Journal of Paleolimnology (2005) 33: 483–498
DOI 10.1007/s10933-005-0813-1

Modeling the history of Lake of the Woods since 11,000 cal yr B.P. using GIS

Zhirong Yang[1,2,*] and James T. Teller[1]
[1]*Department of Geological Sciences, University of Manitoba, Winnipeg, Manitoba, Canada R3T 2N2;* [2]*Geography Department, Hunan Normal University, Changsha, Hunan, P.R. of China 410006; *Author for correspondence (e-mail: yangz0@cc.umanitoba.ca)*

Received 3 August 2004; accepted in revised form 9 January 2005

Key words: Climate change, GIS, Holocene lake evolution, Isostatic rebound, Lake Agassiz, Lake of the Woods

Abstract

Because of differential isostatic rebound, many lakes in Canada have continued to change their extent and depth since retreat of the Laurentide Ice Sheet. Using GIS techniques, the changing configuration and bathymetry of Lake of the Woods in Ontario, Manitoba, and Minnesota were reconstructed for 12 points in time, beginning at 11,000 cal yr B.P. (~9.6 ^{14}C ka B.P.), and were also projected 500 years into the future, based on the assumption that Lake of the Woods continued to have a positive hydrological budget throughout the Holocene. This modeling was done by first compiling a bathymetric database and merging that with subaerial data from the Shuttle Radar Topography Mission (SRTM). This DEM file was then adjusted by: (1) isobase data derived from Lake Agassiz beaches prior to 9000 cal yr B.P. (~8.1 ^{14}C ka B.P.) and (2) modeled isostatic rebound trend analysis after 9000 cal yr B.P. Just after the end of the Lake Agassiz phase of Lake of the Woods, only the northernmost part of the basin contained water. Differential rebound has resulted in increasing water depth. In the first 3000 years of independence from Lake Agassiz, the lake transgressed > 50 km to the south, expanding its area from 858 to 2857 km^2, and more than doubling in volume. Continued differential rebound after 6000 cal yr B.P. (~5.2 ^{14}C ka B.P.) has further expanded the lake, although today it is deepening by only a few cm per century at the southern end. In addition, climate change in the Holocene probably played a role in lake level fluctuations. Based on our calculation of a modern hydrological budget for Lake of the Woods, reducing runoff and precipitation by 65% and increasing evaporation from the lake by 40% would end overflow and cause the level of the lake to fall below the outlets at Kenora. Because this climate change is comparable to that recorded during the mid-Holocene warming across the region, it is likely that the area covered by the lake at this time would have been less than that determined from differential isostatic rebound alone.

Introduction

Changes in the extent and volume of lakes during and after retreat of the Laurentide Ice Sheet (LIS) have been important factors in shaping the landscape of North America, and have contributed to changes in regional climate (Teller 1987; Hu et al. 1997; Hostetler et al. 2000), the global ocean-climate system (e.g., Broecker et al. 1989; Barber et al. 1999; Licciardi et al. 1999; Clark et al. 2001), early human history (Pettipas and Buchner 1983; Buchner and Pettipas 1990; Boyd et al. 2003), and even regional economic development.

Lake Agassiz was the largest lake in North America during the last deglaciation. It formed along the retreating margin of the Laurentide Ice Sheet (LIS), which dammed northward-draining rivers flowing to Hudson Bay and the Arctic Ocean, impounding the drainage of more than 2 million km^2 (Teller and Clayton 1983). The lake's size, volume, areal distribution, and history were related to the location of the ice margin, the elevation and location of its overflow channels, and differential isostatic rebound, all of which changed through time (Teller 2004).

Lake of the Woods (48°50′16″– 49°45′37″ N, 93°49′56″–95°19′26″ W) is a large remnant of glacial Lake Agassiz. Visited by French explorers in 1688 (Macins 1972), Lake of the Woods became an important crossroads for fur traders and occasional explorers and missionaries. The fish, wildlife, navigation on the lake, and the natural beauty of the area, attract many tourists to the area, and its overflow provides hydroelectric power and a favorable setting for various industries, including pulp and paper. In addition, Shoal Lake, which is part of the Lake of the Woods system, is the water supply for the city of Winnipeg, which flows west by pipeline from that lake (Figure 1). Thus, changes in lake depth and extent may impact the environment, tourism, industry and water supply, and will be important for water and environmental management in the future. Importantly, an understanding of these changes is essential for interpreting the sedimentary sequence in the basin and its paleohydrological and paleoclimatic record.

The goal of this research is to show how the extent and bathymetry of Lake of the Woods has changed over the past 11,000 cal years, to identify the spatial relationship with several other lakes in the region, and to predict the change in Lake of the Woods for the next 2–5 centuries. Using GIS modeling techniques, we present twelve stages of lake evolution beginning 11,000 cal yr B.P. (~9.6 ^{14}C ka B.P.). The early three stages of lake evolution (11,000, 10,500, and 10,000 cal yr B.P.; ~9.6, ~9.3, and ~8.9 ^{14}C ka B.P., respectively) were dominated by Lake Agassiz, while the last nine stages of Lake of the Woods (9000, 8000, 7000, 6000, 5000, 4000, 3000, 2000, and 1000 cal yr B.P.; ~8.1, ~7.2, ~6.1, ~5.2, ~4.4, ~3.7, ~2.9, ~2.0, and ~1.1 ^{14}C ka B.P., respectively) occurred after the lake had become independent from Agassiz.

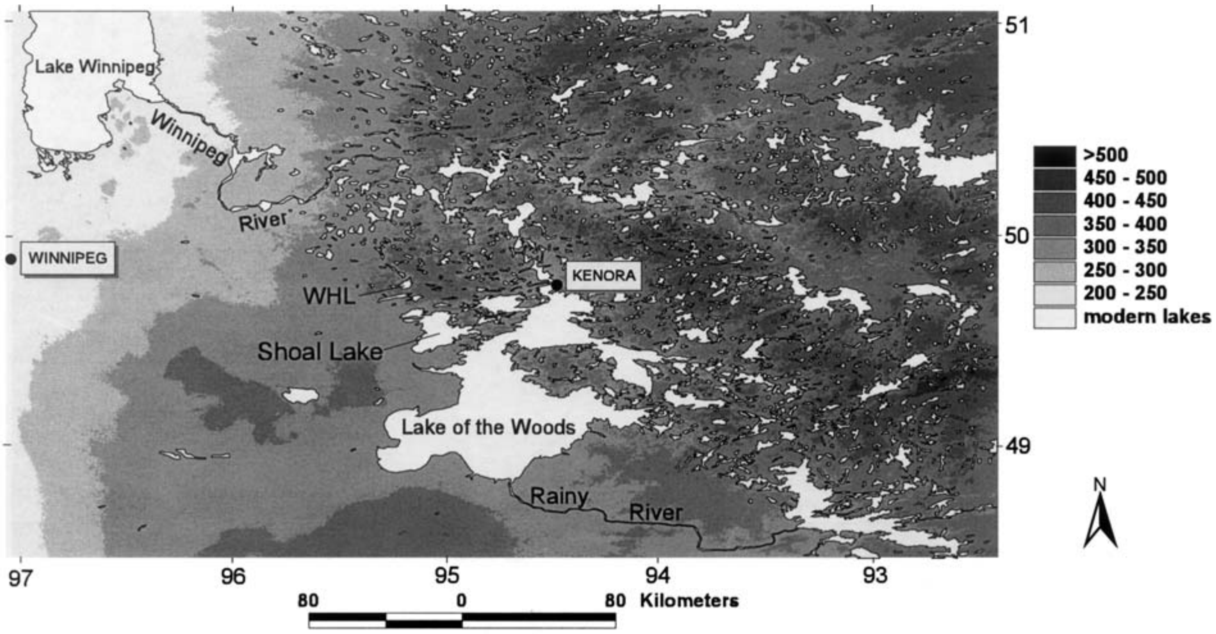


Figure 1. Modern digital elevation model (DEM) of the region studied in southeastern Manitoba, northwestern Ontario and northern Minnesota. Area is shown in Figure 2. Elevation is in meters. WHL = West Hawk Lake.

Site description and modern outlets

Lake of the Woods is an international water body, lying along the Canadian–U.S. boundary of northwestern Ontario, southeastern Manitoba, and northern Minnesota. Irregular in shape, it is 110 km long and up to 95 km wide. It has an area of 4472 km^2, of which 1663 km^2 are in the U.S. It has an estimated 40,000 km of shoreline and more than 14,000 islands (Encyclopedia Britannica 1999). The average depth is 8 m, but some of the northernmost bays are more than 45 m deep. The principal tributary to Lake of the Woods is Rainy River on the international boundary, which provides 75% of the lake's water (Macins 1972). Its outlet river is the Winnipeg River, which carries overflow into Lake Winnipeg; that lake overflows into the Nelson River, which flows to Hudson Bay (Figures 1 and 2).

