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Changes in the Bathymetry and Volume of Glacial Lake Agassiz Between 11,000 and 9300 ^{14}C yr B.P.

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The volume and surface area of glacial Lake Agassiz varied considerably during its 4000-year history. Computer models for seven stages of Lake Agassiz were used to quantify these variations over the lake's early history, between about 11,000 and 9300 ^{14}C yr B.P. (ca. 13,000 to 10,300 cal yr B.P.). Just after formation of the Herman strandlines (ca. 11,000 ^{14}C yr B.P.), the volume of Lake Agassiz appears to have decreased by >85% as a consequence of the abrupt rerouting of overflow to its eastern outlet from its southward routing into the Mississippi River basin. This drainage released about 9500 km^3 of water into the North Atlantic Ocean via the Great Lakes and Gulf of St. Lawrence. Following closure of this eastern routing of overflow, the lake reached its maximum size at about 9400 ^{14}C yr B.P. with an area of >260,000 km^2 and a volume of >22,700 km^3. A second major reduction in volume occurred shortly after that, when its volume decreased >10% following the opening of the Kaiashk outlet to the east into the Great Lakes, and 2500–7000 km^3 of water was released into the North Atlantic Ocean. These discharges may have affected ocean circulation and North Atlantic Deep Water production.

Key Words: **Lake Agassiz; bathymetry; volume; computer model.**

INTRODUCTION

Lake Agassiz was the largest lake in North America during the last deglaciation, covering an area of nearly 1 million km^2 during its 4000-yr history (Teller and Clayton, 1983) (Fig. 1). Because of its large area and volume, the lake had considerable impact on various aspects of late-glacial North America, including the dynamics of the Laurentide Ice Sheet (Clayton *et al.*, 1985), regional climate (Hu *et al.*, 1997; Hostetler *et al.*, 2000), early human history (Pettipas and Buchner, 1983), rivers and lakes that received its overflow (e.g., Teller, 1985, 1987, 1990a; Lewis *et al.*, 1994), and the ocean–climate system (e.g., Broecker *et al.*, 1989; Teller, 1990b; Licciardi *et al.*, 1999; Barber *et al.*, 1999).

In this paper, we present area and volume calculations and bathymetric maps of Lake Agassiz during the early stages of its history, between about 11,000 and 9300 ^{14}C yr B.P. (about 13,000 to 10,300 cal yr B.P. (Stuiver and Reimer, 1993)). This research builds on previous modeling work (Mann *et al.*, 1999) in the following ways: (1) by investigating a sequence of seven lake stages that span an important and dynamic period in the history of Lake Agassiz; (2) by using a topographic database that greatly improves spatial resolution over that of previous modeling efforts; (3) by using a revised methodology in which strandlines were not explicitly defined in advance of modeling, allowing increased shoreline detail to be produced in the bathymetric models; and (4) by estimating volume changes for two major lake-level declines that followed the abrupt opening of new outlets.

LAKE AGASSIZ FROM 11,000 TO 9300 ^{14}C yr B.P.

The history of glacial Lake Agassiz is complex, and its level rose and fell dramatically many times between its inception about 11,700 ^{14}C yr B.P. and its final drainage into Hudson Bay about 7500 ^{14}C yr B.P. (e.g., Elson, 1967; Fenton *et al.*, 1983; Klassen, 1983; Fisher and Smith, 1994; Thorleifson, 1996). Continuing differential isostatic rebound caused a relatively rapid rise in the northeastern part of the basin, which caused lake transgressions toward the south (e.g., Johnston, 1946; Teller and Thorleifson, 1983). Outlet erosion reduced the level of the lake, and retreat of the Laurentide ice margin resulted in the abrupt opening of new outlets that brought about episodes of rapid drawdown of the lake. At other times, ice-marginal readvances and isostatic rebound led to the closure of these outlets and a consequent rise (transgression) of the lake (e.g., Teller and Thorleifson, 1983).

