Glacial Lake Agassiz: A 5000 yr history of change and its relationship to the $\delta^{18}O$ record of Greenland

James T. Teller†
Department of Geological Sciences, University of Manitoba, Winnipeg, Manitoba R3T 2N2, Canada

David W. Leverington‡
Center for Earth and Planetary Studies, National Air and Space Museum, Smithsonian Institution, Washington, D.C. 20560-0315, USA

ABSTRACT

Lake Agassiz was the largest lake in North America during the last period of deglaciation; the lake extended over a total of 1.5×10^6 km^2 before it drained at ca. 7.7 ^{14}C ka (8.4 cal. [calendar] ka). New computer reconstructions—controlled by beaches, isostatic rebound data, the margin of the Laurentide Ice Sheet, outlet elevations, and a digital elevation model (DEM) of modern topographic data—show how variable the size and depth of this lake were during its 4000 ^{14}C yr (5000 cal. yr) history. Abrupt reductions in lake level, ranging from 8 to 110 m, occurred on at least 18 occasions when new outlets were opened, reducing the extent of the lake and sending large outbursts of water to the oceans. Three of the largest outbursts correlate closely in time with the start of large $\delta^{18}O$ excursions in the isotopic records of the Greenland ice cap, suggesting that those freshwaters may have had an impact on thermohaline circulation and, in turn, on climate.

Keywords: Lake Agassiz, history, outbursts, Greenland isotopic record.

INTRODUCTION

During the last period of global deglaciation, large lakes formed along the margins of continental ice sheets in North America, Europe, and Asia. The complex history of North American proglacial lakes has been summarized by Prest (1970), Teller (1987, 2004), Dyke and Prest (1987), Klassen (1989), Dredge and Cowan (1989), Karrow and Occhietti (1989), Lewis et al. (1994), and others. Several special volumes of papers (e.g., Teller and Clayton, 1983; Karrow and Calkin, 1985; Teller and Kehew, 1994) and hundreds of journal articles over the past century have provided important insight into this extensive lake system. Closely linked to the history of proglacial lakes is the chronology of the routing of North American glacial runoff, which has been discussed by many for various regions, and synthesized by several, including Teller (1987, 1990a, 1990b, 1995) and Licciardi et al. (1999), and tied to past global ocean circulation by Broecker et al. (1989), Clark et al. (2001), Teller et al. (2002), and others.

The largest of all proglacial lakes was Lake Agassiz, which developed and expanded northward along the margin of the Laurentide Ice Sheet as it retreated downslope into the Hudson Bay and Arctic Ocean basins in Saskatchewan, Manitoba, Ontario, Quebec, and Northwest Territories. During the history of Lake Agassiz, overflow was carried from the lake to the oceans by way of four different routes (Fig. 1): (A) southward via the Minnesota and Mississippi River Valleys to the Gulf of Mexico, (B) northwestward through the Clearwater-Athabasca-Mackenzie River Valleys to the Arctic Ocean, (C) eastward through a series of channels that led to the Great Lakes (or to glacial Lake Ojibway) and then to the St. Lawrence Valley and North Atlantic Ocean, and, finally, when Lake Agassiz completely drained, (D) northward and eastward through Hudson Bay and Hudson Strait to the North Atlantic Ocean, between the Keewatin and Labrador ice centers.

The extent, configuration, and depth of Lake Agassiz, as with all proglacial lakes, was a result of the interaction among (1) location of the ice margin, (2) topography of the newly deglaciated surface, (3) elevation of the active outlet, and (4) differential isostatic rebound (Teller, 1987, 2001). This interaction was complex, partly because it involved glacier melting and ice-dam failure, as well as ice readvances and surges into the lake basin. Thus, overflow outlets were opened and occasionally closed again by a fluctuating retreat. Outlet erosion also played a role. The interrelationship between differential isostatic rebound and the geographic location of the outlet was especially important in dictating the extent and depth of Lake Agassiz. As described by Teller (2001), whenever the outlet carrying overflow was at the southern end of the basin, lake levels receded everywhere to the north of the isobase (i.e., the contour of equal isostatic rebound) through the outlet (Fig. 2A). Whenever the outlet was not at the southern end of the basin, transgression occurred to the south of the outlet, while regression occurred everywhere to the north of it (Figs. 2B and 2C). Isobases across the Agassiz-Ojibway basin are shown in Figure 3, along with the locations of the main outlets.

As a result of the abrupt opening and closing of outlets, plus the interaction of differential isostatic rebound and the use of many different outlet channels, the level of Lake Agassiz was constantly changing, and its history is very complex. Overall, its history is one of northward expansion, punctuated by abrupt drops in lake level when lower outlet channels were deglaciated; each decline was then followed by transgressive deepening of the lake in the basin south of the isobase through the outlet and regression north of that isobase due to differential rebound.

In this paper, a series of maps are presented

†E-mail: tellerjt@ms.umanitoba.ca.
‡E-mail: leveringtond@nasm.si.edu.

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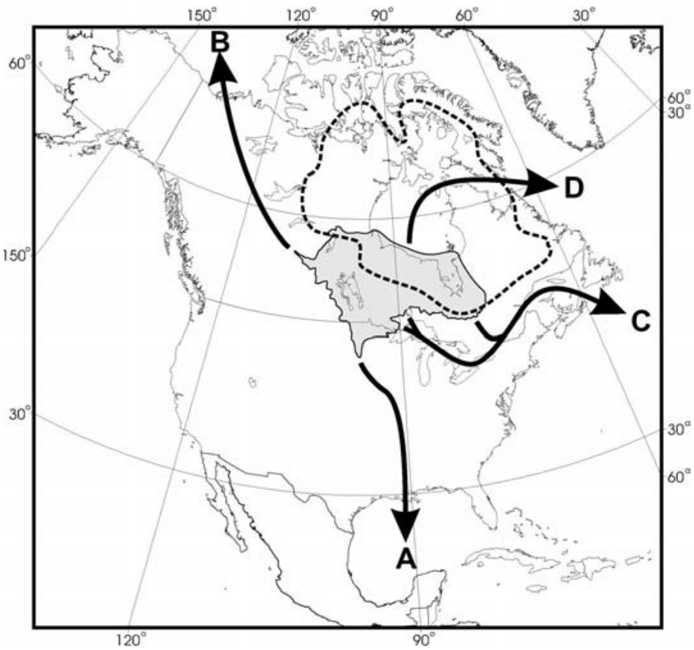


Figure 1. Routing of runoff to the oceans from Lake Agassiz (Teller et al., 2002). Ice margin at ca. 9 ^{14}C ka shown by dashed line. Outflow paths A–D described in text.

that outline the extent of Lake Agassiz at 18 different transgressive maximums and 15 intervening minimums. These span the history of the lake from ca. 10.9 ^{14}C ka (thousand radiocarbon years B.P.; 12.94 thousand calendar years B.P.), until its final stage at ca. 7.7 ^{14}C ka (8.45 cal. ka), after it had amalgamated with glacial Lake Ojibway. In addition, we discuss some of the relationships of these outbursts to the δ^{18}O curves in the Greenland GISP2 and GRIP ice cores.

LAKE RECONSTRUCTION

Maps depicting various stages in the history of Lake Agassiz have been published for more than a century, beginning with those in U.S. Geological Survey Monograph 25 by Upham (1895). Johnston's (1946) field work on the beaches provided important new data on shorelines in the Canadian part of the lake, north of where Upham had worked. Modern reconstructions have been guided by the mapped and correlated strandlines of Upham (1895) and Johnston (1946) and have been supplemented by new topographic map data, field work, and aerial photograph interpretation (see Elson, 1983). More recent reconstructions of Lake Agassiz have extended the area of the lake farther north (Teller et al., 1983; Elson, 1983; cf. Smith and Fisher, 1993), and, because glacial Lake Ojibway amalgamated with Lake Agassiz during the last several hundred years of its history, this eastern extension, as described by Vincent and Hardy (1979) and Veillette (1994), is included in our reconstructions. Elson (1967), Prest (1970), and Dyke and Prest (1987) provided a number of snapshots of the lake through its history, and some, such as Clayton and Moran (1982), Teller (1985), Klassen (1989), Dredge and Cowan (1989), and Thorleifson (1996), presented maps and integrated its history with deglaciation across central North America. Lake reconstructions in various regions and at various times were also published by researchers in the volume *Glacial Lake Agassiz* (Teller and Clayton, 1983), among them Brophy and Bluemle (1983), Dredge (1983), Klassen (1983a, 1983b), Fenton et al. (1983), and Schreiner (1983). Smith and Fisher (1993) and Fisher and Souch (1998) expanded the lake northwestward to the Clearwater-Mackenzie outlet.

