"Glacial Isostatic Adjustment of the Laurentian Great Lakes Basin: Using the Empirical Record of Strandline Deformation for Reconstruction of Early Holocene Paleo-Lakes and Discovery of a Hydrologically Closed Phase"

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GLACIAL ISOSTATIC ADJUSTMENT OF THE LAURENTIAN GREAT LAKES BASIN: USING THE EMPIRICAL RECORD OF STRANDLINE DEFORMATION FOR RECONSTRUCTION OF EARLY HOLOCENE PALEO-LAKES AND DISCOVERY OF A HYDROLOGICALLY CLOSED PHASE*

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ABSTRACT In the Great Lakes region, the vertical motion of crustal rebound since the last glaciation has decelerated with time, and is described by exponential decay constrained by observed warping of strandlines of former lakes. A composite isostatic response surface relative to an area southwest of Lake Michigan beyond the limit of the last glacial maximum was prepared for the complete Great Lakes watershed at 10.6 ka BP (12.6 cal ka BP). Uplift of sites computed using values from the response surface facilitated the transformation of a digital elevation model of the present Great Lakes basins to represent the paleogeography of the watershed at selected times. Similarly, the original elevations of radiocarbon-dated geomorphic and stratigraphic indicators of former lake levels were reconstructed and plotted against age to define lake level history. A comparison with the independently computed basin outlet paleo-elevations reveals a phase of severely reduced water levels and hydrologically-closed lakes below overflow outlets between 7.9 and 7.0 ka BP (8.7 and 7.8 cal ka BP) in the Huron-Michigan basin. Severe evaporative draw-down is postulated to result from the early Holocene dry climate when inflows of meltwater from the upstream Agassiz basin began to bypass the upper Great Lakes basin.

RÉSUMÉ *Réajustement glacio-isostatique du bassin des Grands Lacs : reconstitution des anciens lacs au début de l'Holocène et mise en évidence d'une phase de fermeture hydrologique par des témoins empiriques de la déformation des lignes de rivage.* Dans la région des Grands Lacs, le soulèvement isostatique lié à la dernière glaciation a ralenti selon une courbe de décroissance exponentielle établie à partir du gauchissement observé dans les anciennes lignes de rivage lacustres. Une surface de référence composite de la réponse isostatique datant de 10,6 ka BP (12,6 cal ka BP) a été préparée pour l'ensemble du bassin-versant des Grands Lacs, par rapport à une région au sud-ouest du lac Michigan située au-delà de la limite du dernier maximum glaciaire. Le calcul du soulèvement des sites en fonction des altitudes de la surface de référence a facilité la conversion d'un modèle altimétrique de terrain du bassin actuel des Grands Lacs en des cartes paléo-géographiques pour différents âges choisis. De plus, afin de définir l'évolution des niveaux lacustres, les altitudes initiales des indicateurs géomorphologiques et stratigraphiques des paléo-niveaux lacustres, datés au ^{14}C, ont été reconstituées puis reportées en fonction de l'âge. La comparaison de cette courbe à celles des paléo-altitudes des exutoires lacustres, calculée indépendamment, révèle, entre 7,9 et 7 ka BP (8,7 et 7,8 cal ka BP), une phase d'abaissement majeur des niveaux lacustres au-dessous de ceux des exutoires et la fermeture hydrologique des lacs dans le bassin des lacs Huron-Michigan. La forte évaporation nécessaire à l'abaissement du niveau d'eau est attribuée à un climat sec peu après le début de l'Holocène, dans un contexte de détournement progressif hors des Grands Lacs des eaux de fonte du lac Agassiz.

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INTRODUCTION

The Great Lakes of North America consist of five major basins whose water surfaces comprise 32% of a total watershed area of 766 000 km^2 (The Great Lakes Environmental Atlas, 1995) (Fig. 1). The watershed forms the headwaters of the St. Lawrence River that drains to the Gulf of St. Lawrence and Atlantic Ocean. At the last glacial maximum, the watershed was completely covered by the southern flank of the Laurentide Ice Sheet about 18-21 ka BP (21.3-25.4 cal ka BP). Deglaciation of the basins occurred as the ice margin retreated generally in a northerly direction in a series of oscillations, first exposing the Erie basin about 15.5 ka BP (18.8 cal ka BP), and finally receding to the northern Superior basin about 9.5 ka BP (10.7 cal ka BP) (Table I; Dyke *et al.*, 2003; Dyke, 2004). During this retreat, a series of proglacial lakes formed shorelines of different ages that are upwarped today towards the north-northeast in the direction of thicker and longer-lasting ice (Fig. 2). This deformation is the cumulative isostatic adjustment of Earth's crust since formation of the shorelines as a result of removal of the former ice sheet load. This differential rebound proceeded throughout the period of ice retreat and postglacial time at decelerating rates of uplift. It is continuing today, as evidenced by tilting of the Great Lakes basins measured in long-term records of lake level gauges between southern and northern shores of the Great Lakes (CCGLBHHD, 1977; Tushingham, 1992; Mainville and Craymer, 2005).

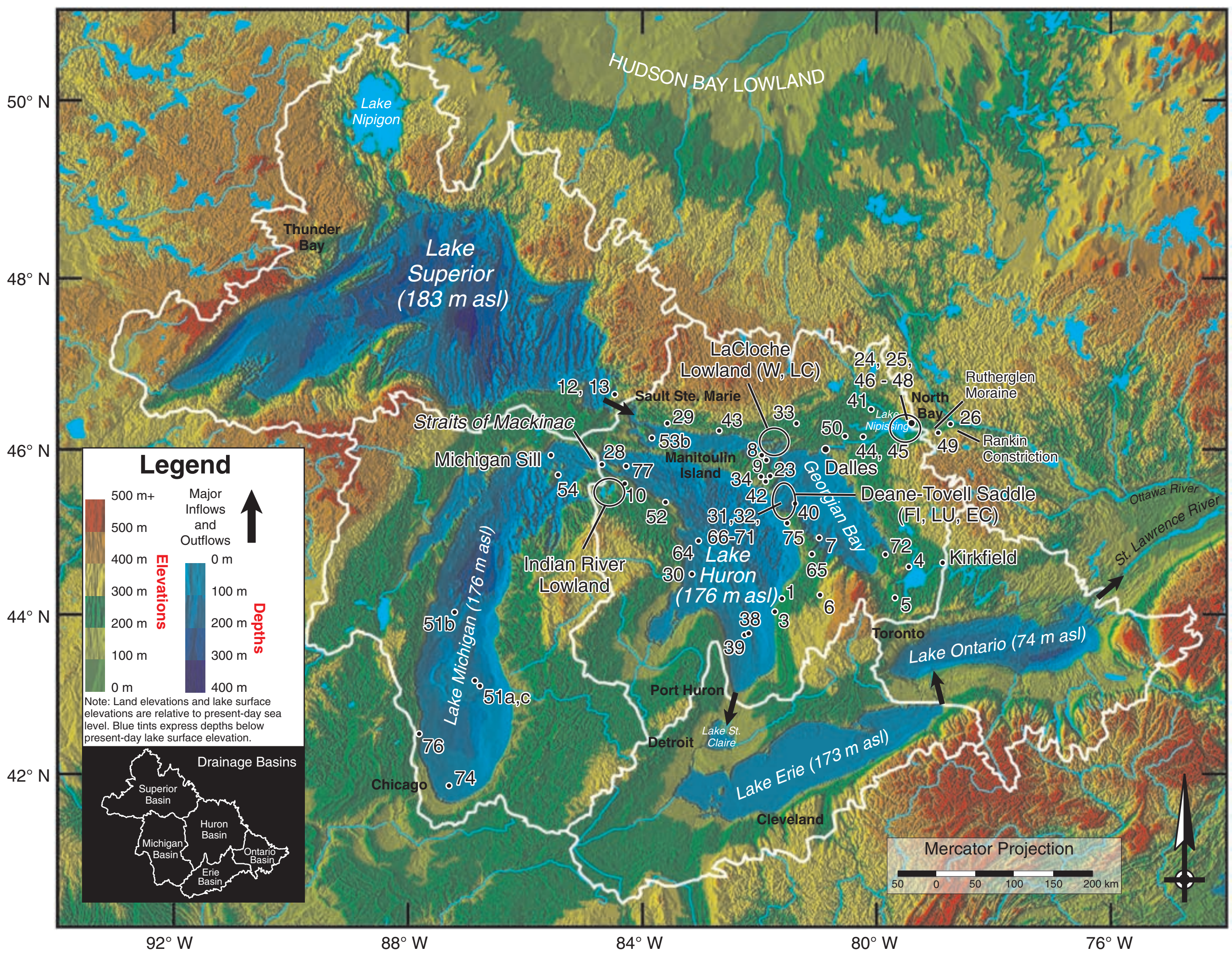

FIGURE 1. Shaded-relief map of the present-day Laurentian Great Lakes with their drainage areas outlined in white from the National Atlas of Canada (1985). The map file is of 30 arc-second resolution, and is from the Canadian Hydrographic Service (1996) with bathymetry added for Georgian Bay and the area west of LaCloche Lowland from United States Lake Survey (1965). Numbers refer to lake-level indicator sites listed in Table II. Lake outlets shown by black arrows.

Carte du relief de la région des Grands Lacs et des limites de ses sous-bassins de drainage (en blanc) établie à partir de l'Atlas national du Canada (1985). La base de données possède une résolution de 30 secondes d'arc et provient du Service hydrographique du Canada (1996), où la bathymétrie ajoutée de la baie Géorgienne et du secteur à l'ouest des basses terres de LaCloche provient de la Commission des Lacs des États-Unis (1965). Les chiffres renvoient aux sites d'indicateurs du niveau des lacs présentés dans le tableau II. Les flèches noires localisent l'exutoire des lacs.