The earliest phases of Lake Agassiz began about 11,700 ^{14}C yr BP (Fenton *et al.*, 1983) and are only recorded as discontinuous high beaches in the southern end of the basin and as high-level routings of its overflow (Bluemle, 1974; Hobbs, 1983; Fenton *et al.*, 1983). The first widespread beaches were named the Herman beaches by Upham (1895). They can be traced from the lake's southern outlet north into Manitoba, and they formed sometime between about 11,700 and 11,000 ^{14}C yr B.P. (Fenton *et al.*, 1983). Our first reconstructions of the bathymetry of Lake Agassiz are for this extensive early high-level stage.

Although there has been some disagreement about the history of Lake Agassiz after the lake stood at the Herman stage, most interpretations (e.g., Fenton *et al.*, 1983; Thorleifson,

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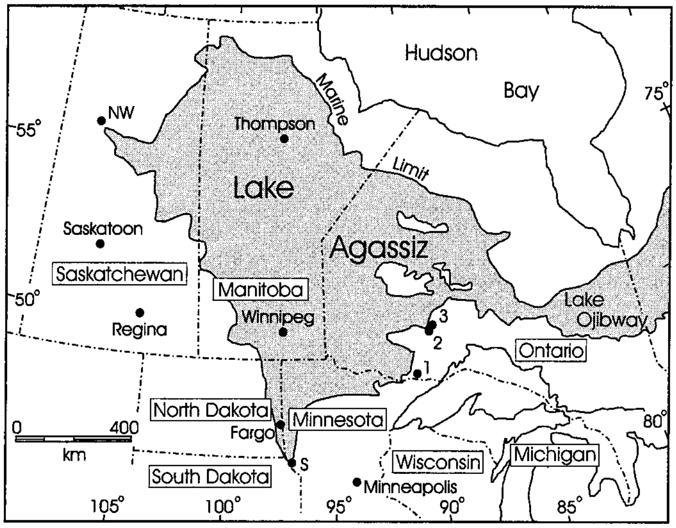

FIG. 1. Map showing total geographic extent of Lake Agassiz during its 4000-yr history (after Teller *et al.,* 1983). The lake margin closest to Hudson Bay marks the marine limit (i.e., the extent that the Tyrrell Sea transgressed over Lake Agassiz sediments after Lake Agassiz drained completely at about 7500 ^{14}C yr B.P.). Lake Agassiz outlets related to outflow between 11,000 and 9300 ^{14}C yr B.P. are labeled as follows: 1, the Kaministikwia route, west of Thunder Bay; 2, the Kaiashk route; 3, the Kopka route; S, the southern outlet (Minnesota River Valley); NW, the northwestern outlet (Clearwater spillway).

1996; Teller *et al.,* in press) suggest that a lower eastern outlet into the Great Lakes was opened shortly after 11,000 ^{14}C yr BP (Fig. 2). This resulted in a significant drop in lake level in the central part of the basin and a large reduction in the areal extent. Using Thorleifson's (1996) reconstructions for this low-water phase (the Moorhead phase), we modeled the bathymetry of Lake Agassiz for two stages (the early and late Moorhead).

Overflow during the Moorhead stages was through the eastern outlets of Lake Agassiz. Isostatic rebound of these outlets produced a transgression of the lake over the subaerially exposed surface of the basin floor until the elevation of that outlet exceeded that of the previously abandoned southern outlet. Erosion of the southern outlet again caused the level of the lake to fall, and the Norcross shoreline, which had formed around the lake margin, was abandoned about 10,200 ^{14}C yr B.P. This was followed by the opening of a still lower outlet to the northwest, the Clearwater spillway (Fig. 2) (Smith and Fisher, 1993; Fisher and Smith, 1994). Readvancing ice then closed the northwestern outlet and forced the lake to transgress back to the southern outlet, where the Tintah beach formed. The bathymetry of both the Norcross and the Tintah lake stages are reconstructed in this paper.

Retreating ice reopened a lower outlet, probably the northwestern route, for Lake Agassiz overflow about 9800 ^{14}C yr B.P., and the level of the lake abruptly fell again. Over the next 400 to 500 years, while the outlet carried overflow, the more rapid isostatic rebound in the north produced a southward transgression of the lake until its level once again reached the southern outlet. Following a brief period of overflow through the southern outlet, lower channels to the east into the Nipigon–Superior basins became available about 9400 ^{14}C yr B.P. as a result of ice retreat (beginning with the Kaiashk outlet) (Figs. 1 and 2); this led to the rapid abandonment of the Upper Campbell strandline (e.g., Elson, 1967; Fenton *et al.,* 1983), which had formed by this transgression. Our reconstruction includes this stage of the lake and the subsequent lower-elevation stage which resulted in the formation of the Lower Campbell beach.

