	10,250–12,000	0.99	Teller and Mahnic, 1988
N	Grim Lake	AA-58456	8130 ± 50	–	Picea and Larix needles	8992–9153	0.85	Teller et al., 2005
						9159–9257	0.15	
O	Devils Crater	AA-58455	8160 ± 50	–	Larix needles	9008–9262	1.00	Teller et al., 2005
P	Lower Vail Lake	TO-8670	9320 ± 70	–	Fine vegetation	10,291–10,697	1.00	Teller et al., 2005
Q	Sioux Pond	WIS-1329	9740 ± 100	–	Lake sediment	9159–9257	0.99	Björck, 1985
This study								
1	Salo Lake	ETH-30180	14,050 ± 100	−24.1	Aquatics	16,527–16,979	1.00	
2	Wampus Lake	ETH-28944	9245 ± 70	−23.3	Wood pieces	10,368–10,507	0.71	
3	East Chub Lake	ETH-31428	8340 ± 70	−29.5	Plant material	9284–9460	1.00	
4	Little Spring Lake	ETH-30171	9915 ± 75	−22.8	Wood	11,228–11,405	0.88	
		ETH-30170	10,420 ± 75	−28.0	Wood	12,144–12,399	0.78	
		ETH-28945	12,000 ± 85	−21.8	Twigs	13,774–13,949	1.00	
5	Ed Shave Lake	ETH-30181	9055 ± 70	−15.2	Wood	10,159–10,289	1.00	
6	Third Lake	ETH-28946	10,000 ± 75	−13.3	Aquatic plants	11,313–11,620	0.98	
7	Swamper Lake	ETH-29224	9440 ± 70	−16.2	Aquatic plants	10,567–10,771	1.00	
8	Bear Cub Lake	ETH-30182	8910 ± 70	−23.1	Wood	9918–10,092	0.73	
		ETH-28940	9970 ± 75	−20.9	Wood frags	11,267–11,412	0.54	
		ETH-28939	10,550 ± 75	−19.5	Wood frags	12,396–12,520	0.45	
9	Sea Gull Lake	ETH-28941	6850 ± 65	−25.0	Wigs	7614–7742	1.00	
		ETH-28942	9605 ± 70	−22.0	Twigs	10,788–10,973	0.66	
10	Rattle Lake	ETH-32339	10,440 ± 75	−31.6	Wood	12,231–12,404	0.51	
11	Slim Lake	ETH-31006	9535 ± 140	−22.8	Wood	10,679–11,101	1.00	
12	Skeptic Lake	ETH-31003	8370 ± 115	−22.8	Wood	9259–9503	0.97	
13	Loopy Lake	ETH-31004	7800 ± 95	−24.9	Wood	8445–8662	0.88	
14	Sunbow Lake	ETH-31429	10,400 ± 120	−22.3	Plant material	12,076–12,403	0.75	
15	Echo Lake	ETH-31653	9360 ± 90	−21.9	Wood	10,264–10,788	0.97	
16	Kettle Lake	ETH-30178	9345 ± 75	−29.9	Wood fragments	10,436–10,458	0.08	
17	Carson Lake	ETH-30165	9780 ± 75	−27.0	Wood fragments	11,132–11,264	1.00	
18	Crawfish Lake	ETH-31430	10,250 ± 75	−24.6	Plant material	11,821–12,127	1.00	
19	Strip Lake	ETH-31007	9165 ± 85	−24.7	Wood	10,237–10,419	0.98	
20	Bruce Lake	ETH-31008	9080 ± 80	−23.5	Wood	10,177–10,300	0.81	
21	Dexter Lake	ETH-30182	8910 ± 70	−23.1	Wood	9918–10,092	0.73	
22	Mokoman Lake	ETH-32172	9345 ± 75	−23.2	Wood	10,436–10,458	0.08	
		ETH-32171	9510 ± 75	−23.6	Wood	10,681–10,830	0.52	
23	Warnica Lake	ETH-30191	8000 ± 65	−24.8	Wood fragments	8846–8998	0.69	
24	Moon Lake	ETH-31005	8865 ± 75	−26.2	Wood	9887–10,161	0.92	
25	Pretty Lake	ETH-31001	9705 ± 145	−22.1	Wood	11,061–11,243	0.49	
26	Righty Lake	ETH-32182	8605 ± 70	−22.2	Plant material	9471–9745	0.99	

practice this meant that many sites were not directly on moraines but were located between them. Moreover the sampling sites were located on local topographic highs. This put the sites above any glacial lake that may have persisting well after the passage of the ice margin. Core sites are listed in Table 1. Laboratory analysis included loss-on-ignition (LOI) and volume magnetic susceptibility. Procedures for determining percent organic matter (550 °C) and percent carbonate (950 °C) were from Heiri et al. (2001). Magnetic susceptibility was determined with a Bartington MS2E probe at 2 cm intervals, and was used as a proxy for allochthonous terrigenous clastic material in the lake sediment.

The radiocarbon dates were determined at ETH (Swiss Federal Institute of Technology) in Zurich, Switzerland using a Beamline Accelerator. Two dates were processed at Beta Analytical. All radiocarbon ages are calibrated to calendar ages using Calib 5.01 and the IntCal04 calibration curve (Reimer et al., 2004). Radiocarbon ages are listed in Table 1 along with their calibrations to calendar years and within the text two-sigma calibrated ages in square brackets follow the radiocarbon ages. To facilitate comparisons we also represent key ages in the form $X \pm Y$ ka BP where X is the median probability (in thousand years) from the calibration and Y is the two-sigma uncertainty. In the Results section the geomorphology of three moraine sets and their most closely associated radiocarbon ages are discussed from south to north. The details of the core stratigraphy can be found in the Supplemental material. Construction of the time–distance diagram assumed all ages are minimum, thus the constraining envelope was placed where the individual site ages dropped to zero probability.