Reconstructions in this paper expand on those in Leverington et al. (2000, 2002a), which used, and projected northward, the correlated beach elevations in Thorleifson (1996), which were derived from Johnston (1946) and Teller and Thorleifson (1983) and were integrated with isobase strandline data to the east from Vincent and Hardy (1979) (Fig. 3A). In contrast to Figure 3A, which shows the trend of isobases at 100 km spacing, Figure 3B presents the trend of isobases at a *vertical* spacing of 25 m on the Upper Campbell water plane. Each Lake Agassiz beach defines its own rebound curve, because isostatic rebound varied through time; older beaches have slightly steeper curves, whereas younger ones have gentler curves (see Teller and Thorleifson, 1983, Fig. 2). The *trend* of the isobases over the life of the lake are assumed to be the same, although we recognize that changes in relative thickness of the Laurentide Ice Sheet during deglaciation are likely to have led to some changes in isobase orientation and configuration.

The isostatically deformed rebound curves were used to generate a paleo–water surface of Lake Agassiz associated with each beach (Mann et al., 1999; Leverington et al., 2002b). Each rebound surface was computationally projected to intersect modern topography (defined by a digital elevation model [DEM] with individual grid dimensions of $\sim 1 \times 1$ km), generating an outline of the lake south of the Laurentide Ice Sheet (Leverington et al., 2000, 2002a, 2002b). The GLOBE (Globe Task Team, 1999) and ETOPO 5 (NGDC, 1988) databases were used as the sources for the modern DEM. Because control points for isostatic rebound are nearly all at the margin of the lake, the simple curvilinear isobases representing this rebound (Fig. 3) may be more complex, as suggested by Dredge and Cowan (1989) and Rayburn and Teller (1999). However, lake bathymetry and total lake volume would have been influenced only slightly by such isostatic variability, and the coincidence of actual beaches and wave-trimmed cliffs with the topographically modeled shorelines indicates that the outlines of various lake stages are accurately depicted, except perhaps in far northern regions where rebound curves were projected and are not controlled by strandline data.

The ice margin across the Lake Agassiz basin through time is controlled in only a few places, mainly by scattered and dated end moraines and by some dated lithostratigraphic relationships. Additional ''local'' control for placement of the ice margin is based on the linkage of lake levels (beach elevations) to specific outlets carrying overflow at a given time, which were controlled by the ice margin. Because of this uncertainty, and because the Laurentide Ice Sheet margin was very dynamic during retreat—repeatedly surging into the lake and rapidly calving back (e.g., Clayton et al., 1985; Dredge and Cowan, 1989)—we have kept the ice margins shown in Figure 4 relatively constant through time across that part of the basin where there is no control; we acknowledge that the ice margin fluctuated, extending beyond our ''average'' margin at times and retreating north of it at other times,

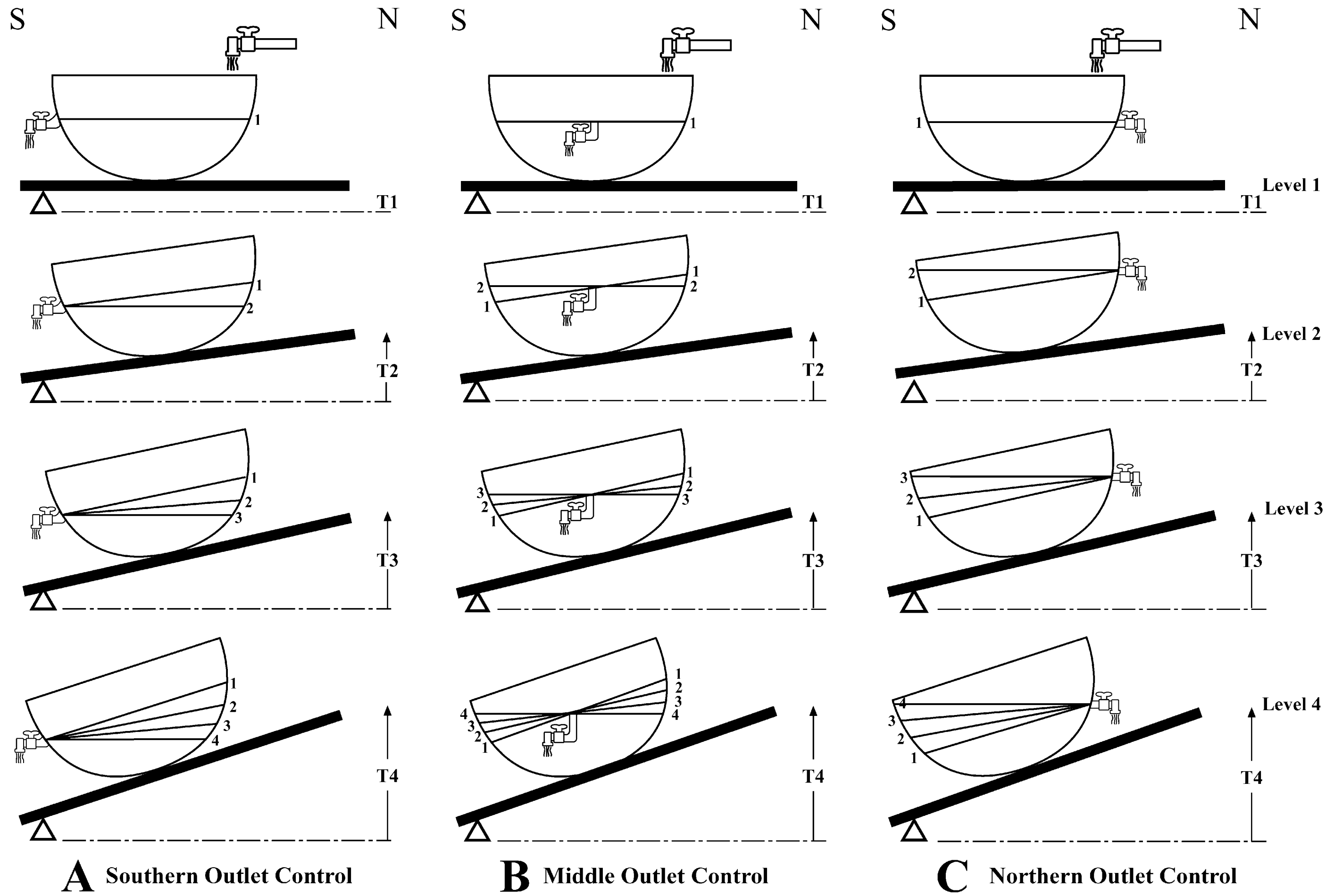


Figure 2. Cartoons showing the influence of differential isostatic rebound through time (T1, T2, T3, T4) on lake level, resulting from overflow through (A) an outlet in the southern end of the basin, (B) an outlet between the northern and southern ends of the basin, and (C) an outlet in the northern end of the basin (after Larsen, 1987; Teller, 2001). Through time, older lake levels (beaches) become elevated or submerged with respect to the contemporaneous (horizontal) level.

but large, deep lake basins are not good places for the evidence of short-lived marginal positions to be preserved.

The level of Lake Agassiz fluctuated considerably throughout its history, with numerous transgressive maximums identified by mapped (and named) beaches and intervening low-level stages controlled by the elevation of newly opened outlets. Subsequent uplift of those outlets led to deepening of the lake before the next, lower outlet was deglaciated; these intervening minimum levels did not form beaches and their outlines are controlled by the elevation of the spillover points in our paleotopographic database. The mapped minimums were implicit in previous volumetric calculations (Mann et al., 1999; Leverington et al., 2000, 2002a; Teller et al., 2002), but have never been published.

Although substantial volumes of water were released between the transgressive maximums and subsequent drawdown levels (see drawdown values in caption to Fig. 4), the lake did not always change substantially in areal extent. In Figure 4, we show a selection of paired maximums and minimums (e.g., A1-A2, B1-B2); six additional pairs showing only relatively small change in area between preceding and subsequent lake outlines are also available.[1] None of the *minimum* lake outlines in Figure 4 has been published before; six of the 12 maps in Figure 4 that show the transgressive *maximums* have been published previously as bathymetric maps by Leverington et al. (2000, 2002a), although two have been substantially modified.

[1]GSA Data Repository item 2004079, Figure DR1, is available on the Web at http://www.geosociety.org/pubs/ft2004.htm. Requests may also be sent to editing@geosociety.org.