METHODOLOGY

In this research, a computer model was used to generate bathymetric maps of seven stages of Lake Agassiz: Herman, early Moorhead, late Moorhead, Norcross, Tintah, Upper Campbell, and Lower Campbell. For each stage, lake bathymetry was calculated by subtracting a stage-specific *rebound surface* from a database of modern elevations. A rebound surface is generated from a database of values interpolated from isobase lines and describes the relative glacio-isostatic rebound that has occurred over a region since a given time (Mann *et al.,* 1999). The geometry of a rebound surface matches the geometry of the corresponding (and now differentially rebounded) water plane.

All topographic and rebound information used here was expressed in the form of raster (i.e., spatially gridded) data layers. Each layer was defined as a rectangular grid of square cells, each cell being assigned a value. The dimensions and cell sizes of each layer in the database were identical, such that the bathymetry of a given stage could be calculated by subtracting,

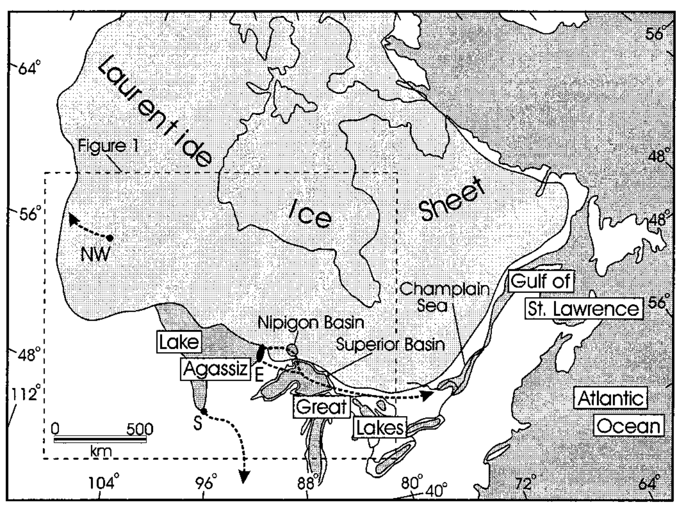

FIG. 2. Schematic map of the Laurentide Ice Sheet shortly after 11,000 ^{14}C yr B.P. (modified after Broecker *et al.,* 1989) showing general locations of Lake Agassiz drainage routes; outlets are labeled as follows: S, the southern outlet (Minnesota River Valley); E, the eastern outlets (including the Kaministikwia, the Kaiashk, and the Kopka routes); NW, the northwestern outlet (Clearwater spillway).