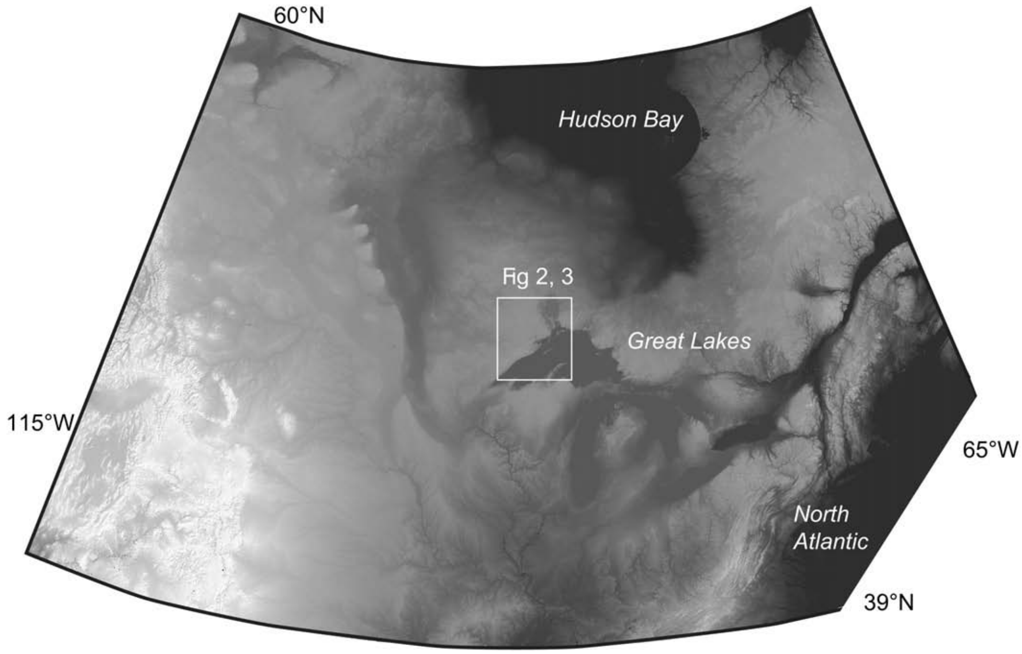

Fig. 1. General setting of the Thunder Bay, Ontario area on the sub-continental divide.

4. Results

4.1. Glacial geomorphology

The primary glacial features of this area are long linear moraines trending across bedrock. Lithologic changes, faults, and other lineations are the primary controls for the distribution of irregular, and often interconnecting lakes. In several areas adjacent to moraines thicker sediment cover partially or totally masks the bedrock with a smoothed texture.

Moraines are represented by linear ridges and they fall into three major sets. Lehr and Hobbs (1992) codified the local names in Minnesota, and Zoltai (1965) did likewise for those in northwest Ontario. We retain these names here for consistency although some of our correlations differ as noted below. We group some of these

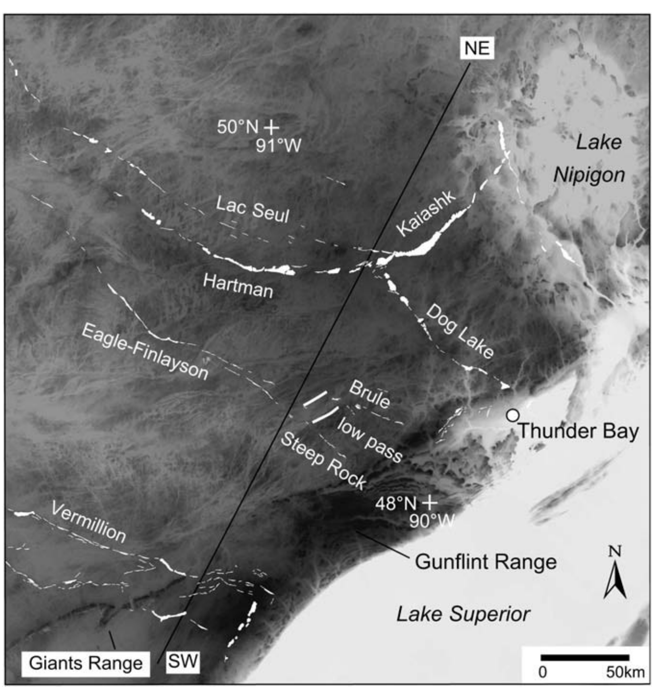

Fig. 2. Grayscale representation of the topography of the Thunder Bay area. Lighter tones are low elevation and darker tones are higher elevation. Elevations in this area range from 183 at Lake Superior to 601 m in the Gunflint Range. Low pass outlines the lowest portion of the continental divide between the present Hudson and Atlantic basins shown. Location of transect in Fig. 7 is shown.

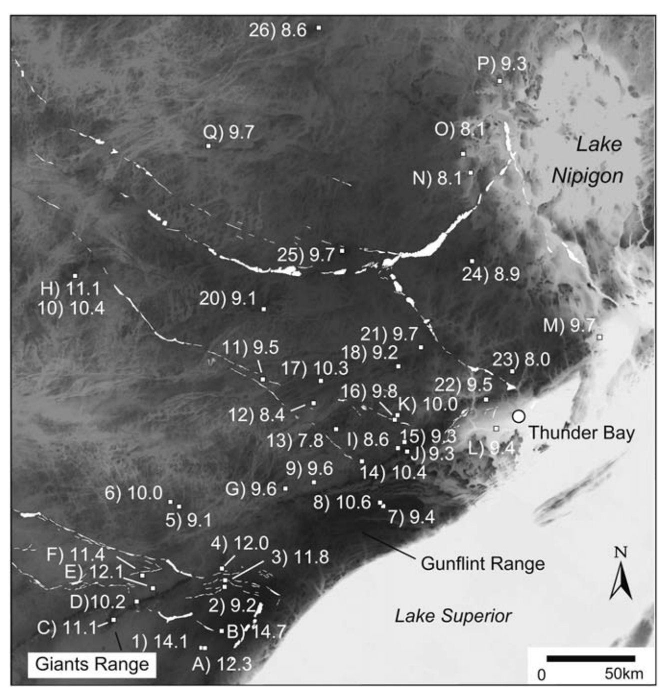

Fig. 3. Significant glacial features and coring site locations. Ages are reported in ka radiocarbon ages. Numbers are keyed to new sites reported here and lettered sites are from prior reports. All details are in the respective portions of Table 1.

individual moraines into three sets where 2–3 of the individual moraines lie in close proximity to each other. There is considerable spacing among the sets. Further, the present sampling resolution does not differentiate among the individual members of any set. The northern moraines have lake basins on their proximal sides and lobate fans locally on their distal sides. The overall lack of drift between moraine sets may point to more rapid deglaciation across the intervening topography. Overall these moraines are taken to represent times when the glacier margin was stationary for some interval of time or of significantly increased sediment production (cf. Sharpe and Cowan, 1990).