HISTORY OF THE RISES AND FALLS OF LAKE AGASSIZ

Dating of Events

There are many beaches and wave-cut scarps in the Lake Agassiz basin. Some are large, distinct, and continuous for many kilometers; others are not (Upham, 1895). The ages of beaches and subsequent outbursts are controlled by a limited number of radiocarbon dates. Organic matter is rare in the beaches of Lake Agassiz, so their ages are commonly based on dated glacial events in the Agassiz basin or correlation with events outside of the basin, such as events recorded in the Superior, Huron, and Michigan basins (e.g., Drexler et al., 1983; Teller and Mahnic, 1988; Lewis et al., 1994; Colman et al., 1994) and in outlet channels that carried overflow (Smith and Fisher, 1993; Fisher, 2003). There are several

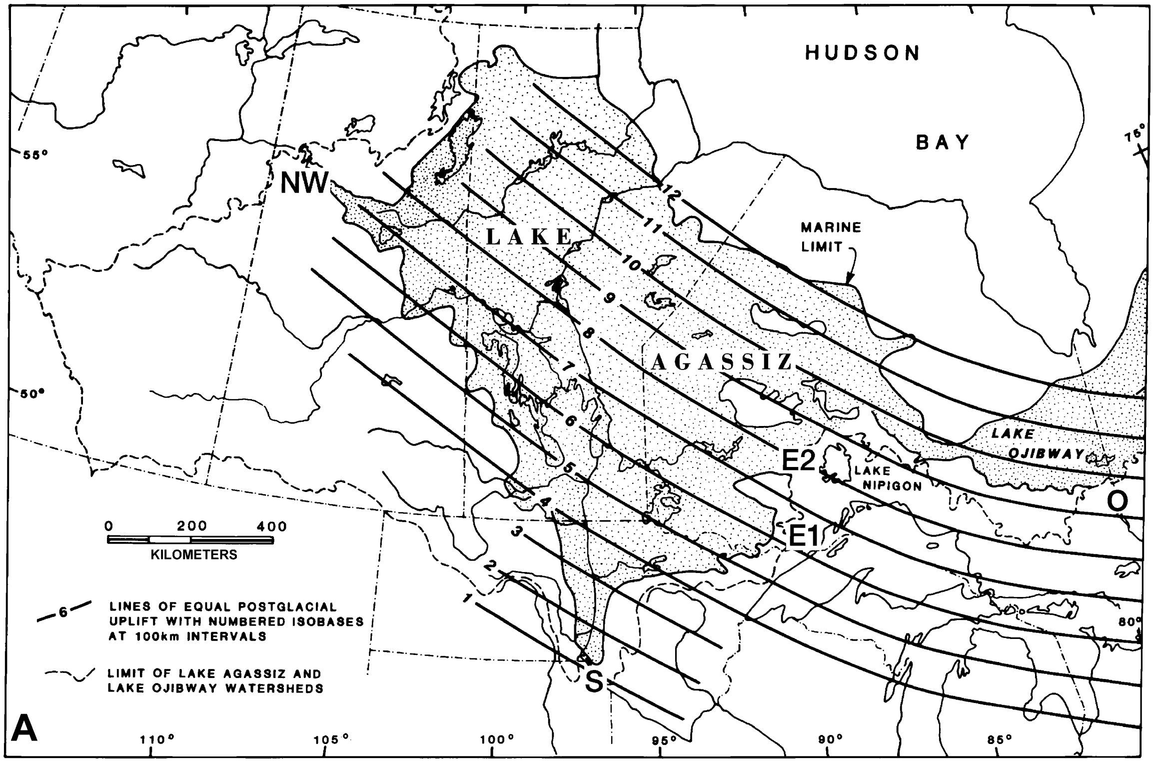


Figure 3. (A) Isobases across the Lake Agassiz-Ojibway basin (Teller and Thorleifson, 1983: after Johnston [1946], Walcott [1972], and Vincent and Hardy [1979]). Lines of equal isostatic rebound (isobases 1–12) are spaced at 100 km intervals. General locations of main outlets are shown (S, NW, E1, E2, and O). (**Caption continued on p. 733.****)**

accepted temporal pins in the Lake Agassiz beach and outburst chronology that are based on radiocarbon dates in the basin; these are related to formation of specific beaches and lie within a generally accepted small age range: (1) the Herman beach (11.0–10.8 ^{14}C ka; Fig. 4A1), (2) Upper Campbell beach (9.4–9.3 ^{14}C ka; Fig. 4D1), and (3) the final drainage of the lake (7.7 ^{14}C ka). The two beaches below the Upper Campbell beach (the Lower Campbell and McCauleyville) (Figs. 4E1 and 4F1) have radiocarbon dates associated with them that indicate that they were deposited shortly after 9400 yr B.P. (Teller et al., 2000). The age of The Pas beach (Fig. 4O1), which dates the end of the routing of Agassiz water into the Superior basin, is 8.2–8.0 ^{14}C ka (Teller and Mahnic, 1988; Thorleifson, 1996; Teller et al., 2002), probably closer to 8.0 ^{14}C ka, on the basis of the paleomagnetically dated change in sediment style in the Superior basin described by Mothersill (1988).

TABLE 1. CONVERSION OF INTERPRETED RADIOCARBON AGES OF LAKE AGASSIZ BEACHES TO THE RANGE OF CALENDAR YEARS (DEFINED BY 1σ) AND MEAN AGES IN YEARS BEFORE PRESENT OF THAT RANGE

Map	Beach name	^{14}C age	Calendar age range	Mean of cal. range
A	Herman	10,900	13,019–12,860	12,940
B	Norcross	10,100	11,697–11,564	11,630
C	Tintah	9900	11,257–11,230	11,240
D	U. Campbell	9400	10,671–10,578	10,620
E	L. Campbell	9300	10,550–10,429	10,490
F	McCauleyville	9200	10,398–10,264	10,330
G	Blanchard	9000	10,208–10,185	10,200
H	Hillsboro	8900	10,149–9935	10,040
I	Emerado	8800	9908–9775	9840
J	Ojata	8700	9684–9600	9640
K	Gladstone	8600	9548–9540	9540
L	Burnside	8500	9527–9494	9510
M	Ossawa	8400	9472–9428	9450
N	Stonewall	8200	9249–9089	9170
O	The Pas	8000	8996–8788	8890
P	Gimli	7900	8701–8636	8670
Q	Grand Rapids	7800	8592–8543	8570
R	Kinojévis	7700	8474–8421	8450

Note: Calendar dates estimated using CALIB 4.3 program of Stuiver and Reimer (1993) and Stuiver et al. (1998). Subsequent drawdowns of the lake (the outbursts) are nearly the same age as the beaches. ''Map'' refers to Figure 4 and Figure DR1 (see footnote 1 in text) lake reconstructions.

The ages of two of the older beaches, the Norcross and Tintah, are controversial. Early researchers attributed them to sequential ''stair-step'' drops in lake level after the Herman beach was formed when lower routes were opened into the Superior basin (e.g., Johnston, 1946; Elson, 1967; Fenton et al., 1983). Dating of two cores in the southern outlet of Lake Agassiz led Fisher (2003) to conclude that this interpretation was correct, and he considered that they formed between ca. 10.9 and 10.8 ^{14}C ka. In contrast, others have concluded that the Norcross and Tintah beaches formed later, after centuries of lower lake level (the Moorhead phase) (e.g., Thorleifson, 1996; Teller et al., 2000; Teller, 2001; Leverington et al., 2000; Fisher and Smith, 1994). These researchers argued that differential isostatic rebound forced Lake Agassiz to rise briefly to the Norcross level sometime