Lake of the Woods originally had two natural outlets, both at the northern end near Kenora. Known as the Eastern outlet and the Western outlet, they have been modified, but continue to carry overflow to the Winnipeg River at an average rate of 392.4 m^3/s, which is more than 91% of the total overflow from the Lake of the Woods (Environment Canada 2004). The Western outlet is the largest, today carrying nearly three-quarters of the total outflow from the lake. The first attempt to control this outlet was made in 1887 with the building of the Rollerway Dam, but this was replaced in 1890 by Norman Dam; the Norman Powerhouse was built in 1925 to meet the power requirements of the paper mill (LWCB 2002). Partial control of the Eastern outlet was achieved in 1892 with the construction of a plant for municipal power supply for the town of Kenora, and full control of this outlet occurred in

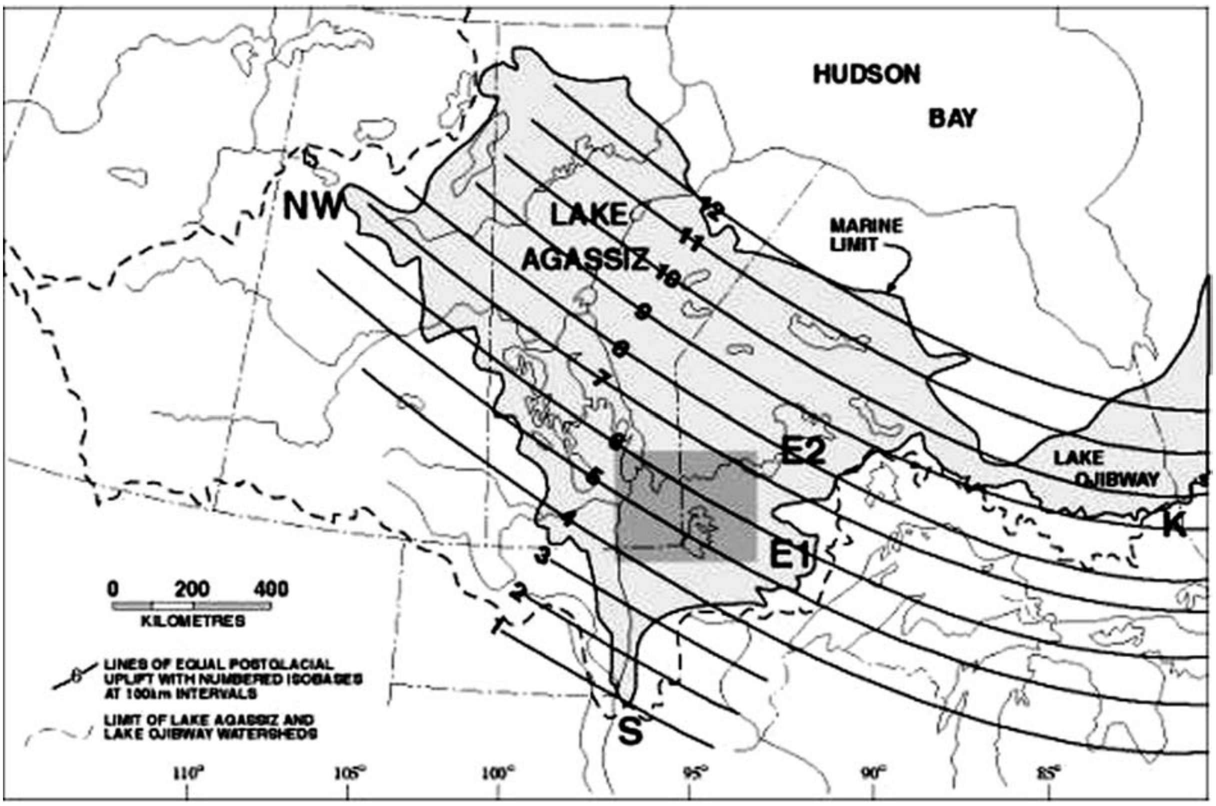


Figure 2. Total area covered by waters of glacial Lake Agassiz (light gray shaded area), showing the main outlets used at various times during its history (after Teller and Thorleifson 1983, Figure 1). Eastward outflow into the Great Lakes occurred initially through routes directly into Lake Superior (E1), whereas later overflow passed through a series of eastern outlets (E2) into Lake Nipigon before reaching Lake Superior; NW = northwestern outlet; S = southern outlet; K = Kinojévis outlet. Lines are isobases representing contours of equal isostatic rebound spaced at 100 km intervals. The area of this research is shown by dark gray shaded area.

1906 when a larger plant was built. All water flowing out the Eastern outlet is used for power generation or mill processing (LWCB 2002). The aqueduct from Shoal Lake to the City of Winnipeg (Figure 1) became one of the outlets of Lake of the Woods when it was completed in 1919. The flow through the aqueduct to Winnipeg represents less than 1% of the total outflow from the lake.

The construction of dams and the control of outflows from Lake of the Woods raised the average lake level by about 1.8 m above the level of the natural Western and Eastern outlets. In order to produce the highest uniform outflow, Lake of the Woods Control Board has made an effort to keep Lake of the Woods water level between 321.87 m and 323.47 m (LWCB 2002).

Methods

As shown by previous research, GIS is an effective method for reconstructing the paleotopographic surface in this region and elsewhere in glaciated North America (Gareau et al. 1998; Mann et al. 1999; Leverington et al. 2000, 2002a, b; Matile et al. 2000; Lewis and Gareau 2001; Teller et al. 2002; Lewis and Thorleifson 2003; Teller and Leverington 2004). Paleotopographic reconstructions are essential for the interpretation of past lake environments, and for the assessment and modeling of past geological, hydrological, and climatic processes.

Seven steps were followed in this research: (1) The modern Digital Elevation Model (DEM) database of the area around Lake of the Woods (Figure 1) was imported into ArcView GIS. (2) Using an empirical model of regional isostatic rebound (Lewis and Thorleifson 2003) and the elevations of glacial Lake Agassiz beaches (Teller and Thorleifson 1983), we calculated isostatic rebound values and generated isobase maps for 14 stages in the Lake of the Woods region. (3) After inputting isobase line data manually, we interpolated the isobase line data using the Triangulated Irregular Network (TIN) algorithm to get isostatic rebound surfaces for each stage at 500–1000 year intervals; in order to do spatial analysis, we converted each stage's isostatic rebound surface into grids (raster data). (4) From a bathymetric contour map of Lake of the Woods, we chose 6500 depth points, and input these point data into ArcView GIS. (5) Again using the TIN algorithm, we interpolated point data and plotted bathymetric surfaces of Lake of the Woods. (6) By subtracting the bathymetric surface from the original DEM, we generated a new DEM of Lake of the Woods area that included bathymetric information of the lake. (7) By subtracting the amount of differential isostatic rebound at each stage from its modern topographic value (the original DEM for first three stages, the new DEM of Lake of the Woods area for last nine stages), we reconstructed paleotopographic surfaces for twelve different points in time, and projected the depth and extent of the lake for two times in the future.

Because this region was submerged by glacial Lake Agassiz as soon as ice retreated from the area about 13,000 cal yr B.P. ($\sim$11,000 ^{14}C yr B.P.) (Teller et al. 2000) and, based on our modeling, because Lake of the Woods area was part of this lake for most of the subsequent 4000 years, reconstructions of the lake during this period are controlled by the history of rising and falling levels of this giant proglacial lake. In short, the depth and extent of water in the Lake of the Woods basin during this period was not controlled by the modern outlets at Kenora, but by factors controlling Lake Agassiz.

Data collection

Database of modern digital elevation model (DEM)

The DEM files for high-resolution terrain analysis used in this research for the first three stages come from USGS Seamless Data Distribution System, which is a source for Shuttle Radar Topography Mission (SRTM) terrain data for the United States and limited areas in other parts of the world. From this system, area selection is seamless and is not constrained by latitude or longitude boundaries, and data can be downloaded immediately free of charge. These data are available only in geodetic (latitude–longitude) coordinates and are offered as Arc-Grid format with high resolution (30 m).

The SRTM data used do not include bathymetric information for Lake of the Woods. Lake depth was digitized from modern bathymetric contour maps prepared by Garmin Map Source Company for the Freshwater Institute in Winnipeg, Department Fisheries and Oceans

Canada. The bathymetric surface was derived from these point data using a TIN interpolating model, and the new DEM of the Lake of the Woods area was calculated by subtracting the bathymetric surface from the original DEM.