TABLE I

Sources of isobase and deglacial information

Basin	References
Agassiz (northwest of Superior basin)	Clayton, 1983; Teller and Thorleifson, 1983; Teller, 1985; Thorleifson, 1996
Barlow and Ojibway basins (northeast and north of the Great Lakes basins)	Vincent and Hardy, 1979; Veillette, 1994; Dyke, 1996
Erie basin	Calkin, 1970; Barnett, 1979; Calkin and Feenstra, 1985
Huron-Georgian Bay basin	Goldthwait, 1910; Stanley, 1936, 1937, 1938a; Deane, 1950; Chapman, 1975; Karrow *et al.*, 1975; Karrow 1980; Chapman and Putnam, 1984; Eschman and Karrow, 1985; Kaszycki, 1985; Karrow, 1986, 1987; Lewis and Anderson, 1989; Lewis *et al.* 1994
Michigan basin	Goldthwait, 1907; Leverett and Taylor, 1915; Stanley, 1938b; Evenson, 1973; Futyma, 1981; Hansel *et al.*, 1985; Taylor, 1990; Kehew, 1993; Colman et al. 1994a
Ontario basin	Coleman, 1937; Muller and Prest, 1985; Pair *et al.*, 1988; Pair and Rodrigues, 1993
Ottawa River valley	Harrison, 1972; Chapman, 1975; Vincent and Hardy, 1979; Barnett, 1988; Veillette, 1994; Simard *et al.*, 2003
Superior basin	Clayton, 1983; Drexler *et al.*, 1983; Attig *et al.*, 1985; Farrand and Drexler, 1985; Thorleifson, 1996; Lowell *et al.*, 1999
Wisconsin basin (west of the Michigan basin)	Clayton and Attig, 1989

The differential and continuing nature of rebound has raised indicators of former lake levels to differing elevations, and this makes their correlation difficult and uncertain, particularly for those that are at scattered locations within the basin. This difficulty is especially evident for lake-level indicators that are now below the surface of the present Great Lakes. To facilitate the reconstruction of original elevations of former lake-level indicators, a reference isostatic response surface with an age of 10.6 ka BP is constructed for the entire basin. This surface is used in conjunction with an exponential function to describe the vertical motion in the time domain of any location in the entire watershed of the Great Lakes. This motion is constrained by, and consistent with, the observed empirical data provided by well-known upwarped strandlines in the Great Lakes basins. With this reference response surface and the exponential expression for uplift, the original elevations of specific lake-level indicator sites and potential overflow outlets are reconstructed and compared. The paleogeography of individual basins and of the entire Great Lakes watershed may likewise be reconstructed for any desired age. Applications of the isostatic response surface and the exponential model of uplift are illustrated for: (1) digital reconstruction of Great Lakes

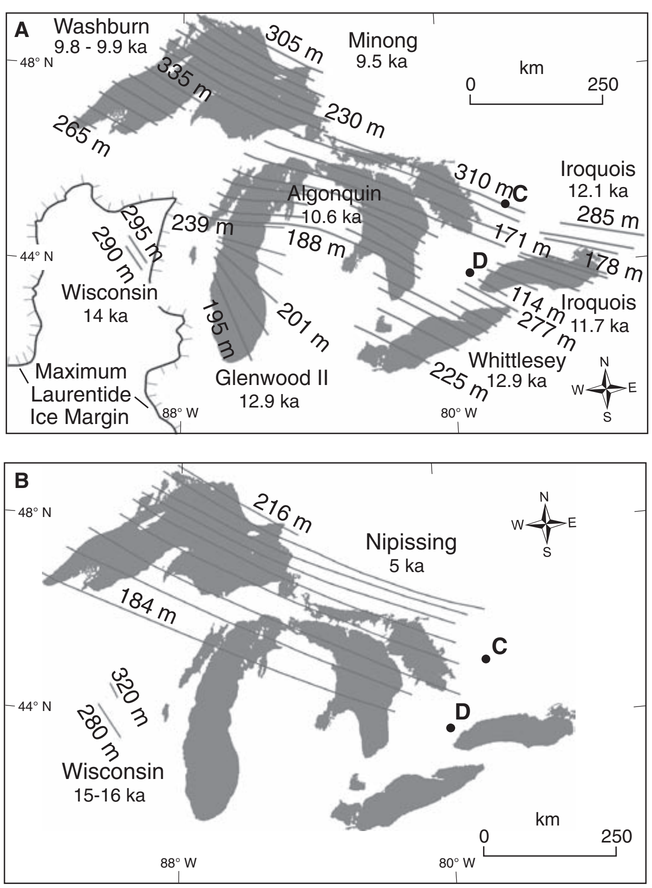

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FIGURE 2. Maps of isobases of selected reference paleoshorelines in basins of the Great Lakes, showing the radiocarbon ages (ka BP) and trends of isobases with their present lowest and highest elevations. Sources of isobases given in Table I. C and D mark end points of a section to which isobases of the Iroquois, Algonquin and Nipissing lakes were projected for illustration in Figure 3. (A) Map of Laurentide maximum ice margin and isobases of lakes Wisconsin, Washburn, Minong, Algonquin, Glenwood II, Whittlesey and Iroquois. (B) Map of isobases for the Nipissing Great Lakes, and for an older phase of glacial Lake Wisconsin.

Cartes des isobases des anciennes lignes de rivage de référence de la région des Grands Lacs, où les âges au ^{14}C (ka BP) et le soulèvement différentiel par rapport aux altitudes actuelles sont illustrées. Les sources bibliographiques sont présentées au tableau I. Les points C et D sont les limites du profil vertical projeté à la figure 3 des lacs Iroquois, Algonquin et Nippising. (A) Carte de l'extension glaciaire maximale de l'inlandsis laurentidien et des isobases des lacs Wisconsin, Washburn, Minong, Algonquin, Glenwood II, Whittlesey et Iroquois. (B) Carte des isobases de la phase Nipissing des Grands Lacs et pour une phase plus ancienne du lac Wisconsin.

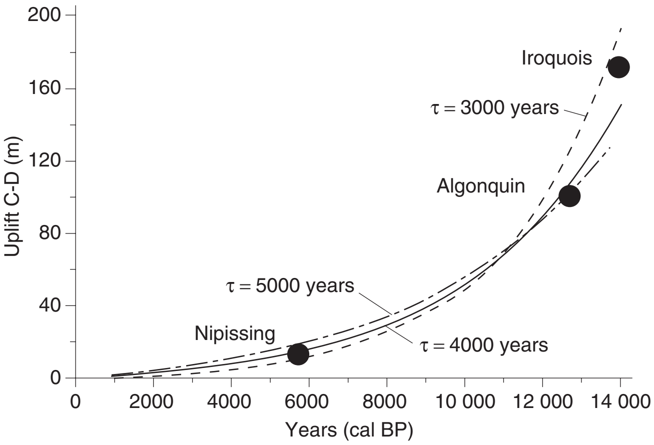

FIGURE 3. A plot of relative uplift vs. age for the Nipissing, Algonquin and Iroquois strandlines showing a relatively good fit to the exponential uplift equation for relaxation times (τ) between 3000 and 5000 years. The relative uplifts are the differences in projected strandline elevations at C and D shown on Figure 2.

Graphique temporel du soulèvement relatif des lignes de rivage des lacs Nipissing, Algonquin et Iroquois montrant un très bon ajustement à l'équation exponentielle de soulèvement calculée pour des temps de relaxation (τ) de 3000 à 5000 ans. Les soulèvements relatifs correspondent aux différences entre les altitudes projetées entre les points C et D de la figure 2.

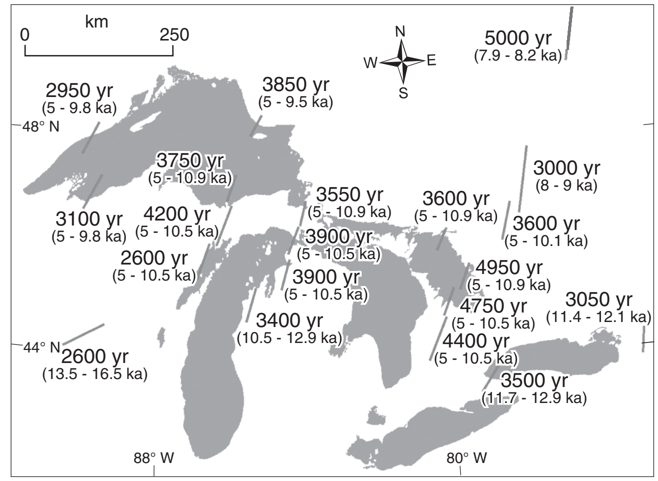

FIGURE 4. Map of the Great Lakes basin showing locations of transects where isobases of two shorelines of known age (shown as ^{14}C ka BP) provide a basis for computing the relaxation time of the uplifting process. The mean value of these estimates, 3700 ± 700 years, is used as the relaxation time throughout the basin.

Carte du bassin des Grands Lacs montrant l'emplacement des transects, où les isobases de deux lignes de rivage d'âge connu (montrés en ^{14}C ka BP) ont servi comme base de calcul du temps de relaxation du processus de soulèvement isostatique. La valeur moyenne de ces estimations, 3700 ± 700 ans, est utilisée comme temps de relaxation sur l'ensemble du bassin.

paleogeography in a geographic information system (GIS) environment, and (2) the assessment of former lake-level indicators in relation to possible overflow outlets, leading to the conclusion that the Michigan, Huron and Georgian Bay basins, if not all the Great Lakes, were once hydrologically closed.

UPLIFT IN THE TIME DOMAIN

ISOBASES OF GLACIO-ISOTATIC REBOUND

In the continental interior, isobases of glacio-isostatic rebound are usually defined by the elevations of a differentially uplifted shoreline of a former lake (Goldthwait, 1907, 1910; Leverett and Taylor, 1915; Hough, 1958; Walcott, 1972; Lewis and Anderson, 1989; Schaetzl *et al.*, 2002). Sets of isobases selected for this study for basins within the Great Lakes watershed are illustrated in Figure 2A and 2B in which the lowest and highest elevations, trends of isobases, the name and uncalibrated radiocarbon age are shown for each lake. Although the Algonquin and Nipissing highstands were confluent in three of the basins about 10.6 ka BP (12.6 cal ka BP) and 5.0 ka BP (5.7 cal ka BP), respectively, most sets of isobases are confined to a single basin. These former, once-level lake surfaces are all warped upward in a north to northeasterly direction (Fig. 2A-B) as a result of differential glacio-isostatic recovery of Earth's crust following deglaciation of the Laurentide Ice Sheet. The sources of isobase and related deglacial information for basins within and adjacent to the Great Lakes are listed in Table I.