4.2. Southern moraines

The southern moraines trace across the northern end of the Giants Range, MN (Fig. 2). South of the moraine belt is strongly lineated, or streamlined, topography. A different number of moraines lie on either side of the Giants range (Fig. 2). To the east along our transect the moraines are named from south to north, the Allen, Wampus, and Outer and Inner Isabelle Moraines (Lehr and Hobbs, 1992). Generally these are thought to be gravel features and commonly esker ridges lie perpendicular to and feed the moraine ridges. The most pronounced and longest moraine is the Vermilion, making up the northern most moraine of this sequence. Proximal to the Vermilion Moraine, drift-poor bedrock is the primary topographic element.

On the lineated topography south of the southern moraine set, Florin and Wright (1969) reported a basal radiocarbon age on moss of 14,690 ± 390 ^{14}C yr BP (W-1763) [16,631–18,751 cal yr BP] from Webber Lake, MN. Our result from nearby Salo Lake at 14,050 ± 100 ^{14}C yr BP (ETH-30180) [16,282–17166 cal yr BP] is similar, indicating that the lineated till surfaces outside the Vermilion Moraine were deglaciated prior to the Webber Lake age of 17.9 ± 0.6 cal ka BP.

Attempts to secure ages inside the various intermediate moraine sets provided ages younger (e.g. (2) Wampus and (3) East Chub Lake, Fig. 3, Table 1) than a site proximal to the moraine set. Interior to the Vermilion Moraine, in a bedrock basin at 1492 cm depth, we recovered a sample yielding an age of 12,000 ± 85 ^{14}C yr BP (ETH-28945) [13,697–14,046 cal yr BP]. Note that Björck (1990) worked a transect west of the Giants Range within the same moraine belts and recovered a similar distribution of ages. The oldest age in Björck's study was from Heikkilla Lake at 12,100 ± 150 ^{14}C yr BP (Lu-2556) [13,639–14,577 cal yr BP]. These results support each other and suggest that the moraines in this southern set formed before 13.9 ± 0.2 cal ka BP and after 17.9 ± 0.6 cal ka BP.

The base of Little Spring Lake core shows a lithologic change, characterized by oscillations in organic carbon, carbonate, and magnetic susceptibility (Fig. 4). The LOI signatures both show a brief increase then decrease tracked inversely by the magnetic susceptibility signature. The lower bracketing age noted above is 12,000 ± 85 ^{14}C yr BP, and the upper bracketing age from a slightly above the oscillation at 1421 cm depth is 10,420 ± 75 ^{14}C yr BP (ETH-30170) [12,065–12,646 cal yr BP]. Provisionally we interpret this oscillation as a climate signal after the formation of the Vermilion Moraine and sometime before 12.3 ± 0.2 cal ka BP.

4.3. Central moraines

The central set of moraines is composed of the Steep Rock, Eagle-Finlayson, and Brule Creek Moraines (Zoltai, 1965). These are linear ridges trending northwest–southeast with the Steep Rock Moraine lying to the southeast of the Eagle-Finlayson Moraine. Zoltai (1965) suggested that the Brule Creek Moraine may be a continuation of the Eagle-Finlayson Moraine, but lack of continuity between the ridges prevents a firm correlation. The Eagle-Finlayson is a narrow, single ridge that may have been overridden (Zoltai, 1965) but the exact evidence for any overriding is not given. Overall these moraines have simpler geometries than the other two moraines' sets.

Exterior to these moraines is the Rattle Lake site (H, 10, Fig. 3). Here Björck (1985) sampled the bulk sediment from 1266–1271 cm and obtained a radiocarbon age of 11,100 ± 110 ^{14}C yr BP (WIS-1327) [12,866–13,199 cal yr BP, 13.0 ± 0.2 cal ka BP]. Our preliminary analysis on individual plant fragments from the same site and stratigraphy is younger: 10,440 ± 75 ^{14}C yr BP (ETH-32339) [12,084–12,662, 12.3 ± 0.1 cal yr BP]. At Bear Cub in the same geomorphic position we obtained an age of 10,550 ± 75 (ETH-28939) ^{14}C yr BP, [12,345–12,801 cal yr BP, 12.6 ± 0.2 cal ka BP], and from just inside the Steep Rock Moraine at Sunbow Lake, a basal age of 10,400 ± 120 ^{14}C yr BP (ETH-31427) [11,951–12,722 cal yr BP, 12.3 ± 0.2 cal ka BP]. These ages have a wide geographical range along the moraine set but have a similar age. Taking the age inside the Steep Rock Moraine as reference we argue that the area between the proximal position of the southern moraines and the older of the central moraines was deglaciated before 12.3 ± 0.2 cal ka BP. This is broadly in agreement with the lithologic oscillation in the Little Spring Lake core (Fig. 4).

At a location proximal to this moraine set, organic remains from the base of Crawfish Lake yielded an age of 10,250 ± 75 ^{14}C yr BP (ETH-31430) [11,706–12,248 cal yr BP, 12.1 ± 0.1 cal ka BP]. This implies that the entire moraine set formed prior to 12.1 ± 0.1 cal ka. The near overlap in ages from both the proximal and distal positions of this moraine set may indicate these moraines formed in a brief interval of time (cf. Sharpe and Cowan, 1990). This set lies across the postulated Lake Agassiz meltwater discharge route. Specifically segments of the Brule Moraine cross directly over the drainage divide (Fig. 5). Since this moraine is not breached at the lowest point in the divide, it is unlikely any drainage could have occurred after its formation or else portions of it would have been eroded.