Isostatic rebound data

Through time, the actual elevation of the regional surface has changed in response to differential isostatic rebound. Because rebound was greatest in the northeastern region, elevations there have become higher relative to those in the southwest (see Figure 2). As a result, both the specific elevation and the configuration of the topographic contour lines (and the lakes that are bounded by those contours) have changed.

In order to reconstruct the paleotopography of the region, and the changing configuration of Lake of the Woods through time, the modern SRTM topography and bathymetry must be adjusted for the isostatic rebound differential at any point in the past. Differential isostatic rebound in the Lake of the Woods area during the early stages of the lake is known from the curvature of Lake Agassiz beaches (Figure 3), while the isostatic rebound values for the late stages of the lake are calculated using an empirical model.

In order to compute changes in the gradient of the Red River valley of Manitoba over the past 9000 years, Lewis and Thorleifson (2003) developed an empirical model based on previous studies (Peltier 1994, 1998), in which postglacial isostatic rebound is quantitatively expressed as an exponential function of age:

$$\mathrm{RU} = A * (e^{t/\tau} - 1) \tag{1}$$

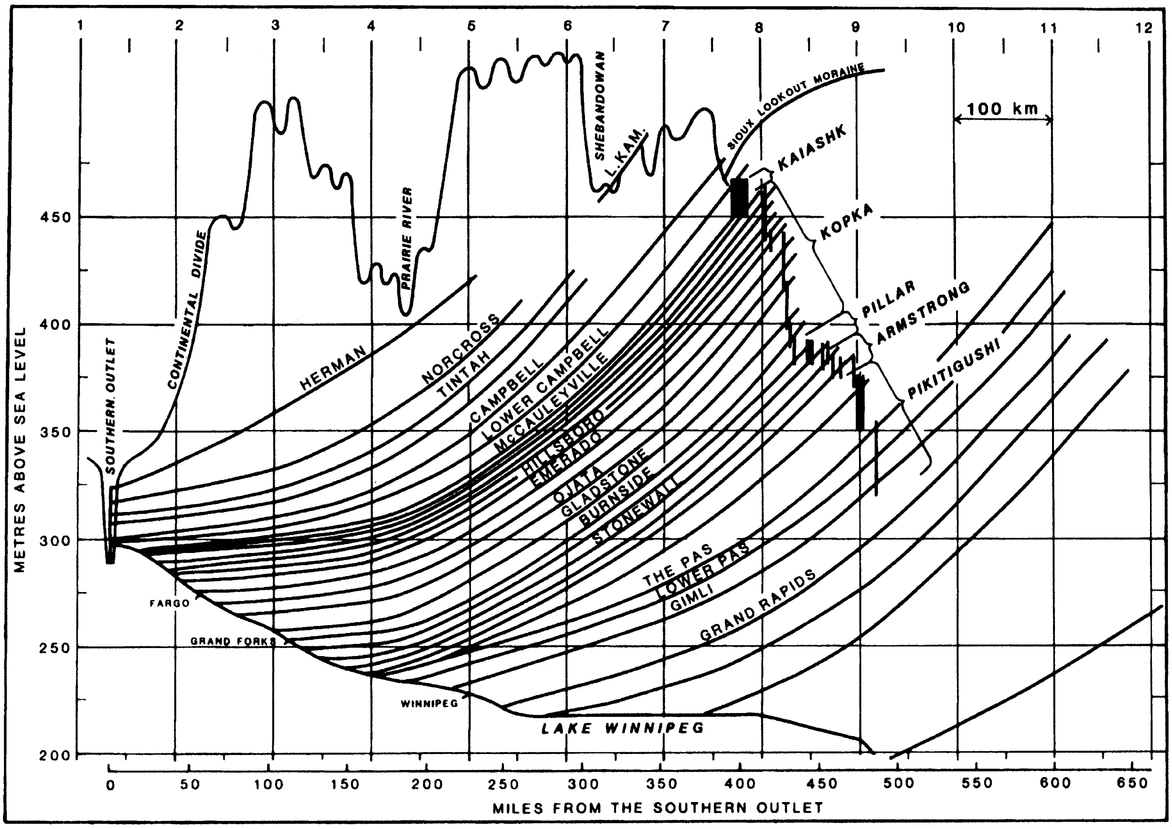


Figure 3. Modern elevation of major water planes of glacial Lake Agassiz, extending from the southern outlet toward the northeast perpendicular to isobase lines in Figure 2, reconstructed from beaches and wave-eroded shorelines (Teller and Thorleifson 1983, Figure 2). Vertical bars represent elevation of eastern outlets between Lake Agassiz and Great Lakes' system.

- *RU*: uplift in m at any point relative to Isobase 1.3 that passes through the southern outlet of Lake Agassiz (Figure 2)
- *A*: amplitude factor for uplift at a particular site
- τ: relaxation time for the uplift, which is roughly the time in years for half the remaining isostatic uplift to occur
- *t*: age in cal years (time elapsed since formation of a specific water level indicator)
- e: a constant (e≈2.71828)

The parameter τ of Equation (1) can be evaluated using dated shoreline elevations. For each pair of shorelines with known relative uplift (RU_1, RU_2) and age (t_1, t_2) at a site, τ was evaluated by:

$$RU_1 = A * (e^{t1/\tau} - 1) \quad (2)$$

$$RU_2 = A * (e^{t2/\tau} - 1) \quad (3)$$

Dividing Eq. (2) by Eq. (3), and eliminating the common factor A and rearranging terms, a single equation for unknown τ was formed:

$$RU_1 * (e^{t2/\tau}) - RU_2 * (e^{t1/\tau}) + RU_2 - RU_1 = 0 \quad (4)$$

The relaxation time τ in Eq. (4) was evaluated for all shoreline pairs at a site using Goal Seek in an Excel spreadsheet (Liengme 1997). Ten plausible values for τ of less than 10,000 years were obtained. A best value was selected by computing differential uplift at points that controlled the middle Holocene diversion of the Saskatchewan River and the transgression of Lake Winnipeg shore since AD 1600 (Lewis and Thorleifson 2003). Both events had been determined from independent investigations by McMartin (2000) and Nielsen (1998). A relaxation time of 3500 years, which gave the best agreement between computed and observed age of river diversion and rate of lake transgression, was selected by Lewis and Thorleifson (2003). This value compares favorably with relaxation times found in adjoining areas, 3426 years and 3399 years in southeastern Hudson Bay and James Bay, respectively (Peltier 1998), and 3700 ± 700 years in the Great Lakes region (Lewis and Gareau 2001). The value of $\tau = 3500$ years was used in subsequent computations of uplift in this research, because the area of this research is included in the area studied by Lewis and Thorleifson (2003).

The amplitude parameter *A* in Eq. (1) varies from place to place within the region depending on the magnitude of uplift. In some regions, if there is a shoreline of known age (*t*) and Relative Uplift (RU), *A* can be determined using Eq. (1), with $\tau = 3500$ years. In this research, elevations of the Lower Campbell beach ($t = 10{,}500$) were obtained at given isobases from the Lake Agassiz strandline diagram (Figure 3), and then, by subtracting the elevation of the Lower Campbell beach at isobase 1.3, Relative Uplift (RU) was calculated. In this example, the amplitude parameter *A* for a given isobase was calculated by:

$$A = RU/(e^{10500/3500} - 1) \quad (5)$$

Relative uplift of the research area at a given isobase through time was calculated by:

$$\text{RU at a given isobase} = A * (e^{t/3500} - 1) \quad (6)$$

Using this empirical model, we calculated isostatic rebound values at specific points in the research area through time. The results are shown in Table 1.

Outlet controls

Before Lake Agassiz receded from this region, water depths over Lake of the Woods were controlled by outlets from this giant glacial lake. Based on previous Lake Agassiz research (e.g., Teller and Thorleifson 1983; Teller and Leverington 2004), the outlet controls of the earliest stages of Lake of the Woods (11,000, 10500, and 10,000 cal yr B.P.; ~9.6, ~9.3, and ~8.9 ^{14}C ka B.P., respectively) were determined, and are shown in Table 2.

Once Lake Agassiz had regressed from the region, all remnant lakes were controlled by their own outlets. As previously noted, Lake of the Woods has two natural outlets, the Western and Eastern outlets at Kenora, which are adjacent to each other and at the same elevation; they carry more than 91% of the overflow to the Winnipeg River. The modern Western outlet of Lake of the Woods is shown in Figure 4. Assuming no erosion, the paleo-elevation of this outlet was calculated by subtracting isostatic rebound from that of the modern Western outlet (Table 3).

Table 1. Values of total postglacial isostatic rebound (in m) at specific isobases (i) since specific times, beginning 11,000 cal yr B.P. to 500 years in the future.