UPLIFT AS A FUNCTION OF TIME

Whereas isobases portray the present configuration of a rebounding surface, it is often desirable to describe the uplift of a given location through time, or to construct surfaces at intermediate times, especially when reference shorelines are widely spaced in time. Following Andrews (1970) and others, the exponential function is adopted to describe relative uplift (U_t) with time where time is expressed as age t (cal years BP) for a landscape previously loaded by an ice sheet. From Peltier (1994, 1998) we use:

$$U_t = A * (\exp(t/\tau) - 1) \qquad \text{(Eq. 1)}$$

where A and τ (tau) are parameters of the equation. A is a site-specific amplitude factor, and is evaluated as:

$$A = U_t / (\exp(t/\tau) - 1) \qquad \text{(Eq. 2)}$$

for a known relative uplift, age and relaxation time. Tau (τ) is the relaxation time or period in years for which decelerating uplift is reduced by 1/exp (1/2.7183 or 36.8%) in successive periods. It is evaluated by solving (Equation 1) at sites where U_t is known for at least two sets of isobases of different ages as shown below. As a first-order approximation, τ is assumed to be time invariant and similar in value throughout the Great Lakes region. Figure 3 shows the reasonably good fit of the exponential uplift curve to Great Lakes relative rebound data for the Nipissing, Algonquin and Iroquois shorelines between sites C and D for trial values of relaxation time between 3000 and 5000 years (see Fig. 2 for location and isobase values).

Evaluation of parameters (τ and A) in the relative uplift equation (1)

Relaxation time (τ) was determined on 20 specific short transects (Fig. 4) where relative uplifts, U_1, U_2, and their ages t_1, t_2, are known for two shorelines, for example the Algonquin and Nipissing isobases whose domains largely overlap one another. Then, on each transect:

$U_1 = A * (\exp(t_1/\tau) - 1)$ and $U_2 = A * (\exp(t_2/\tau) - 1)$.

As the amplitude factor is identical on a specific short transect, these equations were rearranged in terms of A and subtracted to yield a single equation which was solved for τ:

$$U_2 * (\exp(t_1/\tau) - 1) - U_1 * (\exp(t_2/\tau) - 1) = 0 \quad \text{(Eq. 3)}$$

From Figure 4, a mean value of τ, rounded to 3700 ± 700 years, is used for computations of relative rebound in the Great Lakes basin. Similar values of 3400 years and 3500 ± 400 years, respectively, were obtained for the relaxation time of an exponential fit to relative sea-level changes in rapidly-uplifting James Bay south of Hudson Bay (Fig. 3A and 3C; Peltier, 1998), and for relaxation of glacial rebound in the Lake Winnipeg area, Manitoba (Lewis *et al.*, 2000; Brooks *et al.*, 2005).

The amplitude factor is then determined by evaluating Equation 2 using the known decay time τ:

$$A = U_1/(\exp(t_1/\tau) - 1) \text{ or } A = U_2/(\exp(t_2/\tau) - 1) \quad \text{(Eq. 4)}$$

Relative uplift (U_x) or shoreline slope (S_x) at other times

With τ and A known, relative uplift (U_x) or shoreline slope (S_x) for any given age t_x (cal years BP) in the same transect can be computed:

$$U_x = A * (\exp(t_x/\tau) - 1) \quad \text{(Eq. 5)}$$

U_x can be either larger or smaller than U_1 or U_2, and is the basis for removing the effects of rebound from a modern DEM for construction of the watershed paleogeography at a given time t_x, as described in a later section. Average shoreline slope (S_x) = relative uplift (U_x) / transect length (d) or:

$$S_x = U_x/d = (A/d) * (\exp(t_x/\tau) - 1) \quad \text{(Eq. 6)}$$

CONSTRUCTION OF A MULTI-BASIN RESPONSE SURFACE OF ALGONQUIN-AGE REBOUND ABOUT 10.6 KA BP (12.6 CAL KA BP)

Profiles perpendicular to isobases were selected on which an uplifted surface of Algonquin age could be constructed throughout the Great Lakes watershed (see locations of profiles on inset map of Fig. 5). Starting with area B on the Michigan-eastern Superior profile, the slope of the Elderdon phase shoreline of Lake Wisconsin (about 14 ka BP or 16.7 cal BP) was adjusted to the selected reference age, 10.6 ka BP (about 12.6 cal ka BP) using Equation 5 (Fig. 5). Then the reference profile was extended to the northeast using the Glenwood II shoreline. The uplift of the second isobase of the Glenwood II shore relative to its lowest isobase was adjusted for uplift using Equation 5. This procedure was repeated for each Glenwood II isobase until a complete shore profile, adjusted to the reference age, could be plotted. The adjusted Glenwood II profile was then shifted vertically to connect with the uplifted end of the Wisconsin profile. Next, the Algonquin profile, already at the selected reference age, was shifted vertically to connect with the end of the adjusted Glenwood II profile. The Minong isobases were then adjusted from an age of 9.5 ka BP (10.7 cal ka BP) to 10.6 ka BP (12.6 cal ka BP), and the adjusted shore profile was shifted vertically to connect with the end of the Algonquin profile. The continuous curve rising from zero to about 200 m at 800 km distance and beyond represents uplift on the Michigan-eastern Superior profile since 10.6 ka BP (12.6 cal ka BP). This rebound is relative to area B just beyond the Laurentide ice limit in the southwestern corner of the Great Lakes map area.

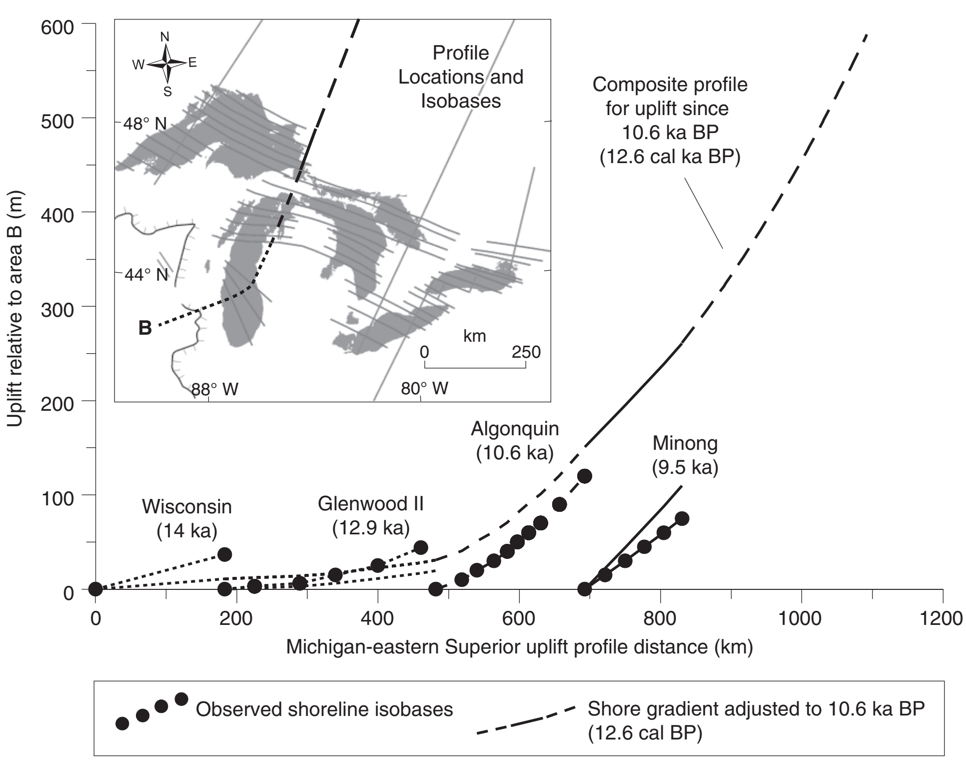

FIGURE 5. Michigan-east Superior profile showing construction of the gradients of shorelines representing uplift since the reference age of 10.6 ka BP (12.6 cal ka BP) relative to area B beyond the limit of the last glaciation. Dots are observed shoreline isobases. Dotted, dashed and solid lines are shore gradients adjusted to an age of 10.6 ka BP (12.6 cal ka BP). Inset map shows locations of reference profiles for contouring the reference response surface.

Profil du Michigan-est Supérieur qui montre le soulèvement des lignes de rivage depuis l'âge de référence de 10,6 ka BP (12,6 cal ka BP) par rapport à la région B située au-delà de la limite de la dernière glaciation. Les points sont les isobases des lignes de rivage observées, les lignes pointillées, les lignes tiretées et les lignes continues représentent les gradients de rivage ajustés à un âge de 10,6 ka BP (12,6 cal ka BP). L'insertion montre l'emplacement des profils utilisés pour délimiter la surface de référence de la réponse isostatique.

The adjusted Algonquin-age uplift values were transferred along connecting isobases to the other profiles (inset map on Fig. 5). After populating these profiles with Algonquin-age uplift values, contours were drawn throughout the region honouring the profile data and isobase trends. These contours constitute the *reference response surface* (Fig. 6) for the Great Lakes basin, representing isostatic rebound of the region from 10 600 BP (12 600 cal BP) to the present relative to area B. For a response surface at any other age, t_x, the contour lines remain the same, but each uplift contour value becomes Ut_x using Equation 5, *i.e.*

$$Ut_x = A_{contour} (\exp (t_x/3700) - 1) \quad \text{(Eq. 7)}$$

where $A_{contour}$ is the amplitude factor of the contour, such that:

$$A_{contour} = U_{10.6\ ka} / (\exp (12\,600/3700) - 1) \quad \text{(Eq. 8)}$$

and $U_{10.6\ ka}$ is the value of the contour from the reference response surface.

PALEOGEOGRAPHIC RECONSTRUCTION

The use of an isostatic response surface in paleogeographic reconstructions is described by Leverington *et al*. (2002) for Arctic Canada, and is similar to the use of relative-sea-level isobase maps as employed for reconstruction of Atlantic Canada paleogeography (Shaw *et al*., 2002). For paleogeographic reconstruction of the Great Lakes basin at a specific age, values for contours of an isostatic response surface at the specific age were first computed as outlined above in Equations 7 and 8. The new reference uplift contours, expressed as a vector data set of isobases, were transformed to a gridded data set or surface using the Triangulated Irregular Network (TIN) interpolation method. The interpolated values from this surface were subtracted from each corresponding pixel value of the modern Great Lakes digital elevation model (DEM) (Fig. 1) to generate a paleo-DEM for the desired age. Paleo-lake shorelines were determined by contouring the paleo-DEM within individual basins according to the elevation of their outlet sill or constriction, which is known from geological data. Paleo-DEM pixel values for areas within the shoreline contours were subtracted from the shoreline elevations to express paleo-lake water depths, which in turn were used to calculate lake area and volume. When present in the map area, an ice cover was superposed using information from the glacial geological literature (Table I) and from syntheses of deglaciation such as Barnett (1992) and Dyke *et al*. (2003). Twelve paleogeographic reconstructions, ranging from the Kirkfield Algonquin phase at 11.4 ka BP (13.3 cal ka BP) to the Nipissing Great Lakes at 5 ka BP (5.7 cal ka BP), were compiled using a GIS; images of two of these reconstructions are shown in Figure 7.