4.4. Northern moraines

The northern set of moraines includes the Hartman, Dog Lake, Lac Seul and Kaiashk (Fig. 2). The Hartman and Dog Lake Moraines are the southern most of this set and likely record a contemporaneous ice margin as indicated by an interlobate, triangular area of kettled drift narrowing to the northeast (labeled Fig. 6) that lies between the two morains. Zoltai (1965) noted that striations north of the Hartman Moraine are aligned slightly west of due south, whereas those interior to the Dog Lake Moraine have a distinct

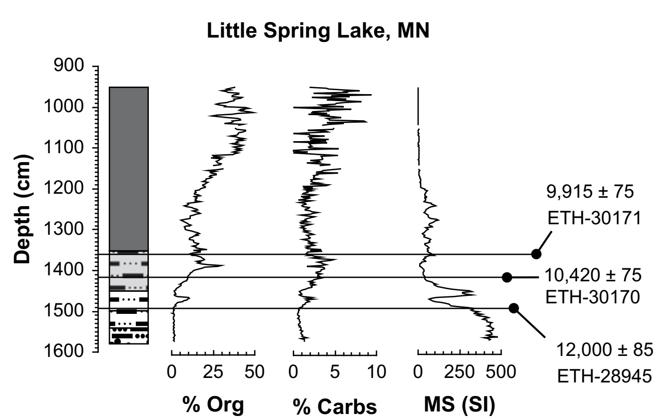

Fig. 4. Stratigraphy of the Little Spring Lake site, including lithology, LOI, MS and associated ages. The stratigraphy for the other sites is in the Supplemental material.

Fig. 5. Oblique image of the Brule Moraine separating Watershed Lake (foreground) and Squeers Lakes (background). These lakes are now connected and drain within the Hudson Bay watershed, the continental divide is near the center of this image. However the presence of this moraine at the topographic divided indicates meltwater flow did not breech it. View to the southeast.

southwest alignment. Since field mapping shows these flow patterns are consistent for more than 200 km upstream from these locations (Zoltai, 1965, 1967), we interpret these observations to represent glacier flow sourced from two different sectors of the ice sheet, converging to the southwest, and forming the interlobate moraine between them.

Interior to these moraines are the Lac Seul and Kaiashk Moraines. The thicker drift comprising the Lac Seul Moraine can be traced directly across the inner limit of the triangular interlobate moraine associated with recession of the ice from the Hartman and Dog Lake Moraines noted above (Fig. 6). The Lac Seul and Kaiashk Moraines form an arcuate pattern extending eastward almost to Lake Nipigon. As noted by Zoltai (1965) and confirmed by our observations, the northern side of this moraine has deep kettles in it and has gullies dissecting its southern side. The relative abundance of kettles increases to the northeast. From these observations we interpret a northern ice-contact slope with sediments and meltwater shed to the southeast. In this view, the entire Kaiashk Moraine is not an interlobate feature, but rather an eastward extension of the Lac Seul Moraine with a much thicker sediment sequence on its eastern side. That the eastern portions of the Kaiashk Moraine run nearly perpendicular to the Dog Lake Moraine implies that the ice building out to the Kaiashk Moraine position traversed the area formerly occupied by ice flowing to the southwest to build the Dog Lake Moraine. Locally, these cross-cutting ice-flow directions are best explained by a readvance of unknown extent.

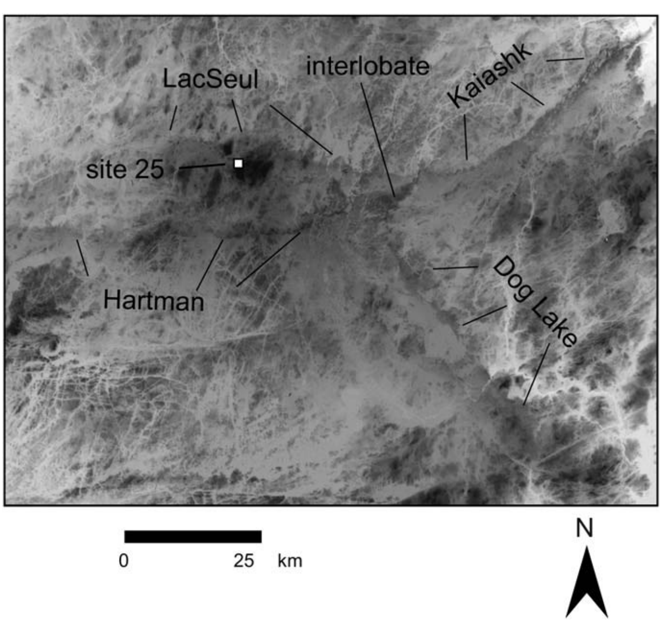

Fig. 6. Detailed view of the Hartman/Dog Lake and Lac Seul/Kaiashk Moraines. This DEM image illustrates the difference in character of the Hartman/Dog Lake Moraine resulting from different lobes but along a contemporaneous ice margin. A triangular interlobate area is labeled, The Lac Seul/Kaiashk Moraine lies inside this interlobate features and is one margin formed at a later time.

Pretty Lake (site 25, Fig. 4) lies between the Hartman/Dog Lake Moraines and Lac Seul/Kaiashk Moraines, and the basal date from that core is 9705 ± 145 ^{14}C yr BP (ETH-31001) [10,642–11,408 cal yr BP] from just above the basal silt. From geomorphic and stratigraphic relationships, the highest probability calibrated age 11.2 ± 0.2 cal ka BP from Pretty Lake requires that the Hartman/Dog Lake Moraine is older than that. Since the Lac Seul/Kaiashk Moraine must be younger based on the same arguments, it might be suggested that Hartman/Dog Lake Moraines formed during or at the end of the Younger Dryas (12.9–11.7 cal ka BP), whereas the Lac Seul/Kaiashk Moraine formed at its end. The Lac Seul/Kaiashk then may be associated with the Pre Boreal oscillation, which occurred at 11.15–11.3 cal ka BP (Björck et al., 1997).