Isobase (i)	3.75	4.0	4.25	4.5	4.82	5.0	5.25	5.5	5.75	6.0	6.125	6.25	6.5	6.75	7.0	7.25	7.5	7.75	8.0
ELC	314.9	318.1	322.2	324.9	332	338.1	347.5	356.1	364.3	373.2	381.1	390	399.5	414.3	425	441.1	455	468	482.5
ELCSO	288.1	288.1	288.1	288.1	288.1	288.1	288.1	288.1	288.1	288.1	288.1	288.1	288.1	288.1	288.1	288.1	288.1	288.1	288.1
RULC	26.8	30	34.1	36.8	43.9	50	59.4	68	76.2	85.1	93	101.9	111.4	126.2	136.9	153	166.9	179.9	194.4
τ	3500	3500	3500	3500	3500	3500	3500	3500	3500	3500	3500	3500	3500	3500	3500	3500	3500	3500	3500
A	1.40	1.57	1.79	1.93	2.30	2.62	3.11	3.56	3.99	4.46	4.87	5.34	5.84	6.61	7.17	8.02	8.74	9.43	10.19
Cal	RU	RU	RU	RU	RU	RU	RU	RU	RU	RU	RU	RU	RU	RU	RU	RU	RU	RU	RU
yr B.P.	at i3.75	at i4.0	at i4.25	at i4.5	at i4.82	at i5.0	at i5.25	at i5.5	at i5.75	at i6.0	at i6.125	at i6.25	at i6.5	at i6.75	at i7.0	at i7.25	at i7.5	at i7.75	at i8.0
+500	0.22	0.24	0.27	0.30	0.35	0.40	0.48	0.55	0.61	0.68	0.75	0.82	0.90	1.02	1.10	1.23	1.34	1.45	1.56
+200	0.08	0.09	0.11	0.11	0.14	0.15	0.18	0.21	0.23	0.26	0.29	0.31	0.34	0.39	0.42	0.47	0.51	0.55	0.60
0	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
1000	0.46	0.52	0.59	0.64	0.76	0.87	1.03	1.18	1.32	1.47	1.61	1.77	1.93	2.19	2.37	2.65	2.89	3.12	3.37
2000	1.08	1.21	1.38	1.49	1.77	2.02	2.40	2.75	3.08	3.44	3.76	4.12	4.50	5.10	5.53	6.18	6.74	7.27	7.85
3000	1.90	2.13	2.42	2.62	3.12	3.55	4.22	4.83	5.42	6.05	6.62	7.24	7.92	8.97	9.73	10.87	11.86	12.79	13.82
4000	3.00	3.36	3.82	4.12	4.91	5.60	6.65	7.61	8.53	9.52	10.42	11.40	12.47	14.12	15.32	17.12	18.68	20.13	21.75
5000	4.46	4.99	5.67	6.12	7.30	8.31	9.87	11.30	12.67	14.15	15.48	16.94	18.52	20.98	22.76	25.43	27.74	29.91	32.32
6000	6.39	7.16	8.13	8.78	10.47	11.93	14.17	16.22	18.18	20.30	22.21	24.31	26.57	30.10	32.66	36.50	39.81	42.91	46.37
7000	8.97	10.04	11.42	12.32	14.70	16.74	19.88	22.76	25.51	28.49	31.17	34.11	37.29	42.25	45.83	51.22	55.87	60.22	65.08
8000	12.40	13.88	15.78	17.03	20.32	23.14	27.49	31.47	35.27	39.38	43.09	47.16	51.56	58.40	63.36	70.81	77.24	83.26	89.97
9000	16.97	19.00	21.59	23.30	27.80	31.66	37.61	43.06	48.25	53.88	58.96	64.52	70.54	79.91	86.68	96.88	105.68	113.91	123.09
10,000	23.05	25.80	29.32	31.65	37.75	43.00	51.08	58.47	65.53	73.18	80.07	87.62	95.79	108.52	117.72	131.56	143.52	154.70	167.16
10,500	26.80	30.00	34.10	36.80	43.90	50.00	59.40	68.00	76.20	85.10	93.11	101.90	111.40	126.20	136.90	153.00	166.90	179.90	194.40
11,000	31.13	34.85	39.61	42.75	51.00	58.08	69.00	78.99	88.52	98.85	108.16	118.37	129.40	146.59	159.03	177.73	193.87	208.97	225.82

Isobase (i) number relates to distance (*100 km) from isobase 1 at southern end of Lake Agassiz (Figure 2).
ELC = modern elevation of Lower Campbell beach at each isobase.
ELCSO = modern elevation of Lower Campbell beach at southern outlet of Lake Agassiz (isobase 1.3).
RULC = uplift of Lower Campbell beach at each isobase, as related to southern outlet (isobase 1.3) = ELC – ELCSO.
τ = relaxation time (cal yr B.P.)
A = amplitude factor at various isobases = RULC/ $(\exp(10500/\tau)-1)$. 10500 is the age in cal yr B.P. of Lower Campbell beach.
RU = relative uplift compared to isobase 1.3 (i1.3), which extends through the southern outlet of Lake Agassiz as shown in Figures 2 and 3 (e.g. i3.75, i4.0, etc.).

Table 2. Elevation of the outlet controlling Lake Agassiz at three different stages.

Cal yr B.P.	Outlet	Elevation (m)[a]
10,000	Eastern	270
10,500	Eastern	286
11,000	Northwestern	288

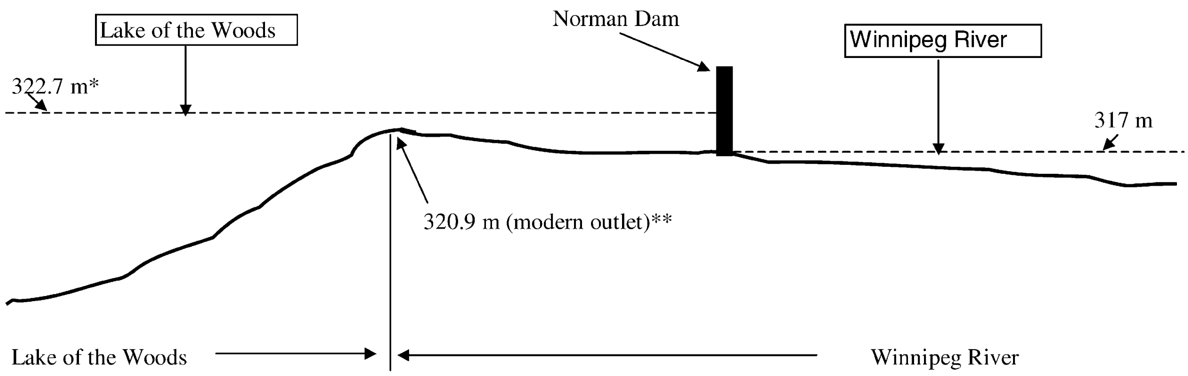
[a]Modern outlet elevation – extrapolated outlet rebound from beach of closest age = original elevation of outlet, adjusted for the age indicated.

Results and discussion of isostatic rebound history

During the early history of the Lake of the Woods basin, all modern lake basins in this region were part of the same large hydrological system, glacial Lake Agassiz. As the level of this giant lake declined, topographically higher basins, or those surrounded by relatively high topography, became independent. The lake area, volume, and maximum and mean depths, determined from the maps of Figures 5 and 6, are shown in Table 4. The early three stages of lake evolution in the area (11,000, 10,500, and 10,000 cal yr B.P.; ~9.6, ~9.3, and ~8.9 ^{14}C ka B.P., respectively) were dominated by Lake Agassiz (Figure 5a–c). As a result of the abrupt opening and closing of outlets, plus the interaction of differential isostatic rebound and the use of many different outlets, the level of Lake Agassiz was constantly changing, and its history is very complex at the century scale (Teller and Leverington 2004). Therefore, the maps of the stages at any given time are only snapshots of lake evolution. Low stages of Lake Agassiz occurred as lower outlets were deglaciated, leading to a draw down of lake level. Then, because of differential isostatic rebound of the outlet, lake level began to rise, until a new lower outlet was opened. Figure 5a is the bathymetric model of the area at 11,000 cal yr B.P. (~9.6 ^{14}C ka B.P.), when water over Lake of the Woods was deepest (~132 m).

After 11,000 cal yr B.P. (~9.6 ^{14}C ka B.P.), the lake area, depth, and volume all decreased step by step (Figures 5a–c and 6a) in response to the abrupt opening of new and lower Lake Agassiz outlets; each was followed by a rise in lake level, as differential isostatic rebound elevated that outlet (Teller 2001). Just before 9000 cal yr B.P. (~8.1 ^{14}C ka B.P.), Lake Agassiz withdrew from most of the Lake of the Woods region (Figure 6a).