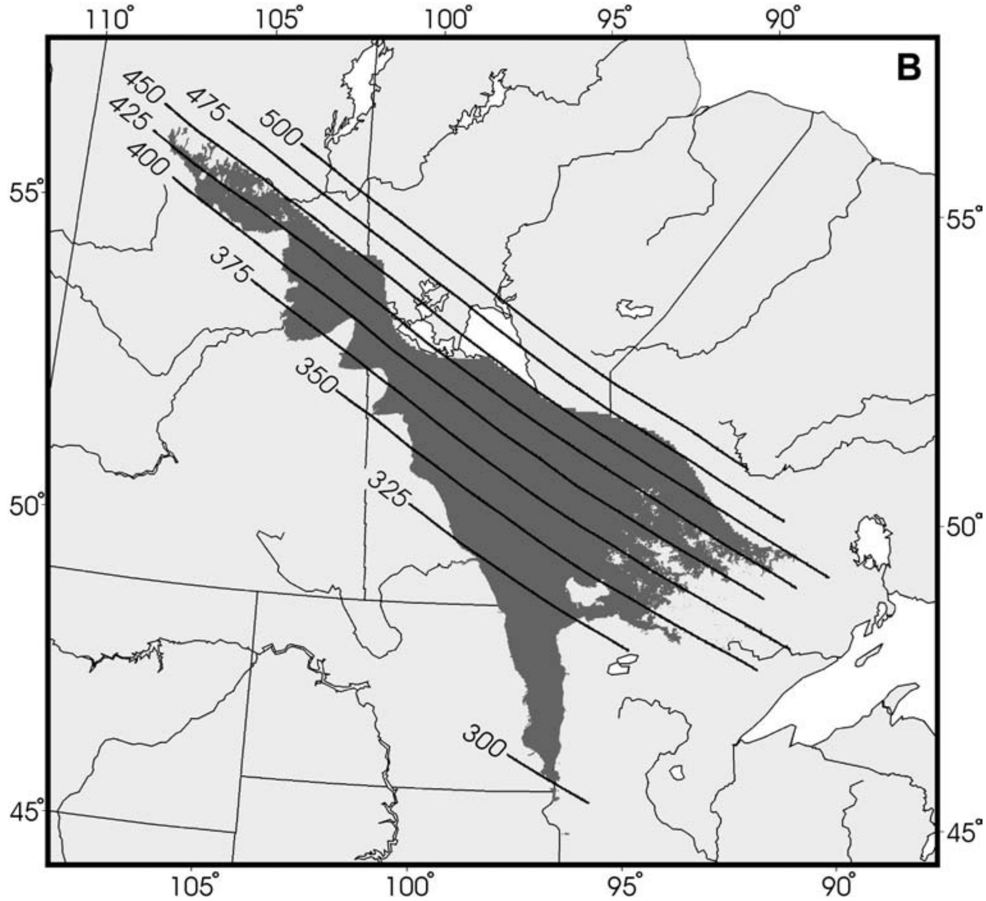


Figure 3. (*Caption continued from p. 732.*) (B) Isobases across the Lake Agassiz basin at *vertical* intervals of 25 m on the rebounded surface associated with the Upper Campbell beach. Maps on older (or younger) beaches, which has undergone more (or less) differential rebound, would show steeper (or gentler) gradients.

between 10.4 and 10.1 ^{14}C ka after the Moorhead low-water phase. Following a drop in lake level, glacial readvance at ca. 10.0 ^{14}C ka closed the newly opened northwestern outlet, forcing the lake to rise to the Tintah level. We adopt the Teller (2001) interpretation in this paper, but we acknowledge that the absence of radiocarbon dates from any of the older beaches and the small number from Agassiz outlets do not allow exact dating of these beaches.

Between 9.4 ^{14}C ka and the final drainage of Lake Agassiz, more than 15 recognized beaches developed in the basin (Johnston, 1946; Elson, 1967; Teller and Thorleifson, 1983); most are linked to a topographically distinct overflow route from the lake (Teller and Thorleifson, 1983; Leverington and Teller, 2003). The ages associated with these lake stages (beaches), and with the outbursts that ended those stages, all lie between 9.4 and 7.7 ^{14}C ka, and it is possible that the time needed to form each beach and the time represented by the intervening transgression were about the same. Thus, 1700 ^{14}C yr (= 2170 cal. yr) divided by 15 outbursts equals an average of 113 radiocarbon years (145 cal. yr) between each beach and each outburst. Complicating the age assignment are the so-called radiocarbon plateau at ca. 10,000 ^{14}C yr B.P. (10.6–10.0 and 9.6 ^{14}C ka) and several other such plateaus during the time of Lake Agassiz overflow (e.g., 8.75 ^{14}C ka and 8.25 ^{14}C ka) (e.g., Bradley, 1999, p. 68) as well as the overall nonequivalence of radiocarbon and sidereal years due to long-term changes in atmospheric ^{14}C content (e.g., Stuiver and Reimer, 1993). Table 1 shows our interpretation of the radiocarbon ages of Lake Agassiz beaches and their conversions to calendar years using the CALIB 4.3 program of Stuiver and Reimer (1993) and Stuiver et al. (1995).

Snapshots of Lake Agassiz Through Time

Although Lake Agassiz began to develop in the southern end of the basin before its first large beach formed at ca. 11.7 ^{14}C ka (13.4 cal. ka) (Fenton et al., 1983), our reconstructions begin with the oldest extensive beaches in the Lake Agassiz basin, the Herman beaches, formed while overflow was through the southern outlet (e.g., Upham, 1895; Elson, 1967; Fenton et al., 1983); the lowest (youngest) of this closely spaced group of beaches is shown in Figure 4A1. The age of the Herman beaches and of the subsequent 110 m drop in lake level is between 11.0 and 10.8 ^{14}C ka (see Licciardi et al., 1999; Fisher, 2003), probably ca. 10.9 ^{14}C ka (12.9 cal. ka). A total of 9500 km^3 of Lake Agassiz waters were released abruptly into the Lake Superior basin near Thunder Bay, Ontario (outlet E1 of Fig. 3A) at this time, and the size of the lake decreased from ~134,000 km^2 to 37,000 km^2. This was the largest outburst until the final drainage of the lake (Teller et al., 2002). On the basis of the interpretation of beach ages by Fisher (2003), as discussed above, this 110 m drop in lake level would have occurred in several steps over about a century, and the Norcross and Tintah beaches formed as Lake

Agassiz overflow eroded the southern outlet below the Herman beach level.

It is important to note that the routing of overflow during this early period remains uncertain. New field work has raised the question of whether overflow from Lake Agassiz was directed east through the Great Lakes (outlet E1 in Fig. 3A) for centuries after the Herman beach was abandoned, as has been the interpretation by all researchers, or whether overflow may have been through the Clearwater outlet to the Athabasca-Mackenzie valley system (outlet NW in Fig. 3A) between 10.8 and 9.4 ^{14}C ka (12.8–10.6 cal. ka). The latter requires that the northwestern outlet was deglaciated shortly after 11 ^{14}C ka, which is much earlier than most evidence indicates. Regardless of the routing of overflow at this time, the extent of Lake Agassiz is not likely to have been much different except along its western margin. The computer-generated extent of the lake immediately *after* this drawdown is shown in Figure 4A2; as previously discussed, this lake is not represented by a beach because subsequent rising waters reworked shoreline sediment upslope.

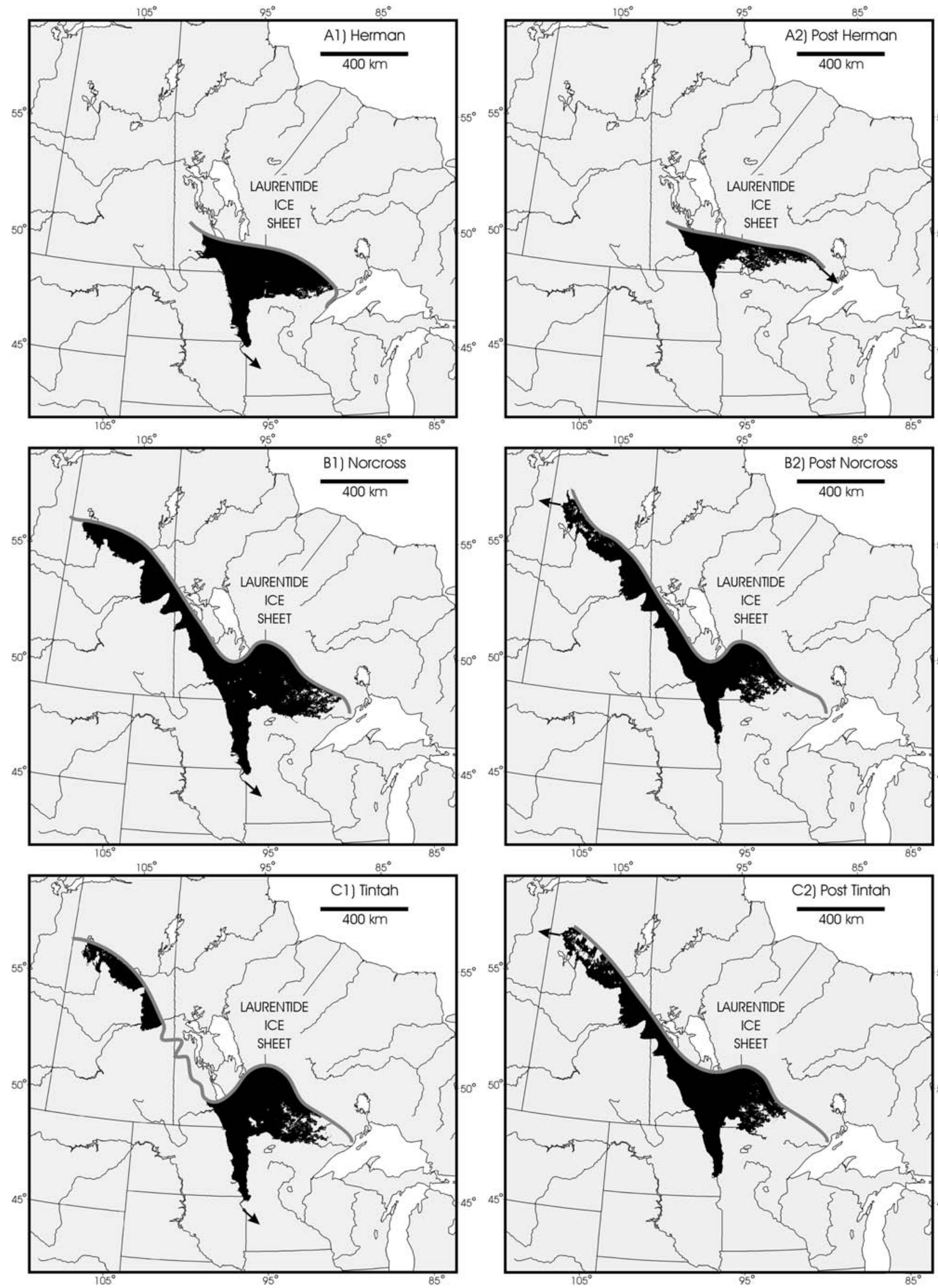
During the next 700–800 yr, waters deepened to the south of the isobase through the eastern outlet (isobase 6 in Fig. 3A), and the lake margin transgressed over the dry lake bed; waters shallowed in the region between that isobase and the Laurentide Ice Sheet margin (see Fig. 2B). There are many radiocarbon dates on vegetation that was buried by sediments deposited during this transgression (see summary of dates in Appendix B of Licciardi et al., 1999). The Norcross beach represents the maximum extent of this transgression just before 10.1 ^{14}C ka (Teller, 2001) (Fig. 4B1); it also dates the start of the next drop in lake level. Figure 4B2 shows the extent of Lake Agassiz following the opening of the northwestern outlet to the Arctic Ocean (outlet NW in Fig. 3A), based on the work of Fisher and Smith (1994). Differential isostatic rebound, combined with the redamming of the northwestern outlet (Thorleifson, 1996), raised water levels to the Tintah beach level by ca. 9.9 ^{14}C ka (Fig. 4C1), prompting renewed overflow through the southern outlet for a short time (Teller, 2001). When ice retreated again from the northwestern outlet, lake level abruptly dropped, and the extent of the lake decreased again (Fig. 4C2). The difference between our interpretation of beach formation during this period of lake history and that of Fisher (2003) is that he has related formation of the Norcross and Tintah beaches to rapid erosion of the southern outlet immediately following formation of the Herman beach.