4.5. Time–distance diagram

For the results discussed above only the individual ages providing the closest brackets on deglaciation for each moraine set were presented. In Fig. 7 the entire data set (as the probability distribution of the calibrated age) is shown by projecting the core sites to a transect line approximately perpendicular to the ice margins (line in Fig. 2). Note that the Dog Lake Moraine could occupy several positions along the transect because its orientation differs from the other moraines, we have plotted where it joins the Lac Seul Moraine. The minimum ages from Teller et al. (2005) are also plotted in Fig. 7. Taken together, these ages form a time-transgressive envelope that deglaciation cannot be younger than. The presence of organic materials implies that they existed at that location and their stratigraphic position in the core requires that the site be deglaciated. For any location these data yield a range of ages. For example, at the transect position of the Brule Moraine the individual ages in that geomorphic position range from about 12.3 to 8.5 cal ka.

As a first approximation the slope of the line defining that envelope represents the retreat rate of the ice margin assuming a simple retreat. The line is placed where the probability, with the available data, drops to zero. Any readvance would reduce the time and thus the retreat rates would be higher than noted here. Thus, from outside to just inside the southern moraine set, ice margin recession rates in this construction are ~13 m/a. Between the southern and the middle moraine sets the rate increased to ~65 m/a, and between the middle and the northern moraine sets the rate more than doubled to ~135 m/a. North of the northern moraine set, the rate increased further to ~161 m/a. Precisely how closely these rates track the exact deglaciation along the transect is discussed in Section 5.1 below and the implications are noted in Section 5.3.

5. Discussion

5.1. Interpretation of basal radiocarbon ages

This report assigns provisional ages to three different moraine sets formed as the Laurentide Ice Sheet retreated west of Lake

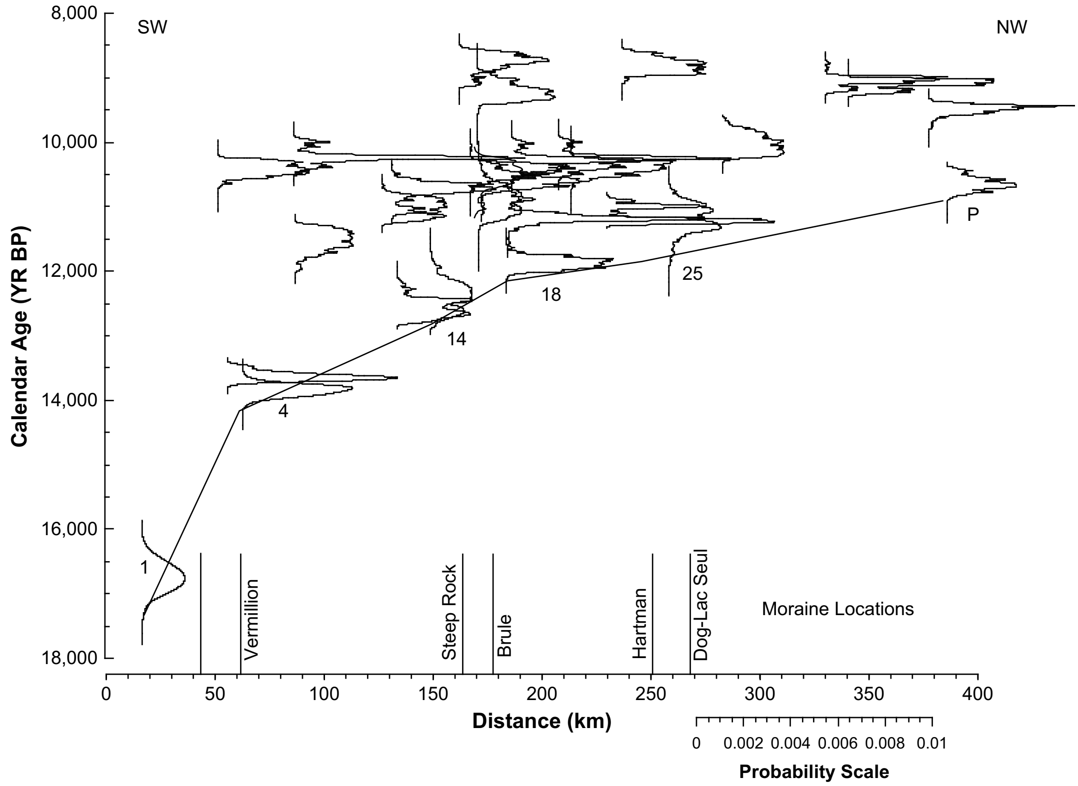

Fig. 7. Time–distance diagram with all calibrated radiocarbon ages shown. The oldest ages from these sites form an envelope and the deglaciation cannot be younger than that envelope. The lines defining the base of the envelope are used to derive the minimum retreat estimates discussed in the text. Labels' curves correspond to site numbers in Table 1.

Superior. Below we discuss caveats of those age assignments and then discuss the implications of the assignments. First, as noted above all basal dates do not directly track deglaciation but provide a bracket on any individual site. Some would argue that they are of little value. However, we note by making multiple assessments for any individual landform, the brackets get tighter. With one or two ages for a given geomorphic position this is difficult to assess, but with more ages the form of an envelope emerges. The number needed to define a good fit remains subjective. In the context of multiple moraine sets, however, additional constraints from relative positions, results in a given age to be both a minimum to the moraine exterior (distal) to it and a maximum to the moraine interior (proximal) to it. Around Thunder Bay the moraines are successively younger to the north, thus older ages to the north trump younger ages to the south. Here we sampled 25 new sites and added the information from 18 more reported by other workers, but only five of them are employed in developing a chronology for ice recession. Is it possible that additional coring would yield older ages? Yes, but the absolute number needed to be sure that no older sites exist is not yet possible to assess. In any event, this problem is not unique to this data set, but exists in all reports of deglaciation sequences that employ basal ages, which includes Dyke (2004). Non-linearities in the radiocarbon time scale also impact any age assignments. These changes are generally a few hundred years and given the other considerations such as the minimum problem discussed below, this is not taken to be the largest source of uncertainty in this work. The offset for plants to reach the sites may not be long. It is clear that vegetation can reoccupy freshly deglaciated landscape within decades (Mathews, 1992; Jones and Henry, 2003). We suspect other factors relating to lake formation are responsible for the range of ages at a particular site.