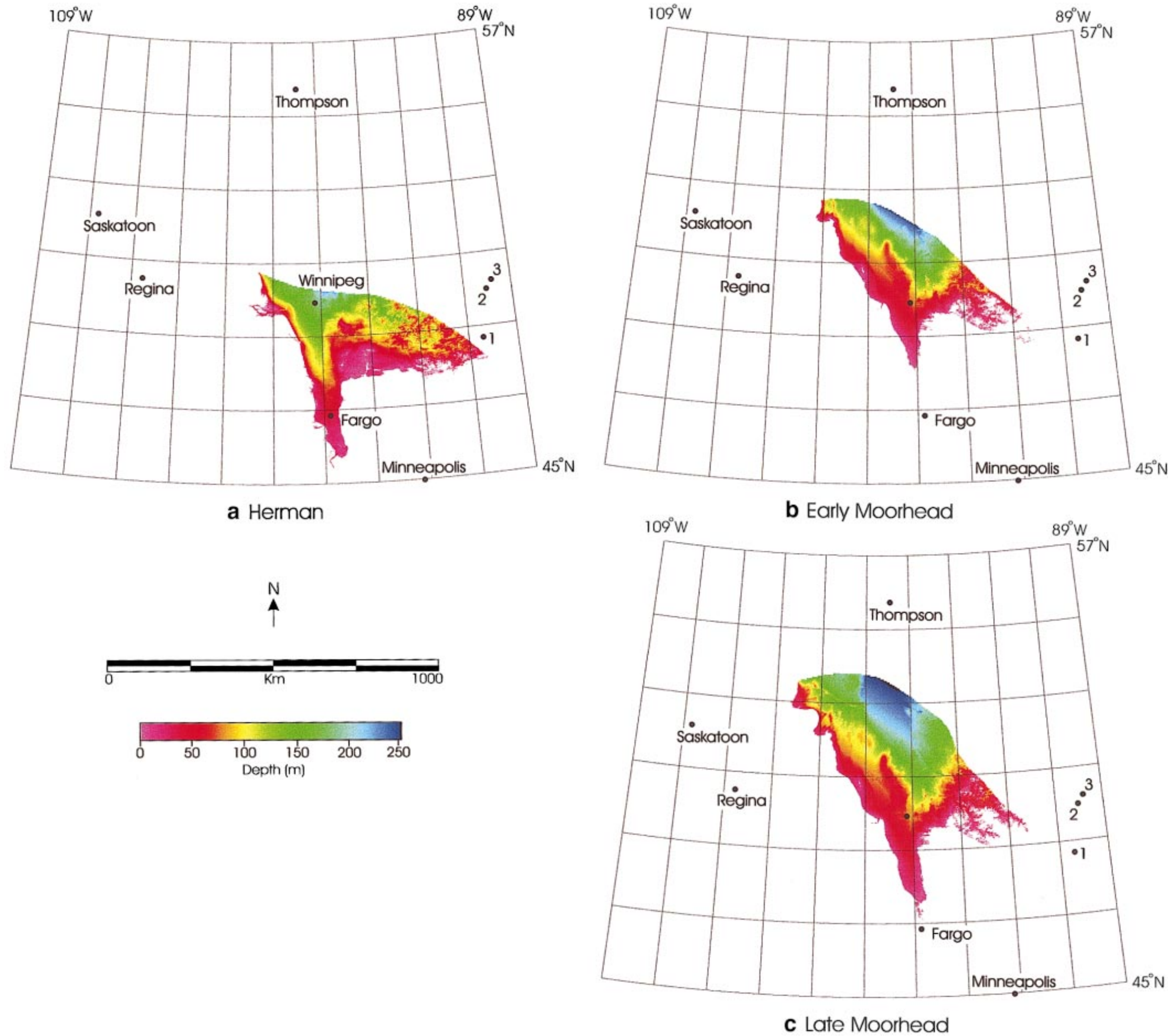

FIG. 3. Computer-modeled maps of bathymetry for seven stages of Lake Agassiz, spanning the period from 11,000 to 9300 ^{14}C yr B.P. Approximate ages for the stages, given in ^{14}C yr B.P., are: (a) 11,000, (b) 10,700, (c) 10,300, (d) 10,200, (e) 9900, (f) 9400, and (g) 9300. Maximum depth is 258 m (late Moorhead). Scale bar corresponds to 51°N. Grid cells are 2° by 2°. The locations of three eastern outlets mentioned in the text are labeled as follows: 1, the Kaministikwia route; 2, the Kaiashk route; 3, the Kopka route. Outlet positions are based on their approximate westernmost extents along the continental divide.

for each cell in the database, the relevant rebound-surface value from the corresponding modern-elevation value.

The bathymetric database produced for each lake stage was used to generate bathymetric maps and to estimate lake volume and area. The surface area of each stage was calculated from its bathymetric database by summing the areas of all cells whose elevations were lower than the shoreline elevation. The volume of each stage was calculated by summing the volume of each relevant cell's water column, which was given by the product of the cell's surface area and its depth below the lake surface.

In the generation and interpretation of results, it was assumed that the quantities of sediment deposition and erosion in the lake basin since the times of the modeled stages were relatively small compared to the volumes of the lake.