Thus, Lake of the Woods became separate from Lake Agassiz and started its own evolution by 9000 cal yr B.P. (~8.1 ^{14}C ka B.P.) (Figure 6a). Initially, Lake of the Woods only covered a small area (~858 km^2), the smallest area in its history, which was less than 20% of its modern area (Table 4). Its volume was only 7.3 km^3, which was also the smallest it has been since becoming independent of Lake Agassiz, assuming that waters remained deep enough to overflow through the Kenora outlets throughout the Holocene. However, at this time, because the northern end of the basin was isostatically depressed, the mean depth of this small stage (8.5 m) was greater than at any time since (Figure 6a–i). During subsequent stages, Lake of the Woods expanded southward



*The normal lake level of Lake of the Woods = (Normal Maximum 323.47 m + Normal Minimum 321.87 m)/2 =322.7 m
**The elevation of modern outlet = the normal lake level 322.7 m - water depth of outlet 1.8 m = 320.9 m

Figure 4. A schematic profile of the modern Western outlet of Lake of the Woods and related elevations (LWCB 2002)

Table 3. Elevation of outlet controlling Lake of the Woods through time after separation from Lake Agassiz.

Cal yr B.P.	Modern elevation−rebound value[a]	Outlet Elevation (m)
1000	320.9−1.61	319.29
2000	320.9−3.76	317.14
3000	320.9−6.62	314.28
4000	320.9−10.42	310.48
5000	320.9−15.48	305.42
6000	320.9−22.21	298.69
7000	320.9−31.17	289.73
8000	320.9−43.09	277.81
9000	320.9−58.96	261.94

[a]Rebound values are taken from Table 1 at isobase 6.125, which extends through the outlet of Lake of the Woods.

and its volume increased (Table 4). Its maximum and mean depth has varied irregularly (see Table 4), reflecting the irregular topography over which the lake transgressed.

Some lake basins remained linked even after Lake Agassiz fell below the basin margin, whereas others merged, as differential isostatic rebound led to expanding southern margins that then enveloped other lakes. For example, West Hawk Lake (WHL, Figure 5a–c) lies in a meteorite impact crater which has a thick (70 m) sequence of Quaternary sediment (Boyd et al. 2002). From our paleo-bathymetric modeling, we know that this lake emerged from Lake Agassiz about 10,000 cal yr B.P. (~8.9 ^{14}C ka B.P.) and also separated from Lake of the Woods at about that time (Figure 5c). Thus, the histories of these two small lakes and their sedimentary records would have been similar until then, and would have been closely related to that of Lake Agassiz. However, it was another 1000 years before Lake of the Woods became independent from Lake Agassiz. Even after that, the record of an independent Lake of the Woods did not begin everywhere in the area covered by the modern lake, and there would have been a hiatus in lacustrine deposition until differential rebound forced the waters to transgress over the exposed southern floor of the basin (Figure 6b–i). In general, the Holocene record in the southern part of the basin spans a shorter and younger period of time than in the northern part. However, as can be seen in Figures 6b–f, waters expanded into the southern part of the Lake of the Woods basin by about 8000 cal yr B.P. (~7.2 ^{14}C ka B.P.) but were largely separated by higher topography from those in the northern part until about 3000 cal yr B.P. (~2.9 ^{14}C ka B.P.).

Lake of the Woods will continue transgressing toward the south. Isostatic rebound values for 200 and 500 years into the future were calculated and are shown in Table 1. From this, we can forecast that lake level will increase about 14 cm and 35 cm, respectively, and that the low land around the southern part of Lake of the Woods will be inundated. The rate and extent of inundation depends on the slope of the southern shore, which needs to be studied further. Rising lake level will result in shore erosion and lake shore property loss.

The impact of climate change on lake level

The history of Lake of the Woods described in the previous section was based on an isostatic rebound model with the assumption that there has always been a positive hydrological budget, with continuing overflow through its outlets. However, there is evidence for mid-Holocene drought in the region, and this would have been an additional influence on water levels of Lake of the Woods. For example, drier conditions are recorded in a small lake 40 km to the west between 6200 and 4000 ^{14}C yr B.P. (Teller et al. 2000). Studies of Lake Winnipeg, 240 km to the northwest (Figure 1), indicate that mid-Holocene drought (7500–4000 ^{14}C yr B.P.) dramatically affected its level, and runoff from its ca. 1 million km^2 drainage basin was so reduced that the lake stopped overflowing (Lewis et al. 2001). Lake of the Woods is part of that drainage basin, with its overflow today supplying about half of the total runoff of the Winnipeg River, which is ~20% of the total runoff to Lake Winnipeg. In adjacent Lake Manitoba, similar low lake levels during the mid-Holocene are partly attributed to decreased precipitation/evaporation ratios (e.g., Teller and Last 1982; Last and Teller 2002), and farther west in Saskatchewan some lake basins became dry at this time (e.g., Porter et al. 1999). Even water levels in the Great Lakes appear to have been lower during the middle Holocene (Blasco 2001; Lewis et al. 2002). Multi-proxy studies of lake sediments indicate mid-Holocene drought in areas to the south such as in Minnesota (e.g., Elk Lake, Dean 1993; Steel Lake,

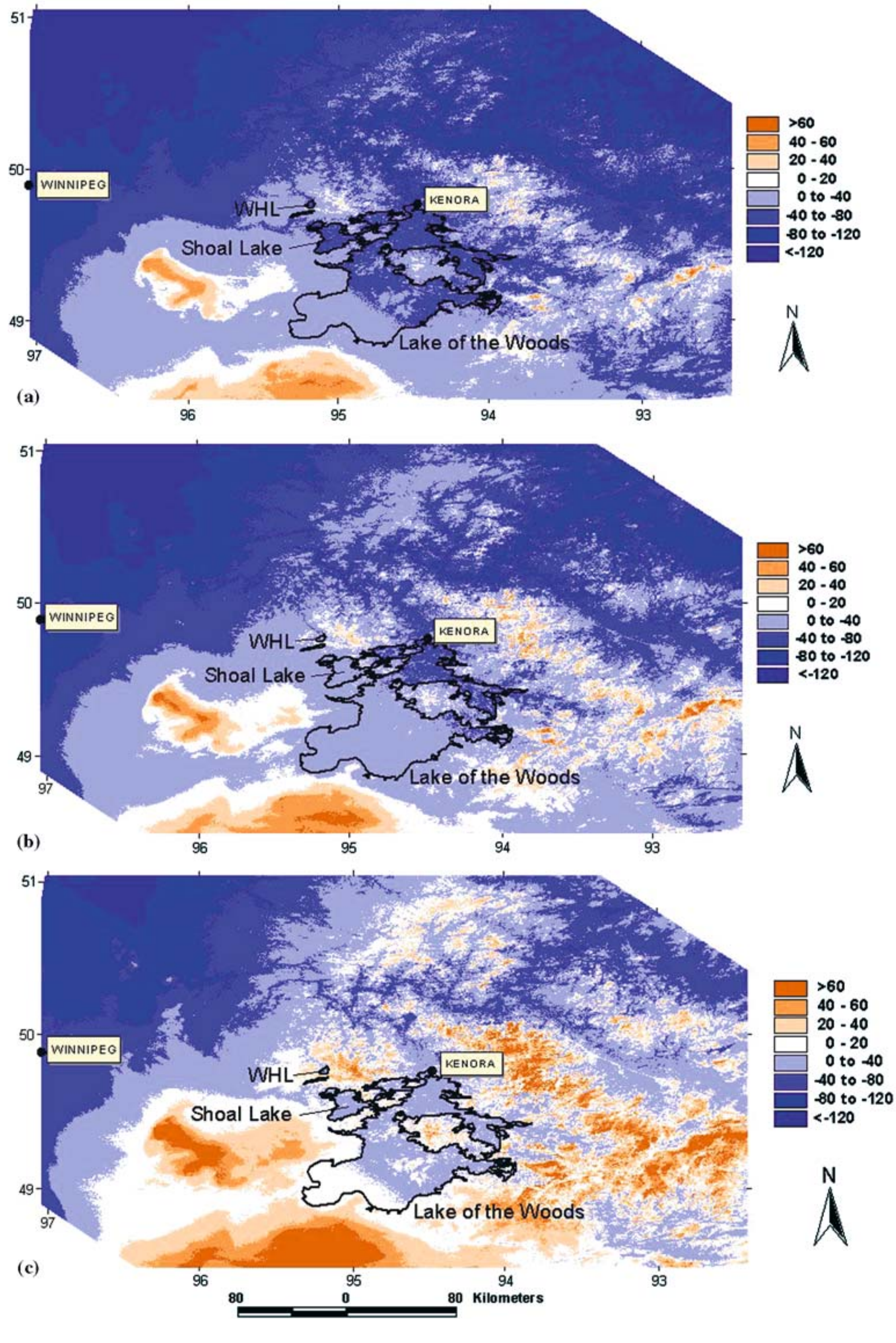


Figure 5. Extent and depth (in meters) of waters in Lake of the Woods region (blue shades) and elevation (in meters) of surrounding land above lake level (white to orange shades) at (a) 11,000, (b) 10,500, and (c) 10,000 cal yr B.P. Elevations of the lake above modern sea in three different stages are 288, 286, and 270 m respectively. WHL = West Hawk Lake.