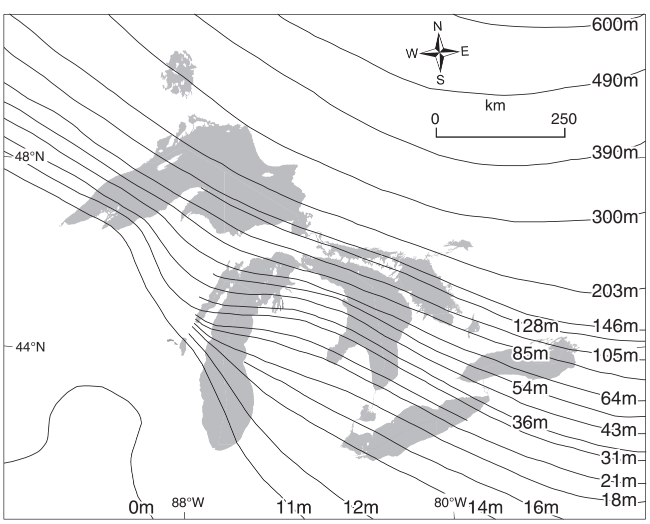

FIGURE 6. The reference isostatic response surface portraying glacial rebound since 10.6 ka BP (12.6 cal ka BP) in the Great Lakes region relative to an area southwest of Lake Michigan beyond the limit of the last glaciation.

Surface de référence de la réponse isostatique montrant le relèvement depuis 10,6 ka BP (12,6 cal ka BP) pour la région des Grands Lacs par rapport à un secteur au sud-ouest du lac Michigan se situant au-delà de la limite maximale de la dernière glaciation.

APPLICATION TO PALEOHYDROLOGICAL MODELING

The 12 GIS paleogeographic reconstructions based on the empirical model of isostatic adjustment were used to measure land and water areas, and lake volumes for studies of the paleohydrology and meltwater flow through the Great Lakes system (Moore *et al*., 2000). Variations in the basin attributes in the 11.4 to 5.0 ka BP (13.3 to 5.7 cal ka BP) period were substantial (Fig. 8). Maximum variation in individual basins under overflow conditions ranges from +72% to -95% for lake area, and from +200% to -97% for lake volume, compared to the present Great Lakes.

RECONSTRUCTION OF FORMER LAKE LEVELS AND DISCOVERY OF CLOSED LOWSTANDS, 8900-7800 CAL BP (7.9-7.0 KA BP)

Seventy-seven radiocarbon-dated, and two other, indicators of former lake levels in the Georgian Bay, Huron and Michigan basins (Table II and examples shown in Fig. 9) have been restored to their original elevations using a site uplift equation:

$$E_t = E_p - A * (\exp (t/\tau) - 1) \quad \text{(Eq. 9)}$$

Here, the elevation at time t cal BP is equated to the present elevation (E_p) minus the uplift of the site since time t. This equation, based on Equations 1 or 5, is used to compare the relative altitudes of two or more lake-level indicators or outlets at various times during their isostatic adjustment.

Original elevations for the reported error limits of each dated lake-level indicator were computed using Equation 9 and are shown in Table II. Most indicators are plotted in Figure 10 for the Huron and Georgian Bay basins, and in Figure 11 for the Michigan basin for the interval 11.7 to 6.2 ka BP (13.5 to 7.1 cal ka BP). On these figures, pairs of symbols joined by tie lines indicate the age and reported error range (X axis) in the date of each water surface indicator (Table II). The original altitude of an indicator is the Y-axis value of the plotted symbol; the tilt of the tie line represents uplift during the reported error range of its age. See Figure 1 for locations of the numbered indicators. The lake level histories

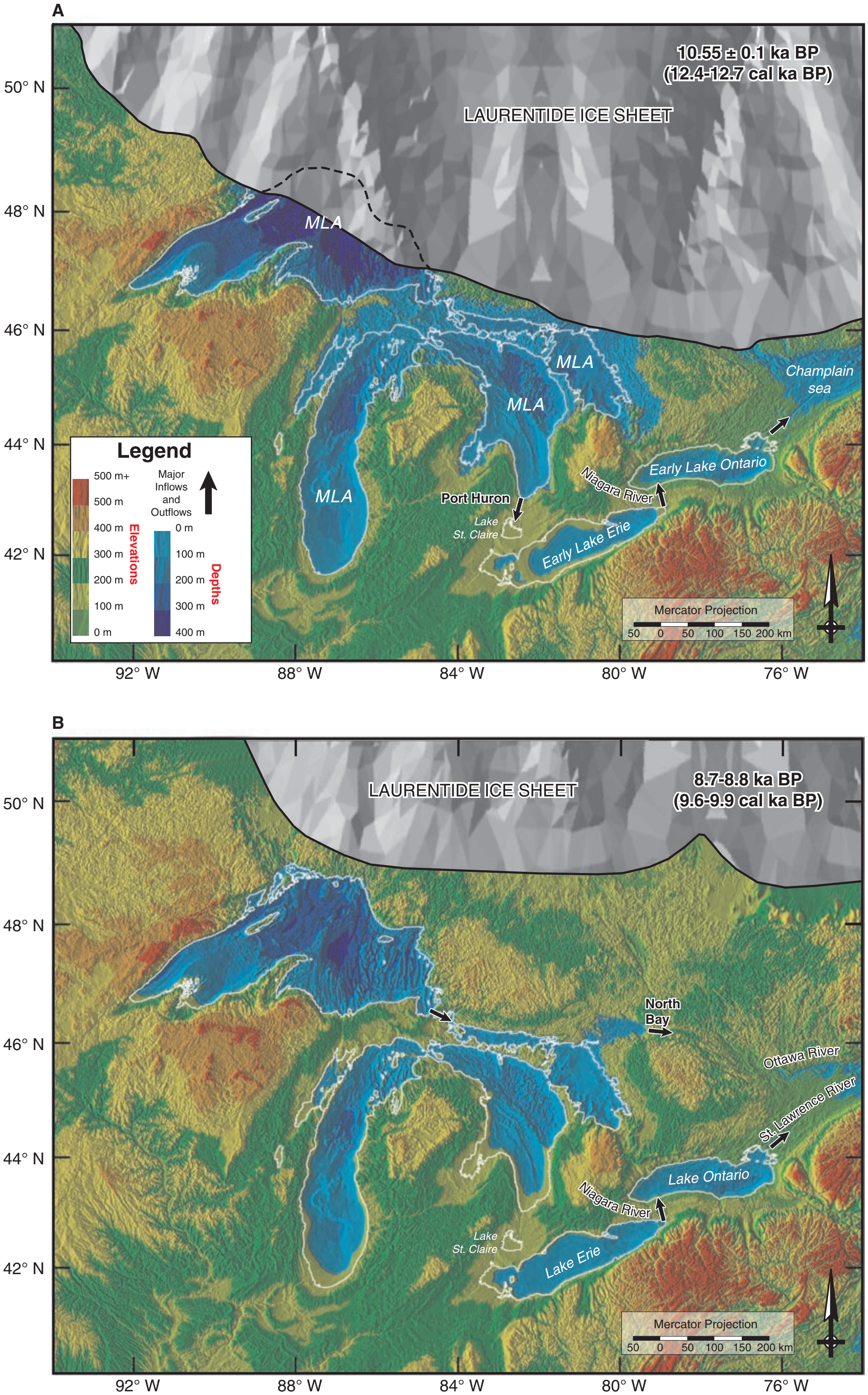

FIGURE 7. Paleogeographic maps showing reconstructions of the bathymetry and topography of the Great Lakes basin. (A) Main Lake Algonquin phase about 10.55 ± 0.1 ka BP (12.4-12.7 cal ka BP). At a late stage of Main Lake Algonquin, shown here, the lake had expanded to a possible maximum area in Superior (dashed line) and Huron basins by calving of icebergs from glacier margins in deep water. This lake overflowed via Port Huron to the Erie basin. Early Lake Erie, controlled by the Lyell-Johnson sill near the present Niagara Falls, overflowed via Niagara River to Early Lake Ontario which discharged via the emerging St. Lawrence River to Champlain Sea. (B) Mattawa highstand phase about 8.7-8.8 ka BP (9.6-9.9 cal ka BP), possibly a phase of high discharge when lake level in the upper Great Lakes basins was hydraulically dammed by constrictions downstream from North Bay at either the Rutherglen Moraine or the Rankin Constriction or both (Fig. 1). Discharge continued via Ottawa River to St. Lawrence River. There was no drainage from the upper to lower Great Lakes at this time; Lake Erie, Niagara River and Lake Ontario drained a separate watershed to the upper St. Lawrence River. Legend as for Figure 7A.

Cartes paléo-géographiques de la reconstitution de la bathymétrie et de la topographie du bassin des Grands Lacs. (A) Phase du lac Algonquin principal en 10,55 ± 0,1 ka BP (12,4-12,7 cal ka BP). Au cours du stade tardif de cette phase, le lac Algonquin a atteint son extension maximale dans les bassins du lac Supérieur (tireté) et Huron par le vêlage des icebergs issus des marges glaciaires en eau profonde. Ce lac se déversait dans le bassin du lac Érié par le Port Huron. Le lac Érié initial, dont le niveau était contrôlé par le seuil de Lyell-Johnson situé près des chutes actuelles du Niagara, se déversait dans le lac Ontario initial par la rivière Niagara, se déversant à son tour (le lac Ontario) dans le Saint-Laurent naissant par la Mer de Champlain. (B) Phase de haut niveau lacustre de Mattawa en 8,7-8,8 ka BP (9,6-9,9 cal ka BP), une phase de fort débit probablement liée à la fermeture hydraulique des Grands Lacs par un étranglement topographique situé en aval de North Bay provoqué par la moraine de Rutherglen, l'étranglement de Rankin, ou les deux. Le déversement dans le Saint-Laurent s'effectuait par la rivière des Outaouais. Il n'y avait pas de drainage entre l'amont et l'aval des Grands Lacs à cette époque : les lacs Érié et Ontario et la rivière Niagara drainaient un bassin-versant distinct vers le haut Saint-Laurent. La légende est la même qu'en A.