DATA COLLECTION

Database of Modern Elevations

The source of modern elevations used in this project was Version 1.0 of the GLOBE (Global Land One-Kilometer Base Elevation) database (GLOBE Task Team, 1999). The GLOBE

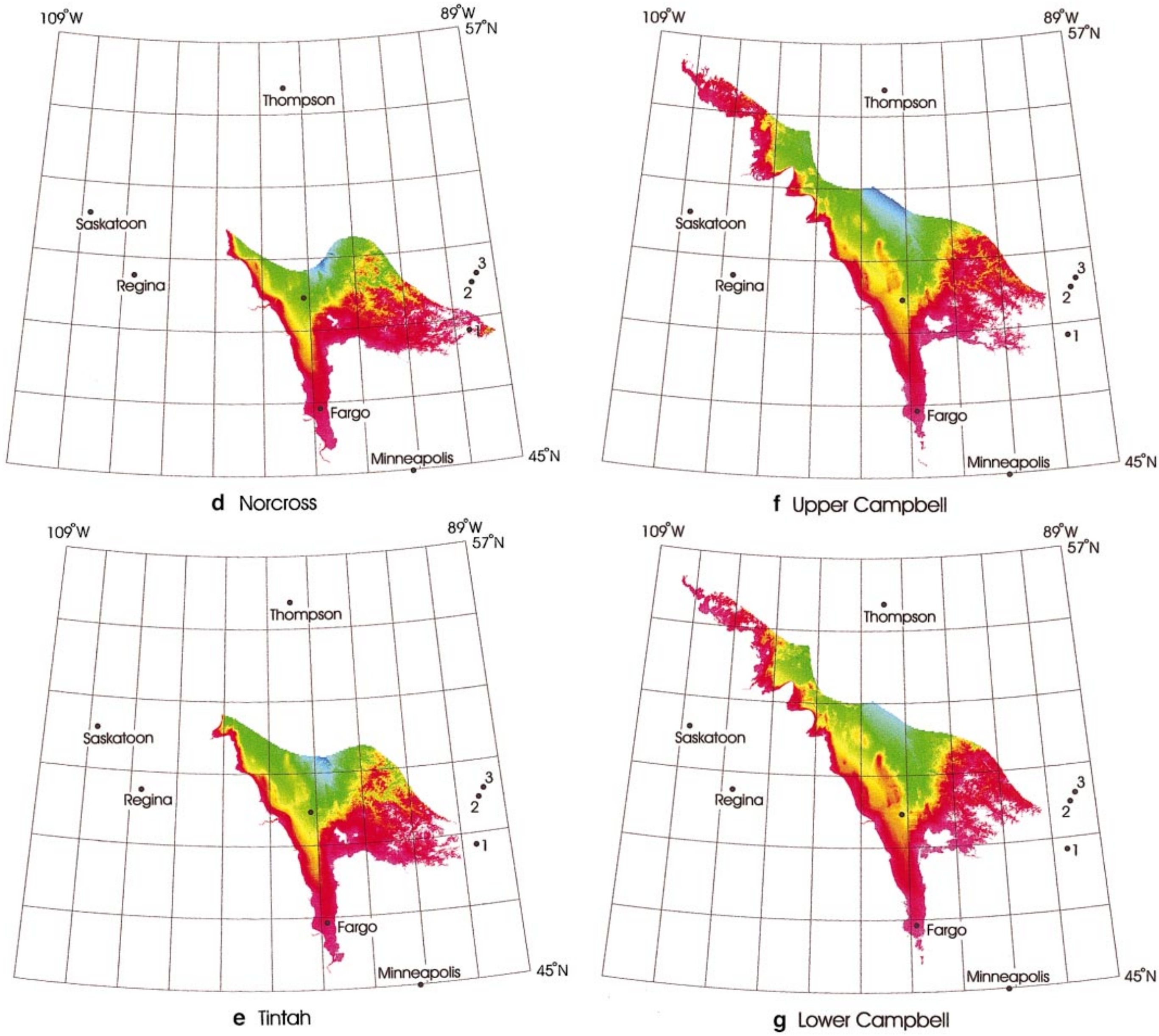

FIG. 3—*Continued*

database is an internationally designed, developed, and independently peer-reviewed global digital elevation model with a latitude–longitude grid spacing of 30 arc sec (Hastings and Dunbar, 1999). The GLOBE database does not include bathymetric data for lakes, including Lake Winnipeg. However, because these lakes are shallow (average depths are less than 15 m), this deficiency had little influence on Lake Agassiz bathymetric estimates and influenced lake volume calculations by less than 1.8%.