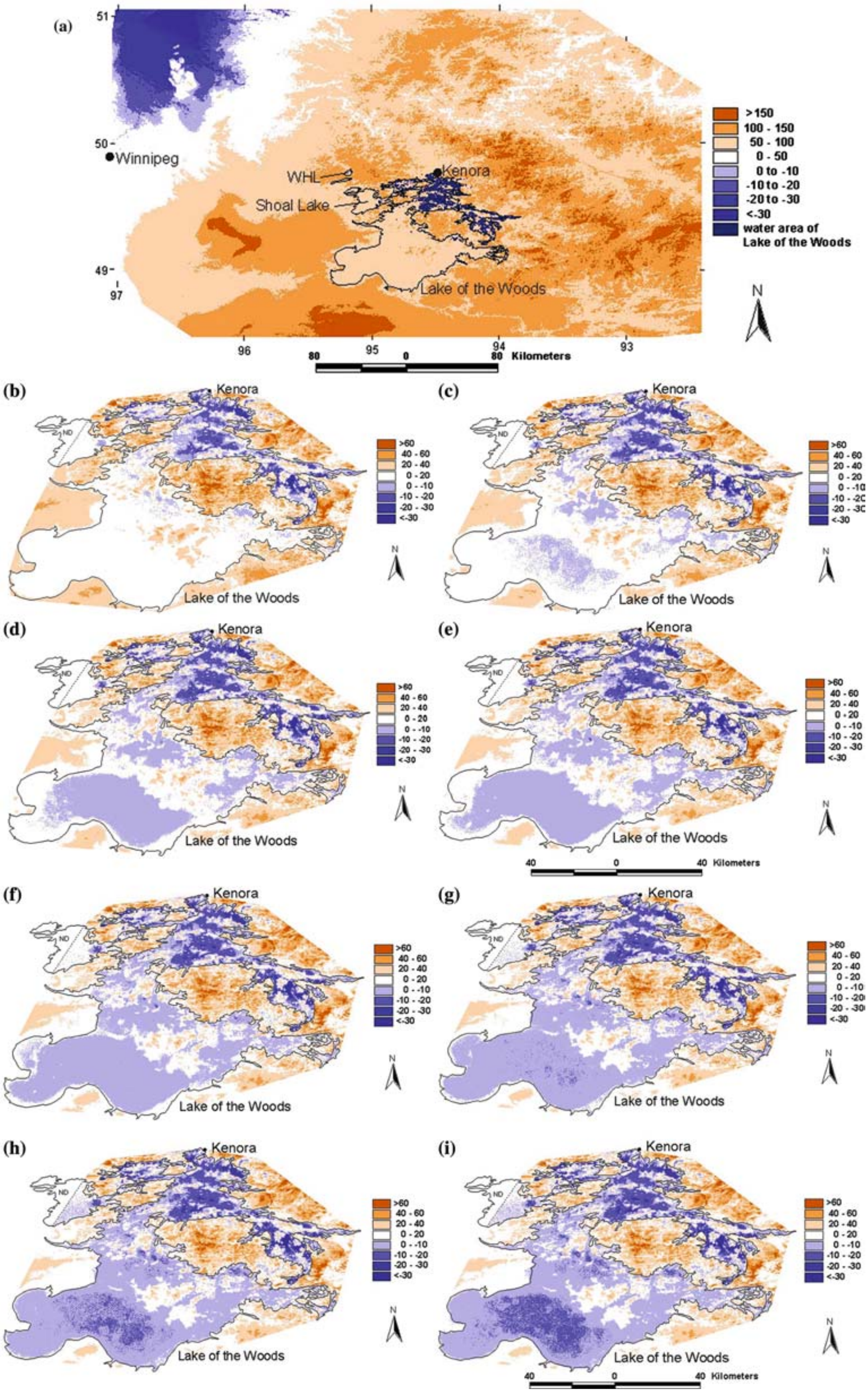

(a)
51
50
49
97
Winnipeg
WHL
Kenora
Shoal Lake
Lake of the Woods
96
95
94
93
80
0
80
Kilometers
>150
100 - 150
50 - 100
0 - 50
0 to -10
-10 to -20
-20 to -30
<-30
water area of
Lake of the Woods
N
(b)
(c)
(d)
(e)
(f)
(g)
(h)
(i)
ND
>60
40 - 60
20 - 40
0 - 20
0 - -10
-10 - -20
-20 - -30
<-30
40
0
40 Kilometers

Table 4. The evolution of lake area, bathymetry and volume of Lake of the Woods.

Periods	Cal yr B.P.	Maximum depth (m)	Mean depth (m)	Area (km^2)	Volume (km^3)
Lake of the Woods Period	Present	66.9	8.1	4524	37
	1000	67.0	7.6	4415	34
	2000	67.0	7.4	4292	32
	3000	66.0	6.9	4052	28
	4000	66.0	6.3	3676	23
	5000	65.0	5.8	3227	19
	6000	65.0	5.6	2857	16
	7000	64.0	6.1	1822	11
	8000	63.0	8.3	1061	9
	9000	61.0	8.5	858	7
Lake Agassiz Period	10,000	89.0	14.7	–	–
	10,500	116.0	31.4	–	–
	11,000	132.0	43.8	–	–

Wright et al. 2004), North Dakota (Coldwater Lake, Xia et al. 1997), and South Dakota (Moon Lake, Valero-Garcés et al. 1997; Pickerel Lake, Schwalb and Dean 1998), although the exact time varies from place to place.

In order to consider what the effect of warmer and drier conditions during the mid-Holocene might have had on Lake of the Woods, we first had to calculate a modern hydrological budget for the system, which had never been done before. If the level of the lake had been drawn-down below the level of the Kenora outlets, total overflow from Lake of the Woods (*S*) would have to have decreased to 0, and that would have required runoff (*R*) + precipitation directly on the lake (*P*) to have decreased from modern values; as well, evaporation from the lake surface (*E*) would likely have increased during a warmer and drier period such as the mid-Holocene. We used the following formula for our calculations of the hydrological budget of the lake

$$R + P + \mathrm{Gi} = S + E + \mathrm{Go} \pm \Delta, \qquad (7)$$

where Gi and Go = groundwater into and out of the lake, and Δ = change in lake level.

Some parameters in the budget equation are not known, so we attempted to estimate the most likely values. Runoff into the lake has only been monitored on the Rainy River, with the closest gauging station 80 km upstream from Lake of the Woods at Manitou Rapids; however, most tributaries to that river lie upstream from that station (Environment Canada 2004), and we concluded that raising the discharge from 365 to 400 m^3/s would provide a reasonable value. Macins (1972) estimated that the Rainy River contributes 75% of the total runoff to Lake of the Woods, so we increased the Rainy River discharge by 25% to 533 m^3/s (16.78 km^3/yr) to obtain the total runoff (*R*) from the entire drainage basin each year. The total outflow (*S*) from Lake of the Woods today is 13.59 km^3/yr, based on a mean annual discharge of 431 m^3/s (Environment Canada 2004). Annual precipitation on the lake averages 700 mm/yr (Environment Canada 2004) for a total *P* of 2.69 km^3/yr. Annual evaporative loss from the lake surface is 650 mm (Fisheries and Environment Canada 1978) for a total *E* of 2.5 km^3/yr. Substituting these values in Eq. (7):

Figure 6. Extent and depth (in meters) of waters in Lake of the Woods region (blue shades) and elevation (in meters) of surrounding land above the lake level of Lake Agassiz (white to orange shades) at (a) 9000 cal yr B.P., and above the lake level of Lake of the Woods at (b) 8000, (c) 7000, (d) 6000, (e) 5000, (f) 4000, (g) 3000, (h) 2000 and (i) 1000 cal yr B.P. The elevation of Lake Agassiz at 9000 cal yr B.P. was 196 m above modern sea level, whereas Lake of the Woods at this time was surrounded by higher topography and was 66 m above Lake Agassiz. The maximum depth of Lake of the Woods at 9000 cal yr B.P. was 61 m. The elevations of Lake of the Woods in the eight different times after that were 278, 290, 299, 305, 310, 314, 317, and 319 respectively (see Table 3). A negative hydrological budget, induced by climate warming during the Holocene, would have resulted in lower lake levels. ND = no data. The elevation of Lake of the Woods natural outlets was used to model the history of the lake, although the modern elevation of Lake of the Woods, controlled by several dams, has been 1.8 m higher than that of the natural outlet since the early 1900's.

$$16.78 + 2.69 + \mathrm{Gi} = 13.59 + 2.5 + \mathrm{Go} \pm 0,$$

$$19.47 + \mathrm{Gi} = 16.09 + \mathrm{Go},$$

so, Gi−Go = −3.38 km^3/yr, which is the net outflow of groundwater (Go).