Over the next 400 yr, between ca. 9.8 and 9.4 ^{14}C ka, greater isostatic rebound of the northwestern outlet (on isobase 7, Fig. 3A) compared to that of the southern outlet (isobase 1) caused Lake Agassiz to transgress southward until it reached the channel that had previously carried overflow at the southern end of the basin; this process resulted in the stranding of the largest and most extensive beach in the basin, the Upper Campbell beach, at ca. 9.4 ^{14}C ka (10.6 cal. ka), as shown by the outline of the lake in Figure 4D1. Within a few years, a lower *eastern* outlet channel from Lake Agassiz was deglaciated, and waters were again routed into the Great Lakes, this time via the Lake Nipigon basin (outlet E2, Fig. 3A), and the level of the lake abruptly dropped (Fig. 4D2).

Subsequent abrupt declines in lake level occurred when new, lower, eastern outlet channels were deglaciated; following each drop there was a new deepening of waters (transgression) south of the isobase through the outlet, as there had been between previous drawdowns. Although there are a few other Lake Agassiz beaches, which lie between these transgressive maximums, most are not well developed and may be storm beaches or offshore bars (see Teller, 2001). Figure 4 shows lake stages related to five transgressive maximums and subsequent minimums after overflow was rerouted eastward (G1-G2, I1-I2, J1-J2, N1-N2, O1-O2), and three late-stage maximums routed through the Ottawa River Valley after ca. 8.0 ^{14}C ka (8.9 cal. ka) (Fig. 4P, 4Q, 4R). Six map pairs showing maximum and minimum stages during this period are not shown in Figure 4, but are available in Figure DR1 (see footnote 1; i.e., map pairs E1-E2, Lower Campbell; F1-F2, McCauleyville: H1-H2, Hillsboro; K1-K2, Gladstone; L1-L2, Burnside; and M1-M2, Ossawa).

At ca. 8.0 ^{14}C ka, Lake Agassiz amalgamated with glacial Lake Ojibway, which had been expanding independently in eastern Ontario and adjacent Quebec (Fig. 3A); it is possible that the drawdown of Lake Agassiz following formation of The Pas beach (Fig. 4O2) may have resulted from this amalgamation. Overflow was subsequently routed out through the Ottawa River Valley; the Kinojévis outlet carried overflow just before the final drainage of the lake (e.g., Vincent and Hardy, 1979) at ca. 7.7 ^{14}C ka (8.45 cal. ka). Figure 4R shows the outline of the lake at this time, which extended over an area of 841,000 km^2 and is correlated to the Ponton beach in the

Figure 4. Snapshots of the history of Lake Agassiz showing the extent of the lake (black area) at various times; maps of the 12 lake stages in this sequence that are not shown here (map pairs E1-E2, F1-F2, H1-H2, K1-K2, L1-L2, and M1-M2; in square brackets in this caption) are available in Figure DR1 (see footnote 1 in text). Each of the map pairs (A1-A2 to O1-O2) shows the transgressive maximums (which are based on mapped beaches) and the subsequent postoutburst minimums; each minimum is followed by a new transgressive phase (south of the outlet) caused by differential isostatic rebound. The last three lake stages (P, Q, and R) are depicted only by transgressive maximums. Values given below for lake-level decline are from Teller et al. (2002) and Leverington and Teller (2003) and relate to change in water level estimated in the outlet area at that time. Arrows show routing of outburst and baseline overflow at time of each lake stage. See comments in text about position of ice margins, Table 1 for age of each lake stage, and Figure 5 for routing of overflow and relationship to Greenland isotopic record. (A1) Herman beach stage (lower level) and (A2) lake following 110 m drawdown. (B1) Norcross beach stage and (B2) lake following 52 m drawdown. (C1) Tintah beach stage and (C2) lake following 30 m drawdown. (D1) Upper Campbell beach stage and (D2) lake following 30 m drawdown. [(E1) Lower Campbell beach stage and (E2) lake following 11 m drawdown.] [(F1) McCauleyville beach stage and (F2) lake following 16 m drawdown.] (G1) Blanchard beach stage and (G2) lake following 10 m drawdown. [(H1) Hillsboro beach stage and (H2) lake following 8 m drawdown.] (I1) Emerado beach stage and (I2) lake following 18 m drawdown. (J1) Ojata beach stage and (J2) lake following 20 m drawdown. [(K1) Gladstone beach stage and (K2) following 9 m drawdown.] [(L1) Burnside beach stage and (L2) lake following 13 m drawdown.] [(M1) Ossawa beach stage and (M2) lake following 15 m drawdown.] (N1) Stonewall beach stage and (N2) lake following 58 m drawdown. (O1) The Pas beach stage and (O2) lake following 18 m drawdown. (P) Gimli beach stage. (Q) Grand Rapids beach stage. (R) Kinojévis beach stage; just before Lake Agassiz-Ojibway drained into the Tyrrell Sea.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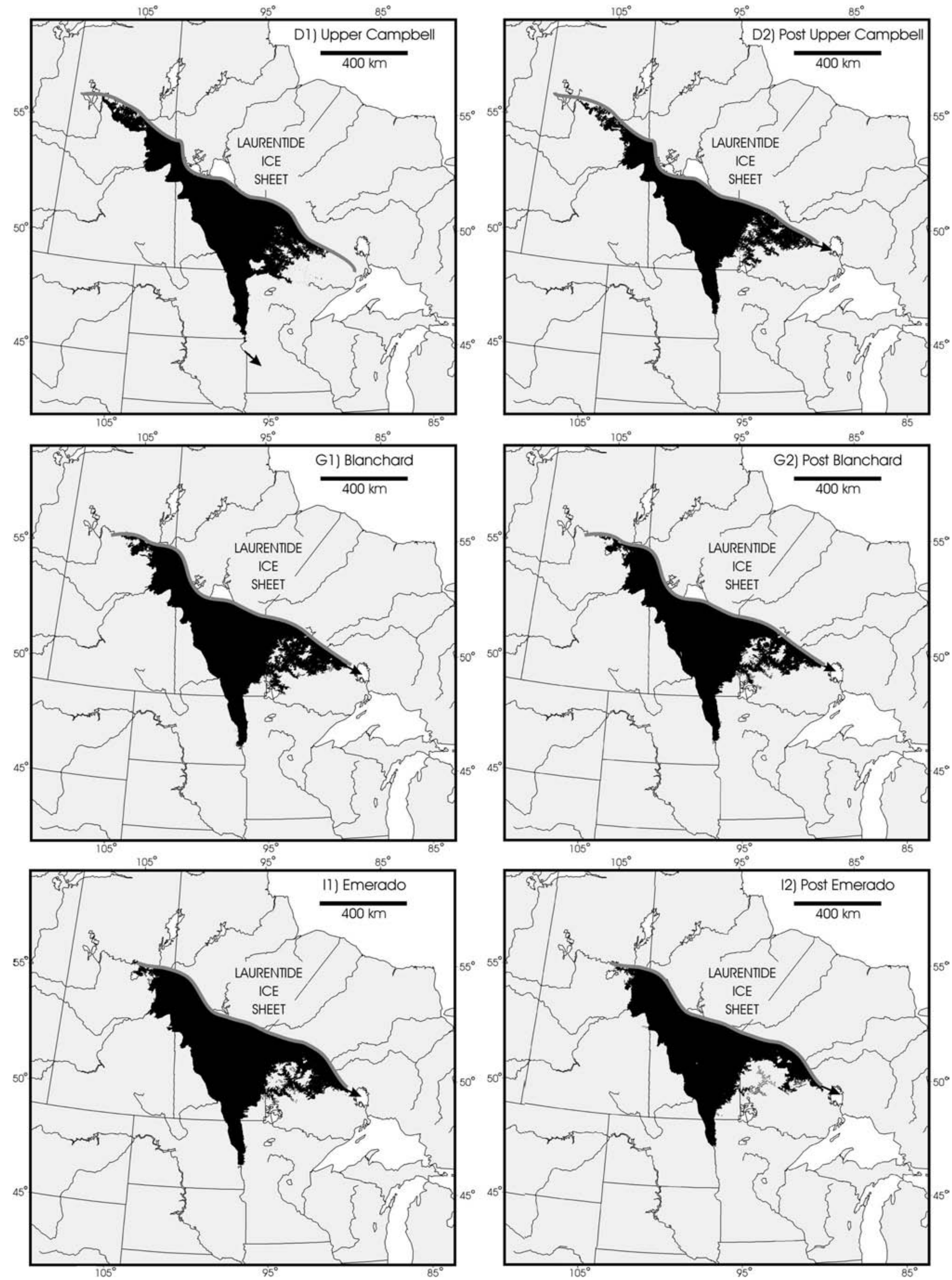