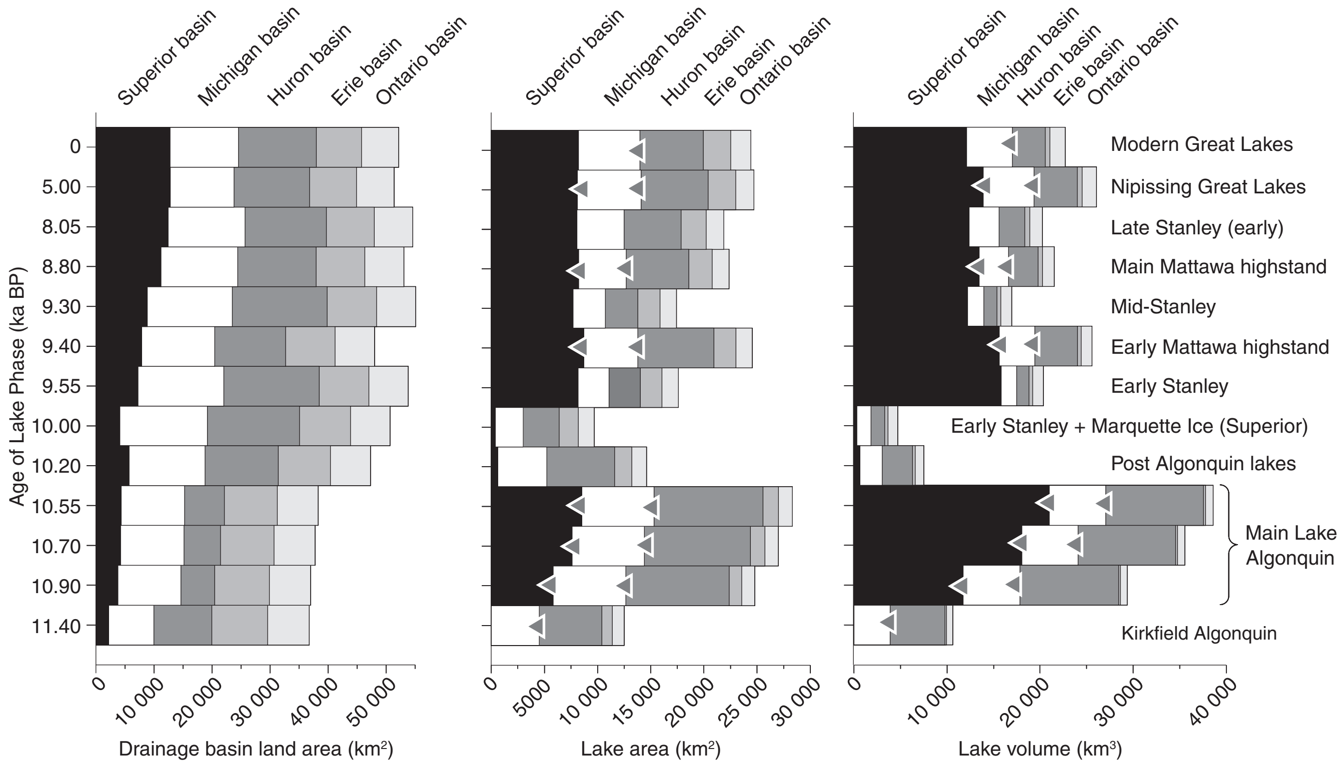

FIGURE 8. Bar graphs showing variation in the land drainage areas, lake surface areas, and lake volumes of selected Great Lakes overflowing phases measured from reconstructions of the paleo-Great Lakes in a GIS computing environment. Triangles indicate adjacent lakes that were and are confluent. Names at right refer to lake phases in Huron and other upper Great Lakes basins.

Diagrammes de la variation des superficies de drainage, des surfaces et des volumes des Grands Lacs au cours des différentes phases de déversement obtenues à partir des reconstitutions des paléo-Grands Lacs par un système d'information géographique (SIG). Les triangles indiquent s'il y a des lacs adjacents ayant ou encore connectés. Les noms à la droite font référence aux phases lacustres du bassin du lac Huron et des autres Grands Lacs.

were interpreted from the restored elevations of the radiocarbon-dated geologic indicators of former water levels in Huron and Michigan basins. For the times when lakes in the Michigan and Huron basins were confluent (relatively high levels), water levels in the Michigan basin were transferred from the Huron basin diagram.

The accuracy for estimating lake levels from empirical indicators is generally considered here to be at best about ±1 m based on survey error and variations in elevations of strandline bluffs, beaches, bars and spits (Schaetzl *et al.*, 2002). The computed initial elevations for lake-level indicators are considered accurate within about 2 m in areas of good isobase control, and somewhat >2 m in areas where constraining isobases have been extrapolated.

A second set of data (Table III) comprising the uplift history of lowest-possible potential overflow outlets for the Huron and Michigan basins, was also computed in the same manner and plotted on the diagrams of Figures 10 and 11 as dotted bands and dashed lines. These constraints include sills at North Bay (dotted band), Dalles Rapids, Deane-Tovell Saddle, and the head of Mackinac River (Lewis and Anderson, 1989). The North Bay sill controlled the overflow level of the water in the combined Georgian Bay, Huron and Michigan basins during the Nipissing transgression (Karrow 1980; Monaghan *et al.*, 1986; Colman *et al.*, 1994a, 1994b). The Dalles Rapids sill controlled overflow levels of the Georgian Bay basin during earlier phases while the North Bay sill was isostatically depressed at lower elevations. The Lucas and Fitzwilliam channel sills were selected as representative of constraints on overflow water levels in the northern Huron basin by the Deane-Tovell Saddle between Bruce Peninsula and Manitoulin Island. The Michigan basin overflow water levels were controlled by a sill at the head of the now-submerged Mackinac River channel between Michigan and Huron basins beneath northern Lake Michigan and the Straits of Mackinac (Stanley, 1938b).

INTERPRETATION OF HURON AND GEORGIAN BAY WATER-LEVEL HISTORY

13 500-11 000 cal BP

In this synthesis, the interpreted water-level history is shown as a thick line in Figures 10A-B and 11, and begins with the transgression of the Kirkfield Algonquin lake level to the Main Algonquin level (sites 4, 2, 1c, 1b, 3) under influence of the isostatically-rising Kirkfield outlet (Fig. 1). After 12 600 cal BP (10 600 BP), lake levels fell, indicated by the initiation of gyttja sedimentation in isolated basins (sites 12, 13, 10), through the Post-Algonquin phases and below, evidenced by the onset of organic accumulation on Manitoulin Island (23) and in Georgian Bay (7b). Land emergence by this fall of lake