During mid-Holocene warming, precipitation and, in turn, runoff to Lake of the Woods would have decreased, and evaporation would have increased. The hydrological budget may have become negative, with overflow ending, lake levels declining below those shown in Figure 6, and the lake shrinking in size. In order to reduce overflow (S) in Eq. (7) to 0, we reduced R and P by 65% and increased E by 40% to have a balanced budget. We kept groundwater as a net loss of 3.38 km^3/yr. In this calculation,

$$\text{The input is}: (2.69 + 16.78) * 35\% = 6.81\,\mathrm{km}^3/\mathrm{yr}.$$

$$\text{The loss is}: 0 + 2.5 * 140\% + 3.38 = 6.88\,\mathrm{km}^3/\mathrm{yr}.$$

In this paleohydrological budget, loss slightly exceeds input, so lake level and extent would decrease.

Lewis et al. (2001) investigated the change in climate required to explain low levels in nearby Lake Winnipeg during the mid-Holocene, which they had interpreted from the stratigraphic record of that lake. In their reconstruction of climate at 6500 ^{14}C yr B.P. (7.43 cal ka B.P.), which was based on pollen data in the region, Lewis et al. (2001) concluded that the grassland boundary had expanded east to the western edge of Lake of the Woods, 100–200 km farther east than its present location; they also moved the mid-Holocene parkland-boreal forest boundary east of the lake by nearly 100 km. In order to account for this vegetative shift Lewis et al. (2001) reduced precipitation and increased evaporation over the drainage basin of Lake Winnipeg. They estimated that mid-Holocene climate over the region that supplied water to the southern end of Lake Winnipeg may have been comparable to modern conditions of southeastern Alberta around Medicine Hat, and that climate impacting runoff into the northern end of that lake may have been comparable to that around Saskatoon, in central Saskatchewan. As a test of our calculation of the mid-Holocene hydrological budget for Lake of the Woods, we selected the paleohydrological parameters of Lewis et al. (2001) for central Saskatchewan during the mid-Holocene, substituting them in Eq. (7), and solving for runoff (R) with overflow = 0; we kept groundwater (Gi−Go) at −3.38 km^3/yr (net loss). In this calculation, $P{-}E$ during the mid-Holocene was reduced by 480 mm/yr from modern values, or 1.85 km^3/yr over the lake. Solving for paleo-runoff into Lake of the Woods:

$$R = (2.69 - 1.85) + 3.38 = 4.22\,\mathrm{km}^3/\mathrm{yr} \quad (8)$$

This reduction in runoff from 16.78 to 4.22 km^3/yr is a 75% reduction from the modern value, slightly more than the 65% we previously determined would produce a negative hydrological budget.

Bengtsson and Malm (1997) noted that the response of runoff to a decrease in precipitation is non-linear, and runoff may become negligible at some point as precipitation decreases. In the nearby Experimental Lakes Area (ELA) of northwestern Ontario, runoff decreases to almost zero when precipitation decreases to about half of current values, probably because evapotranspiration consumes most precipitation below certain values (R. Hesslein 2004, personal communication). Thus, a threshold may have been reached during the mid-Holocene, as precipitation decreased from today's value of 700 mm/yr and runoff approached zero; we believe our values for paleo-runoff of 4.22–6.81 km^3/yr represent reasonable estimates and are in agreement with other hydrological observations. For this reason, we believe it is very likely that water levels in Lake of the Woods did fall below the spillway at this time in response to well-established mid-Holocene warmer and drier conditions in the region. The areas covered by the lake would have been smaller than those shown in Figure 6 during much of the period from about 8 to 4 cal ka B.P., although levels would likely have fluctuated.

To date there is no stratigraphic information from Lake of the Woods to indicate that water levels fell below the spillway at Kenora and that the floor of the lake was subaerially exposed beyond that shown in Figure 6. However, because mid-Holocene aridity probably resulted in a negative hydrological budget, lake levels would have fallen below those shown in Figure 6, which are based only on differential isostatic rebound controls.

Summary

The changing depth and extent of Lake of the Woods since deglaciation of northwestern Ontario has been modeled using GIS techniques and by estimating the hydrological budget of the lake during the warmer and drier mid-Holocene period. By combining a DEM of modern topography, lake bathymetry, a model for differential isostatic rebound, and known fluctuations in glacial Lake Agassiz, snapshots of Lake of the Woods at 1000-year intervals over the past 11,000 years, and into the future, have been generated. These data indicate a southward expansion and deeping of the lake after separation from Lake Agassiz.

During the early postglacial period in northwestern Ontario, 11,000–10,000 cal yr B.P. ($\sim$9.6 to $\sim$8.9 ^{14}C ka B.P.), Lake of the Woods remained part of glacial Lake Agassiz, with waters in that basin at times 22–65 m above the level of the modern lake (Table 4). As Lake Agassiz levels declined, Lake of the Woods became increasingly isolated from the glacial lake, until by about 10,000 cal yr B.P. ($\sim$8.9 ^{14}C ka B.P.) it was only connected by a narrow embayment. By 9000 cal yr B.P. ($\sim$8.1 ^{14}C ka B.P.), Lake of the Woods became independent. Since then differential isostatic rebound has resulted in progressive southward transgression and expansion of the lake. However, reduced precipitation and increased evaporation across the region probably ended overflow and led to lower lake levels and a smaller lake during the mid-Holocene.

Acknowledgements

This research was supported by a Natural Sciences & Engineering Research Council of Canada Discovery Grant to JTT. We are grateful to Andries Blouw for providing the bathymetric map of Lake of the Woods that originally came from Garmin MapSource Company, which we digitized. Thanks also to Greg Keller for help with using Geographic Information System software. Reviews by Matthew Boyd and an anonymous reviewer are greatly appreciated.

References

Barber D.C., Dyke A., Hillaire-Marcel C., Jennings A.E., Andrews J.T., Kerwin M.W., Bilodeau G., McNeely R., Southon J., Morehead M.D. and Gagnon J.M. 1999. Forcing of the cold event of 8200 years ago by catastrophic drainage of Laurentide lakes. Nature 400: 344–348.

Bengtsson L. and Malm J. 1997. Using rainfall-runoff modeling to interpret lake level. J. Paleolimnol. 18: 235–248.

Blasco S.M. 2001. Geological history of fathom five National Marine Park over the past 15,000 years. In: Parker S. and Munawar M. (eds), Ecology, Culture, and Conservation of a Protected Area: Fathom Five National Marine Park, Canada, Ecovision World Monograph Series. Backhuys Publication, Leiden, Netherlands.

Boyd M., Running G.L. IV and Havholm K. 2003. Paleoecology and geochronology of glacial Lake Hind during the Pleistocene–Holocene transition: a context for Folsom surface finds on the Canadian Prairies. Geoarchaeol. Int. J. 18: 583–607.

Boyd M., Kling H., Schnurrenberger D. and Teller J.T. 2002. Coring of West Hawk Lake meteorite impact crater: deepwater technique and initial results. International Association Great Lakes Research, Abstracts of 35th Annual Conference, Winnipeg, 14 pp.

Broecker W.S., Kennett J., Flower B., Teller J.T., Trumbore S., Bonani G. and Wolfli W. 1989. Routing of meltwater from the Laurentide Ice Sheet during the Younger Drysa cold episode. Nature 341: 318–321.

Buchner A.P. and Pettipas L.F. 1990. The early occupations of the Glacial Lake Agassiz basin in Manitoba; 11,500 to 7,700 B.P. In: Lasca N.P. and Donahue J. (eds), Archaeological Geology of North America. Geological Society of America, Boulder, Colorado. Centennial Special Volume. pp. 51–59.

Clark P.U., Marshall S.J., Clarke G.K.C., Hostetler S.W., Licciardi J.M. and Teller J.T. 2001. Freshwater forcing of abrupt climate change during the glaciation. Science 293: 283–287.

Dean W.E. 1993. Physical properties, mineralogy, and geochemistry of Holocene varved sediments from Elk Lake, Minnesota. Geol. Soc. Am. Spec. Paper 276: 135–157.

Encyclopedia Britannica, Inc. 1999. Encyclopedia Britannica 99 CD. Multimedia Edition.

Environment Canada 2004. Water level and streamflow statistics. Website: http://www.wsc.ec.gc.ca/staflo/

Fisheries and Environment Canada 1978. Hydrological Atlas of Canada. Minister of Supply and Services, Canada. p.17.