It is possible that the distribution of basal dates partly reflects climate conditions. It could be argued that conditions favorable to organic growth and preservation increase the number of basal ages on a landscape deglaciated prior to that time. To illustrate this possibility, the probabilities of the radiocarbon dates generated in this study are summed and combined. The resulting plot (Fig. 8) shows the most probable distribution for basal radiocarbon ages across the entire study area. For example the highest probability of a radiocarbon age is about 10,200 cal yr BP, the time immediately following the study area's deglaciation so the cause is not deglaciation, but rather some other regional impact that extends across the entire study area as basal ages about 10.2 cal ka can be found near all moraine set.

A comparison to the oxygen isotope record from the GISP2 ice core shows phasing with the probability of the basal ages. There is a higher likelihood of basal ages in the warm interval after the Younger Dryas with the largest peak after the Pre Boreal oscillation. There is relative low probability in the second half of the Younger Dryas and a relative high zone in the warm interval prior to the Younger Dryas. We do not argue for a direct, one-to-one correspondence between these two because of the limited number of dates in the older part of the sequence. This in itself is telling. We suggest further consideration of the temporal distribution of a large numbers of basal radiocarbon ages may yield insights on why these basins formed at preferential times. To the extent some of the basal ages cluster around climate events, it might be argued that removing those from consideration leaves ages more representative of other processes such as deglaciation.

Turning now to that subset of basal ages we use to constrain the deglaciation history. The southern moraine set has a minimum age that is ~500 yr after the switch to the warmth of the Bölling–Alleröd oscillation. Likewise the middle moraine set minimum ages

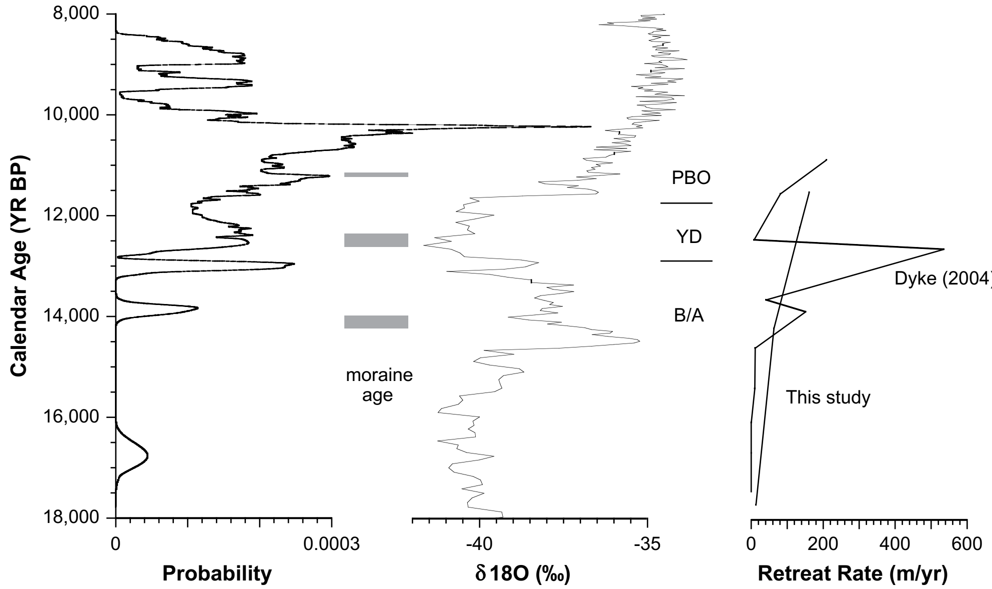

Fig. 8. Probability sum of all radiocarbon ages, the suggested ages for the three moraine sets discussed here, the GISP2 ice core record, and the estimated retreat rates are derived from this work and from Dyke (2004). Abbreviations are PBO – PreBoreal oscillation, YD – Younger Dryas, and B/A – Bölling–Alleröd. The ages for the boundaries are taken from Lowe et al. (2008) but the dating resolution makes further subdivision of the B/A unwarranted here.

are ~500 yr after the warm interval prior to the start of the Younger Dryas. The northern moraines' sets are roughly 500 yr after the end of the Younger Dryas.

We take the northern moraine set as the key to interpreting all three moraine sequences. There are no well-defined moraines for a considerable distance north of them. This is not surprising given that the continued warmth of the early Holocene was probably not conducive to moraine formation. Lowell et al. (1999) suggested that for ice sheet margins, which are unlikely to ever be in true equilibrium, the moraines form at the transition from cold to warm periods. The margins expand during the cold interval when reduced ablation at the margin permits ice sheet advance. An advance would continue as long as cold conditions are present, however at the transition to warm periods, with increased summer ablation, the advance would be halted and the ice margin becomes quasi-stable. Any warming response is more immediate; overwhelming any accumulation signal that may take considerable time to be expressed at the ice sheet margin. A modern example can be seen in Greenland where extensive Little Ice Age moraines ring both the terrestrial portions of the ice sheet and independent land-based alpine glaciers on average some 2 km from the ice margin (Weidick, 1968). Given the possibility that the Little Ice Age represents the most extensive glacier expansion in Greenland since the lateglacial (Kelly et al., 2008), the Little Ice Age moraines represent the start of the warming with recession during the current warming trend not producing significant moraines. If accumulation was the dominant signal since the Little Ice Age, it is unlikely the ice sheet and smaller independent glaciers would show a similar retreat pattern.