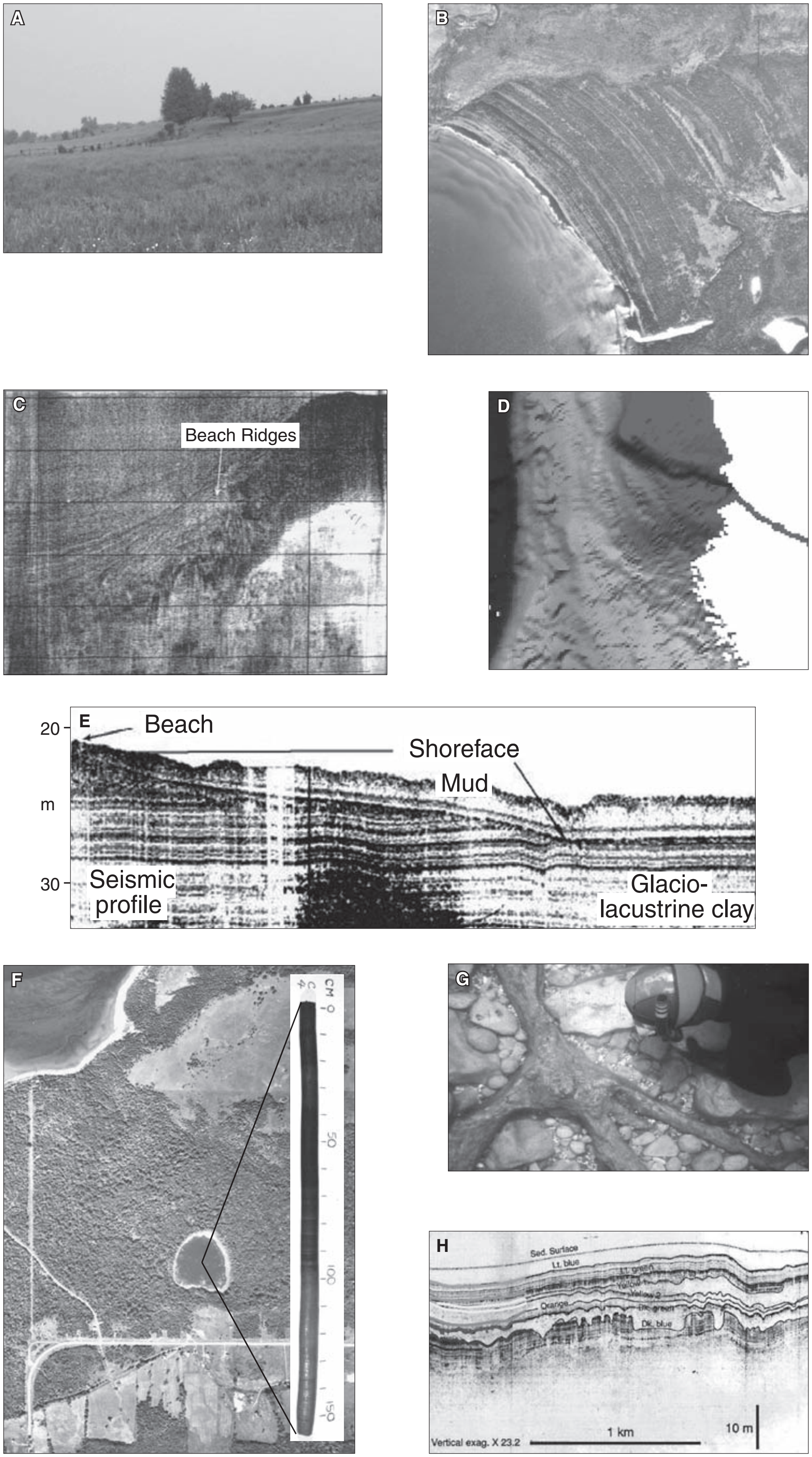

FIGURE 9. Types of evidence of former water levels in the Huron, Michigan, and Erie basins. Abandoned geomorphic shore features: (A) Former lake surfaces are inferred from erosional coastal features at the change in gradient between the gently-sloping shoreface and the steeply-sloping shorebluff, illustrated here for the Algonquin shore near Kirkfield, Ontario. (B) Storm beach ridges above Lake Huron on Manitoulin Island (air photo courtesy of Natural Resources Canada). (C) and (D) Beach ridges 53 m below Georgian Bay imaged by sidescan sonar and multibeam sonar, respectively (Blasco 2001). (E) Submerged beach and shoreface 21-28 m below Lake Erie (from Coakley and Lewis, 1985). Isolation basins, tree stumps, and unconformities: (F) Air photo (courtesy of Natural Resources Canada) view of a small basin isolated during the regression of a large lake, Manitoulin Island. A radiocarbon date of the contact zone between gray clastic large lake sediment and dark-coloured organic small-lake-sediment (photo courtesy of T.W. Anderson) provides the age at which the Great Lake surface passed below the sill elevation of the small isolated basin (Lewis, 1971). (G) *In situ* tree stump on lakefloor at entrance to Georgian Bay. (H) Subsurface reflections in a seismic profile indicate offshore unconformities and sequence boundaries caused by episodes of reduced lake level (Moore *et al.*, 1994).

Indicateurs des anciens niveaux lacustres dans les bassins des lacs Huron, Michigan et Érié. Formes littorales abandonnées : (A) les surfaces des anciens lacs sont estimées à partir de formes d'érosion côtière qui sont situées à la rupture de pente entre la face de la côte en pente douce et le talus littoral en pente abrupte, comme le montre cet exemple du rivage du lac Algonquin près de Kirkfield, Ontario. (B) Crêtes de plage de tempête en bordure du lac Huron sur l'Île Manitoulin (courtoisie de Ressources Naturelles Canada). (C) et (D) Crêtes de plage à 53 m de profondeur dans la baie Géorgienne identifiées respectivement à l'aide du sonar latéral et d'un sonar multi-faisceaux (Blasco, 2001). (E) Plage et avant-côte submergées entre 21 et 28 m sous la surface du lac Érié (d'après Coakley et Lewis, 1985). Bassins isolés, souches d'arbres et discontinuités : (F) Photo aérienne d'un petit bassin isolé durant la régression d'un grand lac, Île Manitoulin (Source : Ressources Naturelles Canada). Une datation au radiocarbone de la zone de contact entre les sédiments clastiques gris du grand lac et les sédiments organiques foncés du petit lac (photo fournie par T.W. Anderson) fournit l'âge auquel le plan d'eau du grand lac est passé sous le niveau du seuil du petit bassin isolé (Lewis, 1971). (G) Souche d'arbre in situ *au fond du lac à l'entrée de la baie Géorgienne. (H) Profil sismique montrant les discontinuités et les limites causées par des épisodes de bas niveau lacustre (Moore* et al.*, 1994).*

TABLE II

Indicators of former lake levels in the Michigan, Huron and Georgian Bay basins, and Georgian Bay-North Bay lowland

Site no.; lake, bog surface elevation[1]	Lat. N and Long. W	Laboratory no. and date	Material dated, elevation (m asl) and stratigraphy	Present elevation[2] and ref. uplift (m)[3]	Original lake elevation[4]	Cal. age years BP[5]	References
1b Kincardine Bog	44° 09.0' 81° 39.0'	GSC-1366 10 600 ± 150	Plant roots 195 from plant detritus over clay	195 46.2	151.3 146.2	12 400 12 800	Karrow *et al.*, 1975
1c Kincardine Bog	44° 09.0' 81° 39.0'	GSC-1374 11 200 ± 170	Plant detritus 194 under clay of Algonquin transgression	194 46.2	143.1 139.1	12 950 13 220	Karrow *et al.*, 1975
2 N Penetangore River	44° 10.0' 81° 38.0'	GSC-1842 11 300 ± 140	*Picea* or *Larix* wood under Algonquin sediment	191 47.0	137.6 133.7	13 060 13 310	Anderson, 1979
3 Eighteen Mile River	44° 01.3' 81° 43.6'	Pooled age 10 550 ± 110[j]	*Picea* wood on Algonquin-level river terraces	189 41.4	149.9 145.8	12 390 12 750	Karrow *et al.*, 1975; Karrow, 1986
			[j]GSC-1126 10 500 ± 150; [j]GSC-11127 10 600 ± 150				
4 Orillia	44° 34.6' 79° 26.3'	n.r.[7] 11 700 ± 250	Collagen of grizzly skull in gravel below Algonquin level	251 102.1	126.5 108.8	13 310 13 790	Peterson, 1965; Tovell and Deane, 1966
5 Cooks town bog	44° 13.3' 79° 37.3'	GSC-1111 10 200 ± 150	Emergent plant remains 229 over gyttja; after Main Lake Algonquin	229 77.4	173.3 160.6	11 430 12 160	Karrow *et al.*, 1975; T. Anderson, pers. comm., 2005
6 Wales Site, Everett	44° 12.2' 80° 57.0'	WAT-493 10 280 ± 100	Wood 225 in peat in gravel	225 58.4	178.0 170.6	11 830 12 350	Fitzgerald, 1985
7a Hope Bay 175.8	44° 55.0' 81° 07.1'	I-7857 8785 ± 145	Top of peat under massive grey clay	147.3 92.0	108.2 104.1	9600 9940	Lewis and Anderson, 1989
7b Hope Bay 175.8	44° 55.0' 81° 07.1'	I-7858 9930 ± 250	Base of peat over laminated clay	147.0 92.0	86.0 69.9	11 140 11 970	Lewis and Anderson, 1989
8 Green bush Swamp ~311	45° 56.1' 81° 59.7'	WAT-579 9930 ± 90	Swamp 3 m above Main Lake Algonquin. Basal gyttja 307.5	311 142.9	213.8 203.4	11 230 11 590	Warner *et al.*, 1984
9b Sheguiandah Bog 216	45° 53.7' 81° 55.4	Beta-92067 7860 ± 50	Plant macro-remains 215.2 from peat	216.4 141.4	171.9 169.9	8580 8730	Anderson, 2002
9c Sheguiandah Bog 216	45° 53.7' 81° 55.4	Beta-92069 9410 ± 60	Plant macro-remains 214.7 in basal peat	215.9 141.4	136.3 133.0	10 570 10 710	Anderson, 2002
9d Sheguiandah Bog 216	45° 53.7' 81° 55.4	TO-2346 9440 ± 80	Base of fibrous peat 214.3	214.3 141.4	135.3 129.7	10 540 10 780	Julig and Mahaney 2002
10 Lake Sixteen 216	45° 36.0' 84° 19.0'	WIS-2000 10 690 ± 100	Basal gyttja 211.9 over clayey silt	216 81.5	133.3 128.9	12 650 12 840	Futyma and Miller, 1986; R. Futyma, pers. comm., 1988
12 Upper Twin Lake 302	46° 32.5' 84° 35.0'	HEL-400 10 650 ± 265	Basal gyttja 291.4 over till	302 142.3	173.7 148.1	12 230 12 880	Saarnisto, 1974
13 Prince Lake 290	46° 33.5' 84° 33.0'	GSC-1715 10 800 ± 360	Basal gyttja 280.8 over clay	290 144.7	159.5 123.1	12 230 13 110	Saarnisto, 1974
23 Tehkummah Lake 191.7	45° 36.0' 81° 59.9'	GSC-1108 10 150 ± 190	Basal gyttja 184.2 over silty clay	191.7 118.2	107.9 89.3	11 380 12 090	Lewis, 1969; Lowdon *et al.*, 1971
24 Thibeault Hill Lake 312.4	46° 21.0' 79° 28.0'	GSC-638 9820 ± 200	Basal gyttja ca. 302 over silty clay	312.4 250.2	161.4 122.4	10 810 11 620	Lewis, 1969; Lowdon and Blake, 1968
25 Kilrush Lake 347	46° 05.0' 79° 28.0'	GSC-1246 9860 ± 270	Basal gyttja 330.5 over silty clay	347 225.7	211.6 166.1	10 790 11 810	Harrison, 1972; Lowdon and Blake, 1975
26 Morel Lake 194	46° 16.3' 78° 48.0'	GSC-1275 10 100 ± 240	Basal gyttja 180.2 over silty clay	194 252.9	20.1 -24.0	11 270 12 070	Harrison, 1972; Lowdon and Blake, 1975
28a Mackinac Straits 175.8	45° 49.1' 84° 43.8'	M-2337 8150 ± 300	Tree root (Tsuga) in growth position 138.5	138.5 90.8	108.7 101.7	8720 9440	Crane and Griffin, 1972
28b Mackinac Straits 175.8	45° 49.1' 84° 43.8'	M-1996 9780 ± 330	Tree stump in growth position 139.2	139.2 90.8	86.3 67.5	10 690 11 760	Crane and Griffin, 1970

TABLE II (continued)

Indicators of former lake levels in the Michigan, Huron and Georgian Bay basins, and Georgian Bay-North Bay lowland