Gareau P., Lewis C.F.M., Sherin A. and Macnab R. 1998. The paleo-Great Lakes, their areas and volumes from digital elevation model studies. Geological Society of America Annual Meeting, Abstracts with Programs, Toronto, Ontario, A-163.

Hostetler S.W., Bartlein P.J., Clark P.U., Small E.E. and Solomon A.M. 2000. Simulated interactions of proglacial Lake Agassiz with the Laurentide ice sheet 11,000 years ago. Nature 405: 334–337.

Hu F.S., Wright H.E. Jr., Ito E. and Lease K. 1997. Climatic effects of Glacial Lake Agassiz in the midwestern United States during the last deglaciation. Geology 25: 207–210.

Last W.M. and Teller J.T. 2002. Paleolimnology of Lake Manitoba: the lithostratigraphic evidence. Géogr. Phys. Quatern. 56: 135–154.

Leverington D.W., Mann J.D. and Teller J.T. 2000. Changes in the bathymetry and volume of glacial lake Agassiz between 11,000 and 9300 ^{14}C Yr BP. Quatern. Res. 54: 174–181.

Leverington D.W., Mann J.D. and Teller J.T. 2002a. Changes in the bathymetry and volume of glacial lake Agassiz between 9200 and 7700 ^{14}C Yr BP. Quatern. Res. 57: 244–252.

Leverington D.W., Teller J.T. and Mann J.D. 2002b. A GIS method for reconstruction of late Quaternary landscapes from isobase data and modern topography. Comput. Sci. 28: 631–639.

Lewis C.F.M. and Gareau P. 2001. Evaluationg Glacio-isostatic Uplift from Tilted Shoreline Data to Reconstruct the Paleo-bathymetry and Topography of the Great Lakes Basin. Program and Abstracts, Canadian Geophysical Union, Ottawa, ON.

Lewis C.F.M. and Thorleifson L.H. 2003. Empirical modeling of regional glacio-isostatic warping for evaluating drainage system development in Red River valley and Lake Winnipeg basin. In: Brooks G.R., George S.S., Lewis C.F.M., Medioli B.E., Nielsen E., Simpson S. and Thorleifson L.H. 2003 Geoscientific insights into Red River flood hazards in Manitoba. The final report of the Red River flood project. Natural Resources Canada.

Lewis C.F.M., Blasco S.M. and Coakley J.P. 2002. Severe Dry Climate Impact on the Laurentian Great Lakes Indicated by Early to Middle Holocene Lake Closure. Internat. Assoc. Great Lakes Research, Winnipeg. Abstracts 45th Annual Conference, 72 pp.

Lewis C.F.M., Forbes D.L., Todd B.J., Nielsen E., Thorleifson L.H., Henderson P.J., McMatin I., Anderson T.W., Betcher R.N., Buhay W.M., Burbidge S.M., Schröder-Adams C.J., King J.W., Moran K., Gibson C., Jarrett C.A., Kling H.J., Lockhart W.L., Last W.M., Matile G.L.D., Risberg J., Rodrigues C.G., Telka A.M. and Vance R.E. 2001. Uplift-driven expansion delayed by middle Holocene desiccation in Lake Winnipeg, Manitoba, Canada. Geology 29: 743–746.

Licciardi J.M., Teller J.T. and Clark P.U. 1999. Freshwater routing by the Laurentide Ice Sheet during the last deglaciation. In: Clark P., Webb R.S. and Keigwin L.D. (eds), Mechanisms of Global Climate Change at Millennial Time Scales. American Geophysical Union Monograph 112: 171–202.

Liengme B.V. 1997. A Guide to Microsoft Excel for Scientists and Engineers. Arnold/John W. Wiley & Sons, Inc., London/New York.

LWCB (Lake of the Woods Control Board) 2002. Managing the Water Resources of the Winnipeg River Drainage Basin. LWCB Brochure, pp. 7–20.

Macins V. 1972. The Fisheries of Lake of the Woods. Sport Fisheries Branch, Ministry of Natural Resources, Ontario, pp. 1–5.

Mann J.D., Leverington D.W., Rayburn J. and Teller J.T. 1999. The volume and paleobathymetry of glacial Lake Agassiz. J. Paleolimnol. 22: 71–80.

Matile G.L.D., Thorleifson L.H. and Lewis C.F.W. 2000. A model for postglacial expansion of Lake Winnipeg based on postglacial uplift. In: Todd B.J., Lewis C.F.M., Forbes D.L., Thorleifson L.H. and Erik N. (eds), 1996 Lake Winnipeg Project: Cruise Report and Scientific Results. Geological Survey of Canada Open File 3470.

McMartin I. 2000. Paleogeography of Lake Agassiz and regional post-glacial uplift history the Flin Flon region, central Manitoba and Saskatchewan. J. Paleolimnol. 24: 293–315.

Nielsen E. 1998. Lake Winnipeg coastal submergence over the last three centuries. J. Paleolimnol. 19: 335–342.

Peltier W.R. 1994. Ice age paleotopography. Science 265: 195–201.

Peltier W.R. 1998. Postglacial variations in the level of the sea: implications for climate dynamics and solid-earth geophysics. Rev. Geophys. 36: 603–689.

Pettipas L.F. and Buchner A.P. 1983. Paleo-indian prehistory of the glacial Lake Agassiz region in Manitoba, 11500 to 6500 B.P. In: Teller J.T. and Clayton L. (eds), Glacial Lake Agassiz. Geological Association of Canada Special Paper 26: 421–451.

Porter S.C., Sauchyn D.J. and Delorme D.L. 1999. The ostracode record from Harris Lake, southwestern Saskatchewan: 9200 years of local environmental change. J. Paleolimnol. 21: 35–44.

Schwalb A. and Dean W.E. 1998. Stable isotopes and sediments from Pickerel Lake, South Dakota, USA: a 12 ky record of environmental changes. J. Paleolimnol. 20: 15–30.

Teller J.T. 1987. Proglacial lakes and the southern margin of the Laurentide Ice Sheet. In: Ruddiman W.F. and Wright H.E. (eds), North America and Adjacent Oceans during the Last Deglaciation. Decade of North American Geology K-3: 39–69.

Teller J.T. 2001. Formation of large beaches in an area of rapid differential isostatic rebound: the three outlet control of Lake Agassiz. Quatern. Sci. Rev. 20: 1649–1659.

Teller J.T. 2004. Controls, history, outbursts, and impact of large late-Quaternary proglacial lakes in North America: Chapter 3. In: Gilespie A., Porter S. and Atwater B. (eds), The Quaternary Period in the United States. INQUA Anniversary Volume. Elsevier, pp. 45–61.

Teller J.T. and Clayton L. 1983. An introduction to glacial Lake Agassiz. In: Teller J.T. and Clayton L. (eds), Glacial Lake Agassiz. Geological Association of Canada, Special Paper 26: 3–5.

Teller J.T. and Last W.M. 1982. Pedogenic horizons in post-glacial sediment of Lake Manitoba, Canada. Earth Surf. Proc. Land. 7: 367–397.

Teller J.T. and Leverington D.W. 2004. Glacial Lake Agassiz: a 5000-year history of change and its relationship to delta ^{18}O record of Greenland. Geol. Soc. Am. Bull. 116: 729–742.

Teller J.T. and Thorleifson L.H. 1983. The Lake Agassiz-Lake Superior connection. In: Teller J.T. and Clayton L. (eds), Glacial Lake Agassiz. Geological Association of Canada Special Paper 26: 261–290.

Teller J.T., Leverington D.W. and Mann J.D. 2002. Freshwater outbursts to the oceans from glacial Lake Agassiz and their role in climate change during the last deglaciation. Quatern. Sci. Rev. 21: 879–887.

Teller J.T., Risberg J., Matile G. and Zoltai S. 2000. Post-glacial history and paleoecology of Wampum, Manitoba, a former lagoon in the Lake Agassiz basin. GSA Bull. 112: 943–958.

Valero-Garcés B.L., Laird K.R., Fritz S.C., Kelts K., Ito E. and Grimm E.C. 1997. Holocene climate in the Northern Great Plains inferred from sediment stratigraphy, stable isotopes, carbonate geochemistry, diatoms, and pollen at Moon Lake, North Dakota. Quatern. Res. 48: 359–369.

Wright H.E. Jr., Stefanova I., Tian J., Brown T.A. and Hu F.S. 2004. A chronological framework for the Holocene vegetational history of central Minnesota: the Steel Lake pollen record. Quatern. Sci. Rev. 23: 611–626.

Xia J., Haskell B.J., Engstrom D.R. and Ito E. 1997. Holocene climate reconstructions from tandem trace-element and stable-isotope composition of ostracodes from Coldwater lake, North Dakota, U.S.A. J. Paleolimnol. 17: 85–100.