Rebound Surfaces

Rebound surfaces were spatially interpolated from point data taken from isobases and strandline rebound curves plotted by Thorleifson (1996; modified from Teller and Thorleifson, 1983); the two Moorhead rebound curves used here were modified from Thorleifson (1996) to make them no steeper than those of earlier stages. All rebound surfaces were interpolated using the "triangulated irregular network" (TIN) algorithm (Mann *et al.*, 1999) and were sampled at the same spatial resolution as the database of modern elevations. The number of isobase points used to define rebound surfaces ranged from 155 (early Moorhead) to 310 (Upper Campbell).

Shorelines were defined implicitly by the interpolated rebound surfaces themselves, whose values were based on isobase data that extended slightly beyond known lake shorelines. With the subtraction of a rebound surface from the modern topographic database, a shoreline was produced having a level of spatial detail limited only by the spatial resolution of the database of modern elevations.

Ice Margins

The geographical configuration of the northern ice margin of Lake Agassiz varied considerably during the period modeled here (e.g., Elson, 1967; Fenton *et al.*, 1983; Thorleifson, 1996). The ice margins used in this study were derived, with minor modifications, from those of Thorleifson (1996), which are based on interpretations of moraine, beach, and stratigraphic evidence. The Thorleifson (1996) ice margin for the lake when it was at the Upper Campbell level is located farther south than the Teller *et al.* (1983) margin at 9900 ^{14}C yr B.P., which was used as a basis for earlier area and volume estimations (Teller and Clayton, 1983; Mann *et al.*, 1999); thus, the lake parameters presented below for the Upper Campbell stage are less than those estimated in the earlier studies. The faces of all ice margins were treated as vertical.

RESULTS

Bathymetric maps for seven stages of Lake Agassiz in the period 11,000 to 9300 ^{14}C yr B.P. (ca. 13,000 to 10,300 cal yr B.P. (Stuiver and Reimer, 1993)) are given in Figure 3; bounding longitudes for these maps are 89° and 109°W, and bounding latitudes are 45° and 57°N. Corresponding lake parameters (area, volume, and maximum depth) are given in Table 1.

TABLE 1
Area, Volume, and Maximum-Depth Values Determined from the Maps of Bathymetry Given in Figure 3

Lake stage	Area (km^2)	Volume (km^3)	Maximum depth (m)
Herman	134,000	10,900	231
Early Moorhead	117,000	10,800	247
Late Moorhead	185,000	19,700	258
Norcross	166,000	13,300	243
Tintah	184,000	15,700	233
Upper Campbell	263,000	22,700	233
Lower Campbell	240,000	19,100	214

It can be seen from Table 1 that, of the seven stages modeled here, the early Moorhead stage was characterized by the smallest lake area and volume, at approximately 117,000 km^2 and 10,800 km^3, respectively (Lake Agassiz may have been much smaller prior to this stage, immediately after the drop in lake level from the Herman strandline; see below). The greatest lake area and volume are associated with the Upper Campbell beach, at about 263,000 km^2 and 22,700 km^3, respectively. The Upper Campbell values differ from those presented by Mann *et al.* (1999) as a consequence of the more southerly ice margin used here. As a check of model consistency, the calculations of the Upper Campbell stage were made using the techniques described in this paper, but with the more northerly ice margin; this produced area and volume estimations (340,000 km^2 and 39,500 km^3, respectively) that closely match those produced in the earlier modeling exercise. The Herman values presented in Table 1 differ from those presented by Mann *et al.* (1999) as a consequence of the updated isobase data used here.

Maximum lake depth did not vary substantially among the seven modeled stages, with the maximum depths of all stages falling between 214 and 258 m (Table 1). The deepest stage modeled here was the late Moorhead, with a maximum depth of 258 m. It is a consequence of this relatively great depth that the lake at the late Moorhead level had approximately the same volume as that at the Lower Campbell level, whose greater areal extent was offset by its relatively shallow maximum depth of 214 m.