Site no.; lake, bog surface elevation[1]	Lat. N and Long. W	Laboratory no. and date	Material dated, elevation (m asl) and stratigraphy	Present elevation[2] and ref. uplift (m)[3]	Original lake elevation[4]	Cal. age years BP[5]	References
29b Bruce Mines Bog 177.4	46° 17.8' 83° 44.6'	GSC-1359 8160 ± 220	Plant detritus bed 174.9 enclosed in clay	177.4 137.0	129.3 122.1	8950 9420	Lewis and Anderson, 1989
29c Bruce Mines Bog 177.4	46° 17.8' 83° 44.6'	GSC-1360 9560 ± 1603	Lowest plant detritus bed 174.1 enclosed in clay	177.4 137.0	97.3 86.3	10 700 11 150	Lewis and Anderson, 1989
30a West Lake Huron 175.8	44° 30.3' 83° 08.0'	GSC-1966 8460 ± 180	Gyttja bed 125.3 enclosed in silty clay	125.3 39.0	110.8 108.6	9150 9630	Lewis and Anderson, 1989
30b West Lake Huron 175.8	44° 30.3' 83° 08.0'	GSC-1943 8830 ± 410	Top woody peat bed 124.5 enclosed in silty clay	124.5 39.0	108.7 103.2	9450 10 470	Lewis and Anderson, 1989
30cd West Lake Huron 175.8	44° 30.3' 83° 08.0'	Pooled age 9370 ± 140[l]	Top and base woody peat bed 124.35 enclosed in silty clay	124.3 39.0	103.8 101.1	10 330 10 760	Lewis and Anderson, 1989
			[l]GSC-1935 9370 ± 180; [l]GSC-1982 9370 ± 220				
30e West Lake Huron 175.8	44° 30.3' 83° 08.0'	GSC-1965 9170 ± 140	Base woody peat bed 124.2 enclosed in silty clay	124.2 39.0	104.5 102.7	10 210 10 520	Lewis and Anderson, 1989
30f West Lake Huron 175.8	44° 30.3' 83° 08.0'	GSC-1983 9680 ± 110	Base woody peat bed 124.5 enclosed in silty clay	124.5	101.1 98.1	10 800 11 220	Lewis and Anderson, 1989
31a Flummerfelt Basin 175.8	45° 22.0' 81° 31.7'	GSC-1847 7740 ± 360	*Populus* driftwood 122.7 in clay	122.7 111.7	91.5 82.8	8190 9010	Sly and Sandilands, 1988
31b Flummerfelt Basin 175.8	45° 22.0' 81° 31.7'	GSC-1830 9770 ± 220	*Salix* driftwood 122.4 in clay	122.4 111.7	56.2 38.3	10 750 11 590	Sly and Sandilands, 1988
32a Middle Island Channel 175.8	45° 16.5' 81° 38.2'	BGS-71 10 305 ± 78	White cedar driftwood 143.8	143.8 104.4	56.0 46.1	11 980 12 360	Sly and Lewis, 1972; This study
32b Middle Island Channel 175.8	45° 16.37' 81° 38.13'	BGS-2166R 8745 ± 75	Poplar driftwood 140.7	140.7 104.6	96.1 92.4	9610 9880	This study
32c Middle Island Channel 175.8	45° 16.38' 81° 38.14'	BGS-2240 9039 ± 80	White cedar driftwood 135.0	135.0 104.6	85.9 80.8	9940 10 280	This study
32d Middle Island Channel 175	45° 16.38' 81° 38.14'	BGS-2241 8742 ± 75	White cedar driftwood 135.0	135.0	90.6 86.7	9590 9880	This study
33 Wood Lake 218	46° 12.9' 81° 44.1'	GSC-606 9620 ± 250	Basal gyttja 205 over clay	218 183.1	114.9 92.4	10 570 11 260	Lewis, 1969; Lowdon *et al.*, 1967
34a Smoky Hollow Lake 192.7	45° 38.1' 82° 04.3	I-4037 6270 ± 190	Top of lowest gyttja189 below silty clay	192.7 119.0	169.9 167.0	6970 7350	Lewis, 1969
34b Smoky Hollow Lake 192.7	45° 38.1' 82° 04.3	I-4036 9130 ± 140	Base of lowest gyttja 187.7 over silty clay	192.7 119.0	132.8 126.6	10 180 10 520	Lewis, 1969
38 South Lake Huron 175.8	43° 53.3' 82° 14.7'	GSC-3656 9350 ± 90	Plant detritus 107.3 under clayey silt	107.3 33.2	89.3 88.0	10 440 10 690	Lewis and Anderson, 1989
39 South Lake Huron 175.8	43° 53.7' 82° 17.2'	GSC-3577 8890 ± 100	Peat 103.2 and shells in silty clay under silty clay	103.2 32.9	88.3 86.5	9810 10 190	Woodend, 1983; Lewis and Anderson, 1989
40 Flummerfelt Patch 175.8	45° 21' 81° 32.9'	GSC-1397 9440 ± 160	Subaerial peat under laminated clay 144.4	144.4 110.5	83.4 76.4	10 500 10 880	Sly and Lewis, 1972; Sly and Sandilands, 1988
41 Tanner Lake 250.5	46° 26.9' 80° 01.0'	Beta-19151 9420 ± 120	Basal gyttja 236.4 over laminated clay	250.5 250.2	112.8 100.0	10 490 10 800	Lewis and Anderson, 1989
42ab South Bay 175.8	45° 34.9' 81° 59.5'	Pooled age 8130 ± 65[m]	Seeds 163.3 in plant detritus bed in silty clay and clay	163.3 116.9	121.7 118.9	8990 9210	Rea *et al.*, 1994a

TABLE II (continued)

Indicators of former lake levels in the Michigan, Huron and Georgian Bay basins, and Georgian Bay-North Bay lowland

Site no.; lake, bog surface elevation[1]	Lat. N and Long. W	Laboratory no. and date	Material dated, elevation (m asl) and stratigraphy	Present elevation[2] and ref. uplift (m)[3]	Original lake elevation[4]	Cal. age years BP[5]	References
			[m]AA-8773 8075 ± 90 ; [m]AA-8772 8185 ± 90				
42c South Bay 175.8	45° 34.9' 81° 59.5'	GSC-1979 8310 ± 130	Upper plant detritus 160.3 beneath silty clay	160.3 117.0	116.7 112.7	9150 9450	Lewis and Anderson, 1989
42d South Bay 175.8	45° 34.9' 81° 59.5'	AA-8774 8910 ± 956	Shell 162.7 enclosed in silt and clay	162.7 116.9	108.4 103.5	9900 10 200	Rea *et al.*, 1994a
42e South Bay 175.8	45° 34.9' 81° 59.5'	GSC-1971 9260 ± 290	Lower plant detritus with shells under silty clay	159.5 116.9	101.0 83.3	10 160 11 080	Lewis and Anderson, 1989
42f South Bay 175.8	45° 34.9' 81° 59.5'	AA-8775 9790 ± 956	Shell 162.6 enclosed in silt and clay	162.6 116.9	86.0 80.8	11 100 11 330	Rea *et al.*, 1994a
43 Blind River Bog 218	46° 12.8' 82° 56.3'	GSC-514 8760 ± 2503	Lagoonal basal gyttja 214 behind barrier beach 221	218 148.3	155.8 147.9	9550 9960	Lowdon *et al.*, 1967
44 Wolseley Bay lake 206.4	46° 06.4' 80° 20.5'	GSC-1178 8110 ± 170	Basal gyttja 191.6 over clay	206.4 209.7	136.6 125.2	8770 9280	Lewis and Anderson, 1989
45 Monet Lake 201.5	46° 09.8' 80° 21.0'	GSC-1389 8250 ± 180	Basal gyttja 190.2 over silt	201.5 215.1	124.3 114.2	9020 9440	Lewis and Anderson, 1989
46 Dreany Lake 213.5	46° 17.4' 79° 21.8'	GSC-815 8200 ± 160	Basal gyttja 204 over clay	213.5 246.0	125.8 114.5	9000 9410	Lewis, 1969; Lowdon *et al.*, 1971
47 North Bay Lake 211.8	46° 17.4' 79° 20.1'	GSC-821 8320 ± 170	Basal gyttja 207 over varved clay	211.8 246.6	120.9 110.2	9110 9490	Lewis, 1969; Lowdon *et al.*, 1971
48 Trout Mills delta	46° 19.8' 79° 24.2'	GSC-1263 8050 ± 190	Wood 212 under clay between cross-bedded sand units	212 248.6	132.6 117.7	8630 9210	Harrison, 1972; Lowdon and Blake, 1975
49 Amable du Fond River	46° 11' 78° 57'	GSC-1097 8750 ± 140	Wood 242 in sand over varved clay	242 243.3	136.6 124.6	9550 9930	Harrison, 1972; Lowdon and Blake, 1975
50 Pure Lake 216.4	46° 8.7' 80° 32.7'	Beta-19153 8750 ± 140	Basal gyttja 206 over masssive clay	216.4 208.3	129.1 118.8	9550 9930	Lewis and Anderson, 1989
51a Lake Michigan 175.8	43° 10' 86° 50'	M-1571 7400 ± 500[6]	Shell 71.3 in sand above eroded clay	71.3 12.8	68.2 67.1	7680 8750	Crane and Griffin, 1965
51b Lake Michigan 175.8	44° 00.0' 87° 14.0'	M-1972 7570 ± 250[6]	Shell 68.8 in sand above eroded clay under silty clay	68.8 16.0	64.6 63.9	8130 8620	Crane and Griffin, 1970
51c Lake Michigan 175.8	43° 08.4' 86° 48.7'	M-1736 7580 ± 350[6]	Shell 79.8 in sand above eroded clay under silty clay	79.8 12.8	76.4 75.5	8000 8790	Crane and Griffin, 1968
52 Thompson Harbour 175.8	45° 23' 83° 36'	M-1012 7250 ± 300	Tree stump 171.3 in growth position on lakebed	171.3 77.6	152.0 148.4	7800 8370	Crane and Griffin, 1961
53b Rains Lake 192.7	46° 06.1' 83° 54.1'	GSC-1368 7090 ± 150	Top plant detritus 183.4 under sandy silt	183.4 119.3	154.3 151.5	7740 8040	Lewis and Anderson, 1989
54 Beaver Island 175.8	45° 42.1' 85° 25.2'	M-1888 6788 ± 250	Tree stump (*Pinus resinosa*) 166.1 in growth position	166.1 74.6	149.5 147.2	7440 7870	Crane and Griffin, 1968
63 Mississagi Strait 175.8	45° 58.2' 83° 11.2'	AA-8770 8985 ± 100	Driftwood 103 in silty clay	103 121.8	46.0 40.6	9930 10 240	Rea *et al.*, 1994a
64 West Lake Huron 175.8	44° 54.9' 83° 00.3'	AA-8871 9235 ± 100	Driftwood 110 in silty clay	110 56.5	80.8 78.7	10 270 10 510	Rea *et al.*, 1994a
65 Colpoy's Bay 175.8	44° 45.9' 81° 07'	n.r.[7] 7660 ± 50	*In situ Thuja* tree stump 166	166 82.2	141.5 140.8	8400 8500	Larson and Kelly, 1994
66 Bad Neighbour 175.8	45° 20.49' 81° 48.00'	Pooled age[p] 9320 ± 30	White cedar tree stump 132.6 in growth position	132.6 105.9	74.1 72.9	10 500 10 570	Blasco *et al.*, 1997; This study