We argue that the ages of the moraines in the Thunder Bay region record the same processes. The similar offset of 500 yr in these three cases may be fortuitous, however some delay must exist from the time a climate transition takes place until the organic materials accumulate in depressions. The magnitude of this effect must be assessed at each study site. Here, this proposed sequence can be tested with a parallel transect using exposure-age dating techniques applied to the same sets of moraines. If the above arguments are correct, it implies a close coupling between the ice sheet and the climate oscillations albeit offset by ~500 yr from the actual climate change. Note that the ages assigned to the moraines here do not represent the time of maximum cold, but instead, the switch from cold to warmth. In this context the moraines are interpreted as the change in behavior of the ice sheet during late-glacial climate oscillations.

5.2. Implications for meltwater drainage

Broecker et al. (1989) suggested that portions of the study area were the conduits for meltwater from glacial Lake Agassiz to the east that altered ocean circulation patterns in the Atlantic basin that thus triggered the Younger Dryas cold event. This comes from the timing of isotopic changes in the Gulf of Mexico and paleogeographic reconstructions. This hypothesis is based partly on extrapolated water planes from rebound reconstructions suggested by Elson (1967) and later modified by Teller and Thorleifson (1983). This is at odds with Zoltai (1965), who based on field mapping in the area where the drainage pathway was suggested, reached the conclusion that no eastward drainage occurred. We note that the middle set of moraines trace directly across this route and their local geometry directly blocks the lowest topography (Fig. 5). In a simple, retreat only model, our suggested minimum age for this moraine is 12.1 ± 0.1 cal ka BP and implies that the ice margin was south of the topographic divide midway through the Younger Dryas event. This interpretation implies that the ice sheet margin would have blocked any eastward drainage during at least the first half of Younger Dryas time.

Although more complex models than simple retreat can be formulated, until supporting evidence for them come to light, it may be fruitful to explore some implications of the above for the history of Lake Agassiz. Lake level indicators at topographic elevations below the southern outlet of the lake require a low phase. Johnson (1916) first reported a disconformity in the lake sequence and that break is now called the Moorehead low. In contrast to earlier works who assigned the timing of this low phase based on one age of 10,960 ± 300 (W-723) [12,060–13,448 cal yr BP] (Usgs, 1960), more recent work (Fisher and Lowell, 2006; Fisher et al., 2008) indicates subareal exposure from 12.6 to 11.4 ka cal BP. Whatever its age or cause the operating assumption has been if the water did not flow out the southern outlet, it must have followed the ice margin and exited the Agassiz basin across the topographic divide either into the Atlantic or Arctic basin. Subsequent rebound would cut off these routes allowing the water levels to rise. The conclusions of this work on the eastern route combined with those from the northwest route (Fisher et al., 2009) do not allow this.

Rather than debate methodology possibilities, which are noted above, let us here examine the assumption that the lowering and subsequent refilling of the basin stem from changing paleogeography only. Two alternative possibles come to mind. First, it may be that water in the Agassiz basin developed sufficient depth to leak northward under the ice sheet by the processes outlined in Clarke et al. (2004). One challenge to this possibility is rather the ice sheet had thinned sufficiently at the time of the Moorehead low to allow this leakage. However, glaciology identifies possible routes that would drain the basin. The second possibility is hydrological based. For sometime, it was assumed that major changes in levels of the postglacial Great Lakes were also controlled by rebound control of the topography. However, the recent work summarized in Lewis et al. (2008) points out this does not explain all recent evidence. Rather they examine the hydrological mass balance and suggest climate conditions can generate the observed closed basins. Following that logic we wonder rather the same argument might explain the Moorehead low. Since the transgression occurred at the end of the Younger Dryas it may be that reduced meltwater fluxes during Younger Dryas time could be offset by evaporation losses both from the free lake surface and by sublimation from the frozen lake surface. Arid air masses penetrating to the Lake Agassiz basin would facilitate these losses. Modeling experiments might be undertaken to define the range of conditions (e.g. basin size, meltwater production, evaporation losses) needed to allow this hypothesis. Unfortunately actual values, derived from proxy indicators that might provide a definitive answered, are still being determined.

5.3. *Implications of retreat patterns*

Next we explore the implications of ice margin retreat rates estimated here. To start we contrast those derived here with those generated from the deglaciation chronology of Dyke (2004). For this exercise the radiocarbon ages in Dyke (2004) are converted to calendar years (Fig. 8). In the older portion of the record the rates from this study are higher than from Dyke (2004) in part because Dyke (2004) suggests the ice margin remained at, or near the Vermilion Moraine for nearly 2500 yr. Just before 14.0 cal ka BP the retreat rates from this study increase, whereas those derived from Dyke (2004) increase also, but contain variable rates peaking as high as 500 m/a at ~12.7 cal ka BP. Subsequently the rate drops off at ~12.5 cal ka BP and then increases again ~12.0 cal ka BP. It should be noted that since rate is a derivative, differences between these two models are amplified. Some of the differences may result from the higher dating resolution reported here, but also from how control points are chosen. Dyke (2004) drew isochrones independent of whether the ice margin was advancing, stationary, or retreating; whereas we assigned ages to moraines that are assumed to represent quasi-stationary positions of the ice margin between periods of advance or retreat of the ice sheet. In any case, the rates show far less variation than those implied by Dyke (2004).

Comparison can also be made with retreat chronologies reported from other sectors of the ice sheet (Fig. 9). In New England, Ridge et al. (1999) employed varve techniques to track the ice sheet margin and found nearly constant retreat rates of 155 m/a except for the Littleton readvance at ~13.8 cal ka BP. Along the southern portion of the Des Moines Lobe, Lepper et al. (2007) used radiocarbon ages from maximum stratigraphic positions and basal organics verified with OSL techniques to propose rates from 105 to ~270 m/a. In the Fort McMurray, Alberta, Fisher et al. (2009) found rates from 23 to nearly 300 m/a. Since these are the simplest retreat patterns, not including any readvances, except in the New England study, they are minimum retreat values.