Figure 4. (*Continued.*)

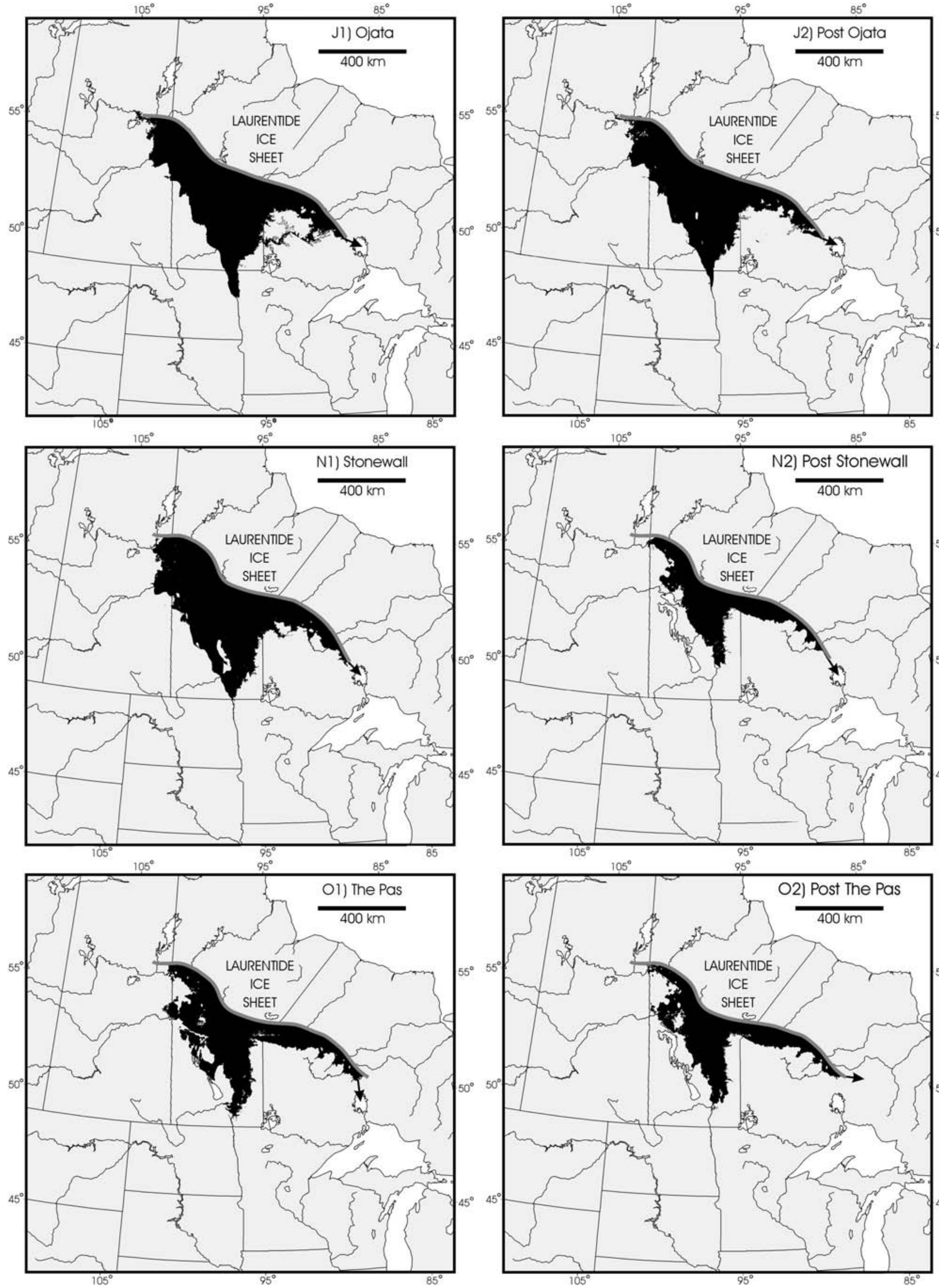

Figure 4. (*Continued.*)

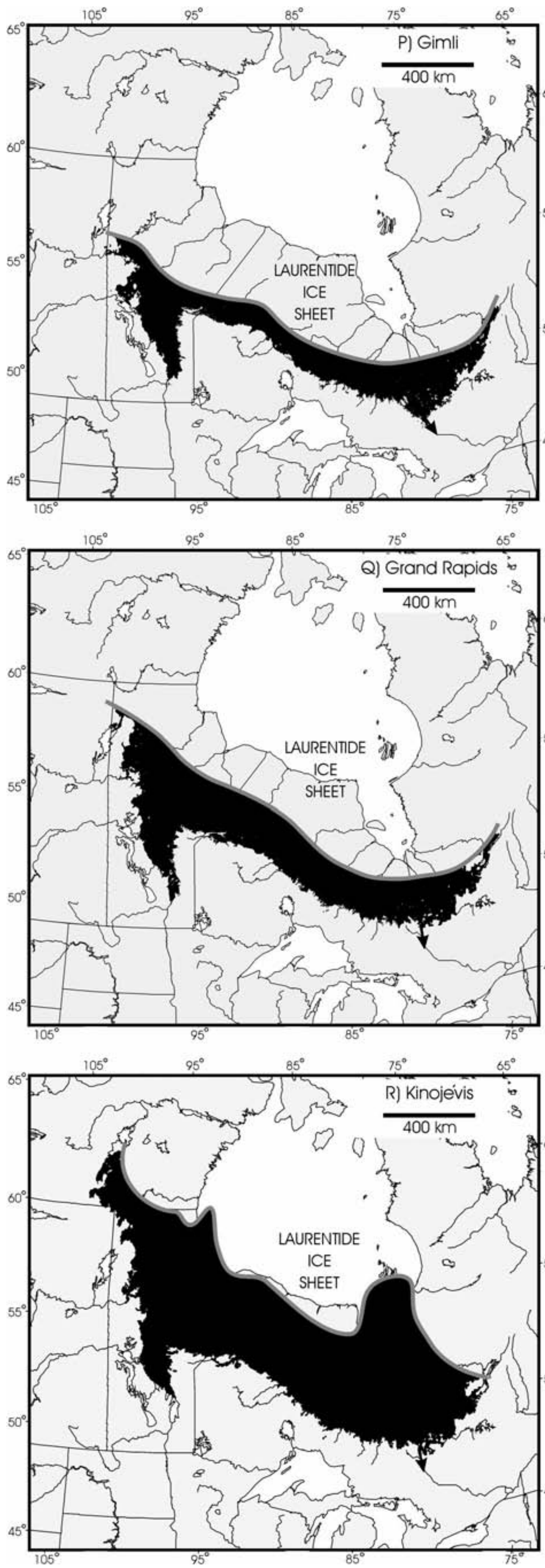

Figure 4. (*Continued.*)

Agassiz part of the basin (Leverington et al., 2002a). An alternative two-step scenario for the final drainage of Lake Agassiz was discussed by Leverington et al. (2002a), where the eastern (Ojibway) part of the lake completely drained but only part of the western region did, leaving a residual lake of ~408,000 km^2 in Manitoba and adjacent northern Ontario (Leverington et al., 2002a, Fig. 2F). Following the final drainage of Lake Agassiz-Ojibway, waters of the Tyrrell Sea transgressed south over lacustrine sediments in the Hudson Bay Lowland (''marine limit'' of Fig. 3A), and modern drainage was established across the old lake bed.