In reviewing the results presented in Table 1, it must be kept in mind that all lake parameters, especially lake volume, are sensitive to the chosen positions and geographic configurations of the ice margins; areas near the ice margin are generally the deepest, and therefore have the greatest influence on volume. For example, a southward shift of the ice margin associated with the Upper Campbell beach by only 10 arc min of latitude will reduce the estimated volume of this lake stage by 10%.

CATASTROPHIC EVENTS

During the period modeled in this study, there were major hydrological events that may have occurred over short intervals of time (Clayton 1983; Fenton *et al.*, 1983; Teller and Thorleifson, 1983; Teller, 1990b; Thorleifson, 1996). Two notable events were (1) the drop in lake level after the Herman strandline formed, as a consequence of the abandonment of the southern outlet with the opening of eastern lake drainage to Thunder Bay through the Kaministikwia valley (ca. 10,900 ^{14}C yr BP, or ca. 12,900 cal yr B.P. (Stuiver and Reimer, 1993)) (Clayton, 1983; Teller and Thorleifson 1983; Thorleifson, 1996), and (2) the drop in lake level from the Upper Campbell beach to the level of the Kaiashk outlet (ca. 9400 ^{14}C yr B.P., or ca. 10,400 cal yr B.P.) (Teller and Thorleifson, 1983; Teller *et al.*, in press).

The configurations of the ice margin and rebound surface immediately following the post-Herman transition of drainage to the east are not well established, but they are likely to have been intermediate between those of the Herman and early Moorhead stages. An estimate of the minimum amount of water that was released with the opening of eastern drainage after the Herman beaches formed was calculated by using the following simplifying assumptions: (1) that water level during this event fell 110 m from the Herman beach to the approximate level of Thorleifson's (1996) early Moorhead level and (2) that this drop occurred with an ice-margin and isobase configuration identical to that used here to model the Herman stage (Fig. 3a). Such a drop in the Herman lake level reduces the lake to a volume of about 1375 km^3, indicating that, under this scenario, a volume of approximately 9500 km^3 of water (representing 87% of the lake's volume at the Herman stage) would have been discharged into the Superior basin during this event. This figure is considered a minimum value for the actual event, since more northerly Herman ice margins (i.e., margins that more closely resemble that of the early Moorhead stage, depicted in Fig. 3b) produce greater estimates of volume change, even though additional crustal rebound should accompany such ice margins. Under this simplified scenario, the area of Lake Agassiz would have declined to approximately 37,000 km^2 immediately after the Herman stage; subsequent ice-margin retreat and increases in lake volume would have led to the early Moorhead stage given in Table 1 and Figure 3b. The length of time over which this abrupt outflow occurred is not known, but it probably was a few decades or less.

The bathymetric and ice-margin configurations modeled here for the lake at the Upper Campbell beach (Fig. 3f) were used to produce an estimate of the volume of water that was released when the eastern (Kaiashk) outlets were accessed by Lake Agassiz at the end of this high-level stage. If the channels of the Kaiashk outlet were the only route used by overflow during this time, the lake level must have dropped by about 10 to 30 m from the level of the Upper Campbell beach (Thorleifson, 1996, Fig. 28). Using the computer model, it was calculated that a drop of 10 to 30 m would have reduced the volume of the lake by 11 to 31%, releasing approximately 2500 to 7000 km^3 of water into the Superior basin. A significantly greater volume of water would have been released if lower (more northerly) outlets such as those in the Kopka system were accessed (Fig. 1). The length of time over which this drawdown in lake level occurred was estimated by Teller and Thorleifson (1983) to have been only about two years. Figure 3g shows the bathymetry of the lake after it had transgressed to the Lower Campbell level.

SUMMARY AND CONCLUSIONS

A computer model was used to generate bathymetric maps for seven stages in the early history of Lake Agassiz, using a digital database of modern-elevation data and adjusting these elevations for isostatic rebound. The area and volume of the lake ranged considerably between about 11,000 and 9300 ^{14}C yr BP, reaching maxima of at least 260,000 km^2 and 22,700 km^3, respectively, during the Upper Campbell stage (ca. 9400 ^{14}C yr B.P.). Using the computer model, investigations of two significant drops in lake level suggest that a minimum volume of about 9500 km^3 of water was released at the end of the Herman stage (ca. 10,900 ^{14}C yr B.P.), and a volume of at least 2500 km^3 of water was released when the lake fell from the Upper Campbell level (ca. 9400 ^{14}C yr B.P.).