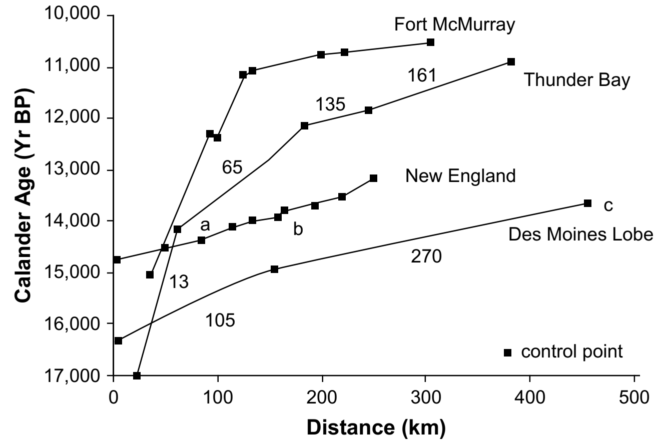

Fig. 9. Comparison of retreat patterns from this study with those reported in New England (Ridge et al., 1999), in the Des Moines Lobe (Lepper et al., 2007) and the Fort McMurray area of northern Alberta (Fisher et al., 2009). Plotted are the time–distance diagrams for each area, thus the retreat rates are the slopes of the respective line segments. They are minimum values as readvances are not accommodated here. Selected slopes have the retreat rate (m/a) identified for comparison. Lower case letters identify sites noted in the text a) middle moraines (this study), b) Littleton readvance (Ridge et al., 1999), and c) Big Stone Moraine (Lepper et al., 2007).

Ideally we would like to invert these geomorphic observations to learn more about the mass balance conditions for these sectors of the ice sheet. Such conditions will quantify meltwater contributions to the oceans. Recently Oerlemans (2005) demonstrated that a temperature signal can be extracted from glacier length. Using historical length records from 169 glaciers, glaciology theory, and empirical calibration to modern mass balance measurements, he estimated temperature warming of 0.6 °C since 1900 A.D., the same as from many other proxy records. These glacier derived temperature estimates stem from knowing the climate sensitivity and response time, which depend upon only three variables: glacier length, the annual precipitation, and mean surface slope of the glacier. Climate sensitivity is defined as the total length change from one equilibrium configuration to another for a specified temperature change (Oerlemans, 2005). For the Laurentide Ice Sheet case, generating reliable values for precipitation and surface slope to infer temperature are premature. However, temperature, precipitation, glacier length and surface slope are likely the key variables to understanding the retreat patterns recorded by the ice sheet in terrestrial settings.

Returning to the examples of retreat rate information (Fig. 9), we note that despite coming from widely different geographical areas and settings, similarities exist. First, we note that the southern moraine complex (>13.9 cal ka BP) formed about the same time as the Big Stone Moraine from the Des Moines lobe (>13.5 cal ka BP) and the Littleton Moraine (~13.8 cal ka BP) in New England. A tentative correlation can also be made between the middle moraine complex (>12.3 cal ka BP) and Don's Moraine (>12.7 cal ka BP) in Fort McMurray, Alberta and between the northern moraine complex (>11.2 cal ka BP) and Survive Moraine (>11.6 cal ka BP) in Fort McMurray, Alberta (Fisher et al., 2009). While dating improvements could verify these correlations, taken together these temporal correspondences suggest that moraine formation represents something other than just local factors. Moreover, the maximum terrestrial retreat rate at each of these four sites is similar and ranges from ~150 to 250 m/a. The younger portion of the Fort McMurray transect was in contact with a significant glacial lake (Fisher et al., 2009) and should reflect higher retreat rates. With the present resolution of deglaciation retreat rates, it appears that rates are independent of the glacier bed type, whether hard or soft.

During the last century typical retreat rates of alpine glaciers, over a wide range of precipitation and glacier length, are $\sim$10 m/a. A simple explanation for the higher retreat rates noted above is that faster rates result from lower surface slopes of the ice sheet. This may well be, but it is curious that over areas that would be described as hard (New England, Thunder Bay) and those that would be called soft (Des Moines, Fort McMurray) the retreat rates are similar. Typically reconstructions over similar bedrock areas would call for surface profiles over hard beds to be nearly an order of magnitude higher than over soft beds (e.g. Clark, 1992). If the Oerlemans (2005) empirical formula applies, the climate sensitivity should scale in direct ratio to the surface slopes. For example, for the surface slopes from two different ice masses that have a ratio of 7:1 but the same precipitation of 3 m/a, the climate sensitivity ratio is 7.02:1. In order to observe similar retreat rates from different sectors of the ice sheet, then, requires either different amounts of temperature forcing, very different response times for these sectors of the ice sheet, or that the surface slopes were similar over different types of beds. Given that these moraines appear to form at similar times, which requires a switch from retreating to advancing or stationary ice margins, vastly different temperature forcings or internal ice dynamics may be the least likely explanation. Future work should hone these terrestrial retreat estimates, perhaps adding news ones covering different time slices, and conduct modeling experiments to assess the reasons that retreat rates are similar in different settings.

6. Conclusions

Based on radiocarbon control, when retreat of the Laurentide Ice Sheet took place during the lateglacial climate swings, the retreat rates appear to be similar and mimic the swings in climate. In the present study three major moraine sets trace across the Thunder Bay area. The largest and most complex moraine sets are at the southern and northern portions of the study area, and are assigned minimum ages of formation of 13.9 ± 0.2 and 11.2 ± 0.2 cal ka BP, respectively. These ages coincide with the Bölling–Alleröd warming and the Pre Boreal warmth at the end of the Younger Dryas but are offset some 500 a. The smaller, simpler moraines in the middle of the study area have a minimum age assignment from 12.3 ± 0.2 to 12.1 ± 0.1 cal ka BP. The simplest, minimum retreat rates range from 63 m/a from between the southern and middle moraine sets, to 161 m/a to the northern moraine sets. These are baseline values that define the rate at which the Laurentide Ice Sheet retreated when subjected to the climate swings that occurred during the last termination. When contrasted to retreat rates from other sectors of the ice sheet these rates appear similar even though the subglacial bed conditions differ. The age assignments made to moraines in this study do not permit eastern meltwater drainage from Lake Agassiz to the Superior basin until after the Younger Dryas event.

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Appendix. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi: 10.1016/j.quascirev.2009.02.025.

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