RELATIONSHIP OF LAKE AGASSIZ OUTBURSTS TO THE GREENLAND ISOTOPIC RECORD

Influxes of fresh water to the North Atlantic Ocean from North America have been correlated with changes in thermohaline circulation (THC) and climate (e.g., Broecker et al., 1985, 1989; Rahmstorf, 1995; Manabe and Stouffer, 1995, 1997; Alley et al., 1997; Clark et al., 2001). This relationship has been linked to the main cooling episodes that interrupted late glacial warming, namely, Heinrich 1, the Younger Dryas, the Preboreal Oscillation, and the 8.2 cal. ka cooling. Changes in the site of meltwater delivery to the oceans during late-glacial time (associated with changes in continental-scale glacial drainage basins) and short high-flux injections of water related to catastrophic outbursts from Lake Agassiz have been interpreted as possible forcing mechanisms (Clark et al., 2001; Teller et al., 2002). Some outbursts from Lake Agassiz may have triggered changes in THC, and these changes may have been sustained when there was an associated re-rerouting of overflow to a different ocean.

Some evidence for episodes of climatic cooling comes from the oceans, but the best record of temperature fluctuation during the 5000-cal.-yr-long period when Lake Agassiz waters may have had an impact on THC and climate is found in the isotopic and ionic records of the GRIP and GISP2 ice cores in Greenland (e.g., O'Brien et al., 1995; Grootes and Stuiver, 1997; Johnsen et al., 2001). Given the potential impact on climate of freshwater injections to the North Atlantic Ocean, it is relevant to compare the chronology of Lake Agassiz outbursts to the isotopic record of Greenland.

Figure 5 shows the fluctuations of $\delta^{18}O$ values in the GISP2 and GRIP ice cores of Greenland, which serve as proxies for tem-

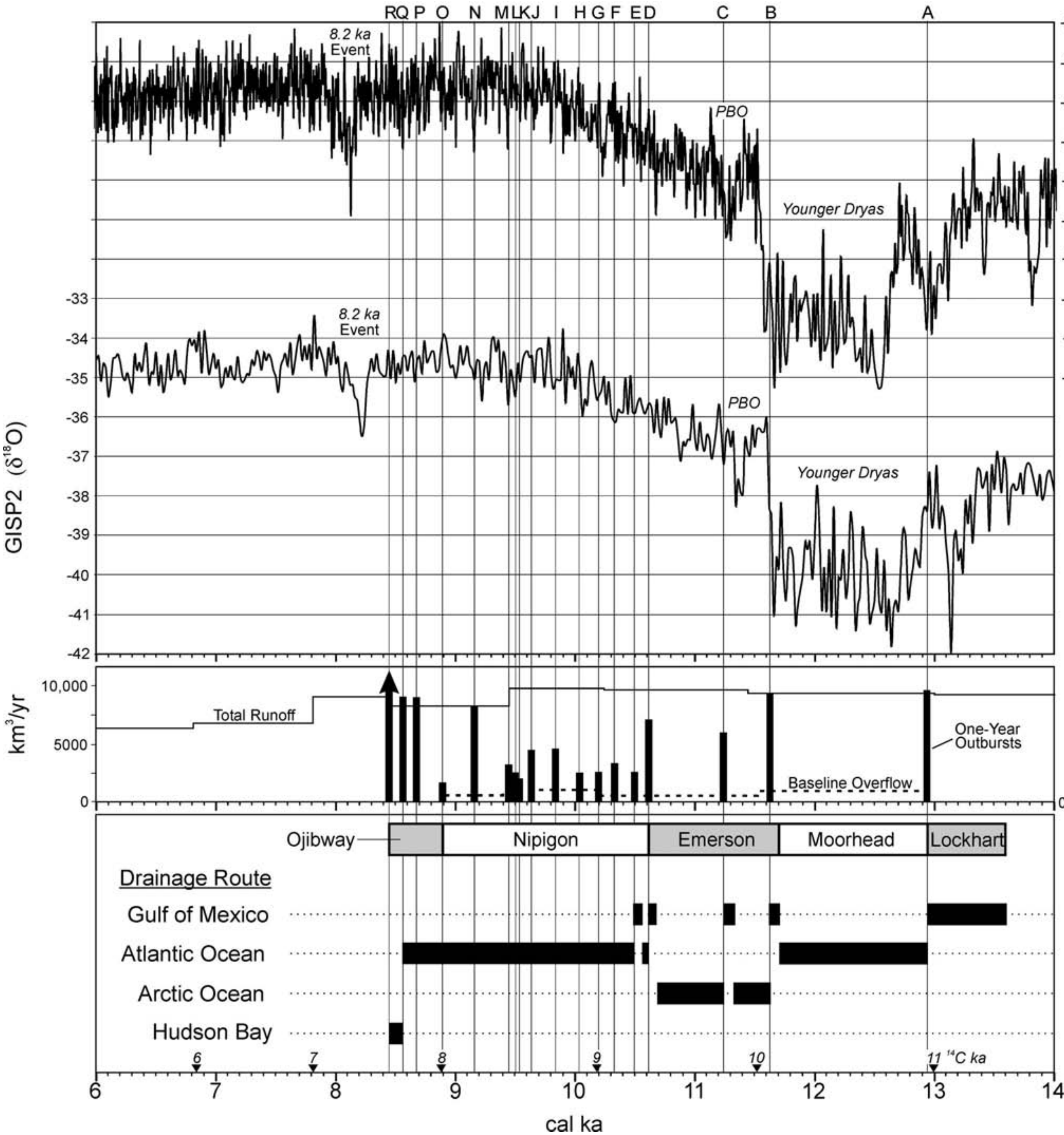

Figure 5. Record of $\delta^{18}O$ values in Greenland ice of GRIP (Johnsen et al., 2001) and GISP2 (Stuiver et al., 1995) cores plotted with the record of catastrophic outbursts from Lake Agassiz (A–R); GRIP data are 55 cm averages, and GISP2 data are bidecadal (isotopic data from Cross, 2002). Negative isotopic excursions related to the Younger Dryas, Preboreal Oscillation (PBO), and 8.2 ka event are identified. Ages are in calendar years B.P.; selected ages are shown in radiocarbon years. Each Lake Agassiz outburst has been interpreted as occurring in ~1 yr and is represented by a bar whose height relates to the total volume of the outburst (Teller et al., 2002); the final outburst was 163,000 km^3 and is indicated by an arrow. Note that if each lake drawdown occurred in 1 yr or less, as suggested by Teller et al. (2002), a 10,000 km^3 outburst would result in a flux of 0.32 Sv (i.e., 320,000 $m^3 \cdot s^{-1}$ for 1 yr). Dashed line represents baseline overflow from Lake Agassiz. The "total runoff" includes all flow through routes A, B, C, and D of Figure 1 (Licciardi et al., 1999, Appendix A) but excludes the outbursts; note how little flow changed through time.

perature variation (Johnsen et al., 2001). Also shown are the times (and magnitudes) of Lake Agassiz outbursts; the radiocarbon dates of these outbursts have been converted to calendar years as shown in Table 1.