The volumetric changes in overflow from the lake that are modeled in this paper were superposed on the baseline outflow from Lake Agassiz, which Licciardi *et al.* (1999) calculated for the earliest phase of the lake to have been about 0.05 Sv (1577 $km^3\ yr^{-1}$; 1 sverdrup = $10^6\ m^3\ s^{-1}$). This baseline outflow was abruptly shifted into the North Atlantic Ocean via the Great Lakes and the Gulf of St. Lawrence (Fig. 2) about 10,900 ^{14}C yr B.P., when an eastern outlet of Lake Agassiz became free of ice. Added to this flow was the release of water from the lake as a result of the decline in level from the Herman strandline; this may have added another 9500 km^3 to the total, over a period of several decades or less. If the lake lowering occurred as a catastrophic event in only one year, the flux of freshwater into the North Atlantic about 10,900 ^{14}C yr B.P. would have increased by about 0.35 Sv for a period of one year. If the event occurred over 10 years, the flux increase would have averaged about 0.08 Sv (with 0.03 Sv contributed by the drop in lake level), whereas if the event occurred over 100 years, the flux increase would have averaged about 0.053 Sv over this period (with 0.003 Sv contributed by the drop in lake level).

The abrupt drawdown of Lake Agassiz about 9400 ^{14}C yr B.P., when the Upper Campbell beach was abandoned, resulted in an estimated 2500–7000 km^3 of water being released into the Great Lakes. This outflow would have been added to the 0.04 Sv baseline flow from Lake Agassiz (Licciardi *et al.*, 1999). Again, the drawdown of the lake at 9400 ^{14}C yr B.P. was associated with a shift in the routing of overflow into the North Atlantic Ocean. Thus, if all the water was released in only one year, the flux of freshwater into the North Atlantic would have increased by 0.12–0.26 Sv for a year. If the event occurred over a period of 10 years, the flux increase would have averaged about 0.048 to 0.062 Sv over this period (with 0.008 to 0.022 Sv contributed by the drop in lake level).

Some ocean models (e.g., Tziperman, 1997; Fanning and Weaver, 1997) show changes in thermohaline circulation in the North Atlantic that may have occurred as a result of "pulses" of freshwater superposed on a system that was made more sensitive by earlier meltwater runoff. Such changes could have led to a reduction in North Atlantic Deep Water (NADW) formation and, in turn, to climate change. Models suggest that widely distributed freshwater fluxes into the North Atlantic of about 0.16 to 0.32 Sv over a period of about 4 years (Rahmstorf, 1994, 1995a), or of about 0.06 Sv over a period of hundreds of years (Rahmstorf, 1995b), are sufficient to lead to a collapse of thermohaline circulation in the Atlantic Ocean. Furthermore, highly localized North Atlantic freshwater perturbations of about 0.006 Sv over 10 years can initiate chain reactions that result in completely different ocean convection patterns (Rahmstorf, 1995b), although the ultimate effects of such perturbations are a function of their geographic positions (Fanning and Weaver, 1997).

Based on these models, it is unclear if the magnitudes estimated here for the post-Herman and post-Upper Campbell drainage events were sufficient to influence NADW production; these events may not have produced freshwater fluxes in suitable locations or over sufficiently long periods to be significant. Interestingly, the 10,900 ^{14}C yr B.P. burst of outflow at the termination of the Herman stage approximately coincides with the beginning of Younger Dryas cooling, which has been suggested as having been triggered by the outflow of waters from Lake Agassiz (Broecker *et al.*, 1988, 1989, 1990).

Of course, the abrupt releases of Lake Agassiz waters also had an impact on the continental record. Large spillways, coarse fluvial sediments, and lacustrine sediment sequences in many regions downstream from Lake Agassiz bear testament to these events (e.g., Teller and Thorleifson, 1983; Lewis *et al.*, 1994; Fisher and Smith, 1994). Furthermore, the large spatial changes in the lake must have played a role in regional climate (Teller, 1987; Hu *et al.*, 1997; Hostetler *et al.*, 2000) and lake sedimentation.

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