Negative excursions in $\delta^{18}O$ are correlated with decreases in atmospheric temperature (e.g., Dansgaard, 1961; Bradley, 1999) and, in turn, with climate cooling in the North Atlantic region; some have related these isotopic excursions to the influx of freshwaters that inhibited THC and delivery of warm waters into high latitudes (e.g., Broecker et al., 1989; Björck et al., 1996; Barber et al., 1999; Clark et al., 2001; Renssen et al., 2001). Because the dating of isotopic excursions in the Greenland ice cap is controlled by the measurement of annual snow layers, the excursions' ages in calendar years have generally been accepted, although there are variable differences of up to nearly 200 yr in interpreted age between the GISP2 and GRIP ice cores; events in the GRIP core generally show younger ages than those in the GISP2 core (Southon, 2002), as can be seen in Figure 5. These age differences are very important when trying to establish a cause-and-effect relationship between climate change and Agassiz flood outbursts that are spaced at 100–200 yr, especially because the impact of freshwater on THC and, in turn, on climate and, in turn, on the isotopic record of Greenland is not known with any certainty. Models show changes in THC to be variable in response time, in duration, and in the magnitude of change in flux and temperature, as a result of variation in duration, magnitude, and site of influx to the ocean. Furthermore, the time of the freshwater flux into the ocean may greatly influence the THC response, because oceans can be pushed over critical thresholds when combined with other contemporary forcing events (cf. Alley et al., 2001) and when oceans are in circulation modes that make them more vulnerable to change (e.g., Fanning and Weaver, 1997). In short, both the ocean response to freshwater forcing and the freshwater forcing itself are complex, and changes are nonlinear (see Rahmstorf, 2000; Ganopolski and Rahmstorf, 2001).

Lake Agassiz Outbursts and the Greenland $\delta^{18}O$ Isotopic Record

Several $\delta^{18}O$ isotopic excursions in the Greenland record have been correlated with outbursts from Lake Agassiz or with the redirection of Lake Agassiz overflow (Fig. 5), which may have reduced the flux of warm waters into the North Atlantic:

1. The Younger Dryas cooling began at ca. 12.9 cal. ka and is clearly recorded in the Greenland isotopic record (Fig. 5). Broecker et al. (1989), Fanning and Weaver (1997), Clark et al. (2001), Teller et al. (2002), and others related the start of this well-known cooling to either a Lake Agassiz outburst and/or the redirection of Agassiz overflow into the North Atlantic at 12.9 cal. yr ka. This is the time of the first (Herman) drawdown of Lake Agassiz (Figs. 4A1 to 4A2), which initiated a period of overflow from Agassiz to the North Atlantic Ocean that lasted for more than 1000 cal. yr. Contrary to the interpretation that large fluxes of Agassiz water led to a reduction in THC, it is interesting to note that the 11.6 cal. ka outburst occurred at the end of the negative isotopic excursion associated with the Younger Dryas and close to the time when overflow from Lake Agassiz shifted from being routed through the Great Lakes to being routed into the Arctic Ocean (Fig. 5). An analysis of dated lacustrine records, tree rings, and the GRIP ice core by Björck et al. (1996) places the end of the Younger Dryas at 11.45–11.39 cal. ka, which is 100–200 yr later than is indicated in the isotopic records of the GISP2 ice core (Fig. 5).

2. The Preboreal Oscillation (PBO) began with a slow cooling almost immediately after the end of the Younger Dryas and ended with an abrupt short cold episode between 11.4 and 11.3 cal. ka, as recorded in the GISP2 core (Fig. 5). In the GRIP ice core, the PBO cooling is dated ~200 yr later at 11.2–11.05 cal. ka, which immediately followed the Lake Agassiz outburst at 11.2 cal. ka (Fig. 5) as noted by Fisher et al. (2002). In contrast, in the GISP2 core, the PBO *precedes* the 11.2 cal. ka outburst but follows the 11.6 cal. ka outburst by several centuries. Complicating the attempt to link Agassiz outbursts with the PBO is the fact that there was a comparable outburst of water into the North Atlantic Ocean from the Baltic Ice Lake of Europe just before the PBO, which probably had an impact on THC and contributed to the PBO cooling event (Björck et al., 1996). We think that the chronological association of the PBO cold event—recorded in both Greenland and northwestern Europe 200–300 years after the end of the Younger Dryas (Björck et al., 1996)—with large post–Younger Dryas outbursts from Lake Agassiz and the Baltic Ice Lake suggests a cause-and-effect relationship.

3. The 8.25–8.1 cal. ka negative $\delta^{18}O$ excursion in the GISP2 core, and shortly after that in the GRIP core (Fig. 5), followed the final drainage (outburst) from Lake Agassiz, dated at ca. 8.4 cal. ka. A number of researchers have linked this increased freshwater flux with the 8.2 ka isotopic event (e.g., Alley et al., 1997; Barber et al., 1999; Clark et al., 2001; Renssen et al., 2001; McDermott et al., 2001; Teller et al., 2002; Clarke et al., 2003, 2004). Other cooling events dated between 8.4 and 8.0 cal. ka have been recorded in Europe, North America, and elsewhere (e.g., von Grafenstein et al., 1998; McDermott et al., 2001; Hughen et al., 1996). Dean et al. (2002) associate the abrupt change in proportion of land, water, and ice at the time of the drainage of Lake Agassiz with a fundamental change in atmospheric circulation.

It is interesting to note that the 7.7 ^{14}C ka (8.45 cal. ka) mean calibrated age of the first postglacial marine shells in Hudson Bay, determined by Barber et al. (1999) and correlated with the final drainage of Lake Agassiz, precedes the start of the Greenland isotopic excursion by ~150 yr in the GISP2 core and ~250 yr in the GRIP core (Fig. 5). This difference in age between the freshwater influx and isotopic response could be reduced by increasing the ''local deviation'' (ΔR) for ''global-mean surface reservoir age for carbonate'' used by Barber et al. (1999) to values closer to those found in Hudson Bay itself. Alternatively, we suggest that this seeming lag in response can be accounted for by the two-step model for the final drainage of Lake Agassiz that was proposed by Leverington et al. (2002a), in which the second step of the final drawdown provided the critical mass of freshwater that triggered a change in THC in an ocean that had reached a relatively stable interglacial mode. In fact, a two-stage cooling around the time of the 8.2 ka event has been identified in speleothems of Ireland (Baldini et al., 2002) and lacustrine records in Norway (Nesje and Dahl, 2001), and a two-step release of Lake Agassiz waters has been modeled by Clarke et al. (2004).

It is possible that the small isotopic excursions in the GISP2 core at ca. 10.4–10.3, 10.1–10.0, and 9.8–9.9 cal. ka are related to small Agassiz outbursts (Fig. 5). In an analysis of $\Delta^{14}C$ ka records of German oak and pine, Björck et al. (1996) noted a distinct anomaly at 10.1 cal. ka that corresponds to a 1.5–2.0 °C temperature drop in the GRIP core, as well as other $\Delta^{14}C$ anomalies at ca. 11.0 and 9.4 cal. ka. However, correlation of these small isotopic excursions with relatively small Agassiz outbursts is very tenuous, partly because there is uncertainty in precisely dating the outbursts and partly because there are differences in chronologies between the GISP2 and GRIP cores.

SUMMARY AND CONCLUSIONS

Lake Agassiz had a long history of frequent and abrupt changes in size and depth. As the Laurentide Ice Sheet retreated, progressively lower outlets from Lake Agassiz carried overflow to the oceans through four main routes (Fig. 5). These outflow changes, in combination with differential isostatic rebound, resulted in a lake that repeatedly expanded and abruptly contracted (Fig. 4). Beaches and wave-cut strandlines record the transgressive maximums of at least 18 stages between 10.9 and 7.7 ^{14}C ka (12.9 and 8.45 cal. ka). Between these maximums were low-level stages that developed when new and lower outlets opened and brought about an abrupt drawdown of the lake. These outbursts were followed by a slow deepening in the southern part of the basin, as waters south of the outlet transgressed because of differential isostatic rebound.

The largest Lake Agassiz outbursts were those related to abrupt drawdowns in lake level from the Herman, Norcross, Tintah, Upper Campbell, Stonewall, and Kinojévis lake stages (Figs. 4A, 4B, 4C, 4E, 4N, and 4R). Three of these outbursts occurred near the start of large δ^{18}O isotopic excursions in the GISP2 and GRIP ice cores from Greenland (Fig. 5), namely, those related to the Younger Dryas, Preboreal Oscillation, and 8.2 cal. ka cooling. These large Agassiz outbursts may have triggered those cool episodes by affecting the THC. Smaller outbursts have no clear relationship to the isotopic excursions in the ice cores. We suggest that the difference between the estimated 8.45 cal. ka age of the final drainage of Lake Agassiz and the 8.2 cal. ka cooling event may be accounted for by a two-step drainage of Lake Agassiz or by increasing the ‘‘local deviation’’ for the marine carbonate reservoir effect used by Barber et al. (1999) to date the final drainage.

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