TABLE II (continued)

Indicators of former lake levels in the Michigan, Huron and Georgian Bay basins, and Georgian Bay-North Bay lowland

Site no.; lake, bog surface elevation[1]	Lat. N and Long. W	Laboratory no. and date	Material dated, elevation (m asl) and stratigraphy	Present elevation[2] and ref. uplift (m)[3]	Original lake elevation[4]	Cal. age years BP[5]	References
			[p]Beta-98369 9360 ± 805; [p]Beta-103563 9300 ± 50; [p]Beta-104923 9320 ± 50				
67 Lucas Island Channel 175.8	45° 23.39' 81° 45.23'	Beta-87732 8560 ± 70	White cedar tree stump 157.5 in growth position	157.5 109.7	112.4 111.2	9480 9570	Blasco *et al.*, 1997; This study
68ab Lucas Island Channel 175.8	45° 23.36' 81° 45.29'	Pooled age[q] 8200 ± 40	White cedar tree stump 159.1 in growth position	159.1 109.3	119.1 117.1	9090 9250	Blasco *et al.*, 1997; This study
			[q]Beta-81977 8200 ± 60; [q]Beta-81978 8200 ± 60				
69 Lucas Island Channel 175.8	45° 23.36' 81° 45.29'	Beta-81979 7770 ± 60	White cedar tree stump 158.4 in growth position	158.4 109.3	125.0 123.8	8480 8600	Blasco *et al.*, 1997; This study
70 Cassels Cove 175.8	45° 19.04' 81° 43.50'	Beta-87733 7490 ± 80	Tamarack stump 172.8 in growth position	172.8 105.6	142.9 141.5	8230 8380	Blasco *et al.*, 1997; This study
71 Cassels Cove 175.8	45° 19.04' 81° 43.50'	Beta-87734 7230 ± 90	Cedar tree stump 172.8 in growth position	172.8 105.6	145.1 143.6	7980 8150	Blasco *et al.*, 1997; This study
72 Wye Marsh 176	44° 43.4' 79° 51.5'	n.r.[7] 7940 ± 110	Pine wood 174 at base of marsh sediment	174 105.2	140.1 137.1	8660 8950	Chittenden, 1990
74a Olson Forest 175.8	41° 49' 87° 18'	ISGS-2036 8120 ± 100	Oak tree stump 151 in growth position	151 9.1	147.8 147.5	8970 9270	Chrzastowski *et al.*, 1991
74b Olson Forest 175.8	41° 49' 87° 18'	Beta-34357 8380 ± 100	Oak tree stump 151 in growth position	151 9.1	147.5 147.2	9290 9500	Chrzastowski *et al.*, 1991
75a Little Eagle Harbour 175.8	45° 08.54' 81° 35.09'	Beta-103564 8020 ± 60	White cedar stump 166.7 in growth position	166.7 97.7	134.0 131.8	8790 9010	This study
75b Little Eagle Harbour 175.8	45° 09' 81° 35'	BGS-2167 8126 ± 90	White pine driftwood 166.7	166.7 98.1	131.9 129.0	8990 9250	This study
75c Little Eagle Harbour 175.8	45° 09' 81° 35'	BGS-2168 7972 ± 100	Small white cedar in situ stump 166.8	166.8 98.1	134.8 131.9	8700 8990	This study
76 Southwest Lake Michigan shore	42° 30.0' 87° 50.0'	ISGS-187 7370 ± 90	Forest bed 176.2, branches and roots	176.2 8.1	174.0 173.9	8060 8310	Fraser *et al.*, 1990
77 Stanley unconformity 175.8	45° 48.3' 84° 16.8'	PSV[8] 7900 ± 300	Sand in silty clay over eroded clay 125.8	125.8 95.5	97.3 90.6	8410 9110	Lewis *et al.*, in press
FB Flower pot Beach 175.8	45° 16' 81° 37'	Estimated 7900 ± 300	Sand beach ridges 125.3 over plant detritus	125.3 104.3	94.1 86.9	8410 9110	Blasco, 2001

1. Surface elevation (m asl) of sampled lake or bog.

2. Present elevation (m asl) of feature representing former level of a Great Lake.

3. Uplift (m) since 10.6 ka BP (12.6 cal ka BP) from reference isostatic surface (Fig. 6).

4. Elevations (m asl) computed using equation (9) for the age limits of the calibrated age range of the lake-level indicator as listed in column 7.

5. Ages at limits of the calibrated range (1σ) of the lake-level indicator date and its uncertainty reported in column 3. Calibration by the Calib 5.0.1 program (Stuiver and Reimer, 1993) using the INTCAL98 calibration dataset (Stuiver *et al.*, 1998).

6. Shell age uncorrected for reservoir effect.

7. n.r. = not reported.

8. PSV = Age by comparing paleomagnetic secular variation with a ^{14}C-dated standard for northeastern U.S.A. (King and Peck, 2001).

TABLE III

Present and early Holocene elevations[1] of North Bay, Georgian Bay, Huron and Michigan sills (m above present sea level)

Age years		North Bay	Georgian Bay sill	Huron basin sills					Michigan basin sill
				LaCloche Lowland		Deane-Tovell Saddle			
BP	cal BP		Dalles	Whitefish	Little Current	Yeo-Fitzwilliam	Lucas	Echo	Head Mackinac River
0	0	204[2]	178[3]	180[3]	171[3]	138[4]	144[4]	145[4]	131[4]
6000	6840	158.2	144.5	148.7	143.4	117.9	123.7	125.5	115.2
7000	7840	141.4	132.1	137.2	133.2	110.4	116.3	118.4	109.4
8000	8890	118.0	115.0	121.2	119.2	100.0	106.0	108.4	101.4
9000	10 190	78.2	85.8	93.9	95.2	82.3	88.5	91.5	87.7
10 000	11 470	22.7	45.2	55.0	61.7	57.6	64.1	67.9	68.6
11 000	12 920	-68.4	-21.6	-6.4	6.8	17.1	24.1	29.1	37.2

1. Computed using Equation 9.

2. Sill elevation; full discharge water level is 9 m. higher. Estimated from Lewis (1969).

3. Drom Lewis and Anderson (1989).

4. This study

levels through the Post-Algonquin phases is recorded further by additional, but somewhat delayed, organic sediment accumulation in small basins south of Georgian Bay (5, 6), near North Bay (24, 25) and on Manitoulin Island (8).

11 000-10 000 cal BP

Relative lake levels continued to fall toward the level of driftwood deposited in Georgian Bay (31b). During the lowstand, trees grew in the Straits of Mackinac (28b), and organic sediments accumulated at relatively low-elevation at northern and southern Huron basin sites (29c, 30f). A subsequent abrupt lake-level rise and inundation is recorded by a shift to more aquatic species in Georgian Bay peat at site 7 (Lewis and Anderson, 1989). The lake level rise at about 10 800 cal BP (9500 BP) was possibly in response to the onset of outburst floods from upstream glacial Lake Agassiz through its eastern outlets to Superior basin (Clayton, 1983; Teller and Thorleifson, 1983; Farrand and Drexler, 1985; Lewis and Anderson, 1989; Breckenridge *et al.*, 2004; Breckenridge and Johnson, 2005). The rise of lake level during high-discharge events may be related to constrictions in the Mattawa River valley at the Rutherglen Moraine (Harrison, 1972; Chapman, 1975) and in Ottawa River valley at the Rankin Constriction (Lewis and Anderson, 1989) (Fig. 1). The maximum of the lake-level rise possibly exceeded the altitude of the Sheguiandah site on Manitoulin Island (9c, 9d) where swamp organic sediment began accumulating just after the lake-level rise. Some sites with thick clay beneath gyttja (41, 33) suggest the waters of this inundation carried abundant fine-grained sediment in suspension. This pulse of higher water (Lake Mattawa) was apparently short-lived, and its decline may account for the emergence of small basins and onsets of gyttja sedimentation or peat accumulation at several sites (9c, 9d, 29c, 30b, 30c, 30d, 30f, 33, 34b, 38, 40, 41). The lake level drop was low enough to expose the eastern Huron basin lakebed for tree growth at site 66. Within a century or so, rising water level allowed for deposition of driftwood (64 then 32c), and terminated peat accumulation at slightly higher sites in southern Huron (39) and Georgian Bay (7a) basins.

10 000-9000 cal BP

The rising lake level that terminated peat accumulation (7a) continued to rise to a second high-level Lake Mattawa phase (about 9800 cal BP, 8700 BP) that inundated Pure Lake (50), and possibly culminated at a high enough level to induce sediment aggradation in Amable du Fond River (49) and to construct a barrier lagoon north of Lake Huron (43) (Fig. 10B). Thick clay deposits beneath gyttja in Pure Lake suggest that this water-level rise was also accompanied by a high influx of fine-grained suspended sediment. An extended period of moderately low water level following the Mattawa highstand is indicated by tree growth on the saddle between Huron and Georgian Bay basins (67, 68a, 68b), and by peaty marsh sediments in southwestern Huron basin (30a) and South Bay, Manitoulin Island (42a, 42b, 42c). Oddly, an erosional event and sequence boundary in the deep water sediments of Huron basin is not recorded at this time of low water level (Fig. 10; Moore *et al.*, 1994; Rea *et al.*, 1994a).

At about 9050 cal BP (8100 BP) water levels again rose, transforming sedimentation in South Bay from plant detritus to clastic silty clay (42a, 42b), and possibly overflowing the North Bay outlet where delta construction occurred at nearby Trout Mills (48). Water levels then rose abruptly to the last Mattawa highstand, likely as a result of overflow from large volumes of subglacial meltwater discharged during the Nakina ice advance over the Great Lakes-Hudson Bay divide, and recorded throughout the Superior basin as a set of 36-40 thick varves dated 9035 ± 170 cal BP or about 7950-8250 BP (Breckenridge *et al.*, 2004). Although the maximum lake rise in Huron and Michigan basins is not known, it apparently fell short of the Sheguiandah archeological site on Manitoulin Island as it is not recorded in the peat sequence there between dated samples 9b, 9c and 9d (Anderson, 2002). Land emergence following this short-lived high Mattawa phase lake

A

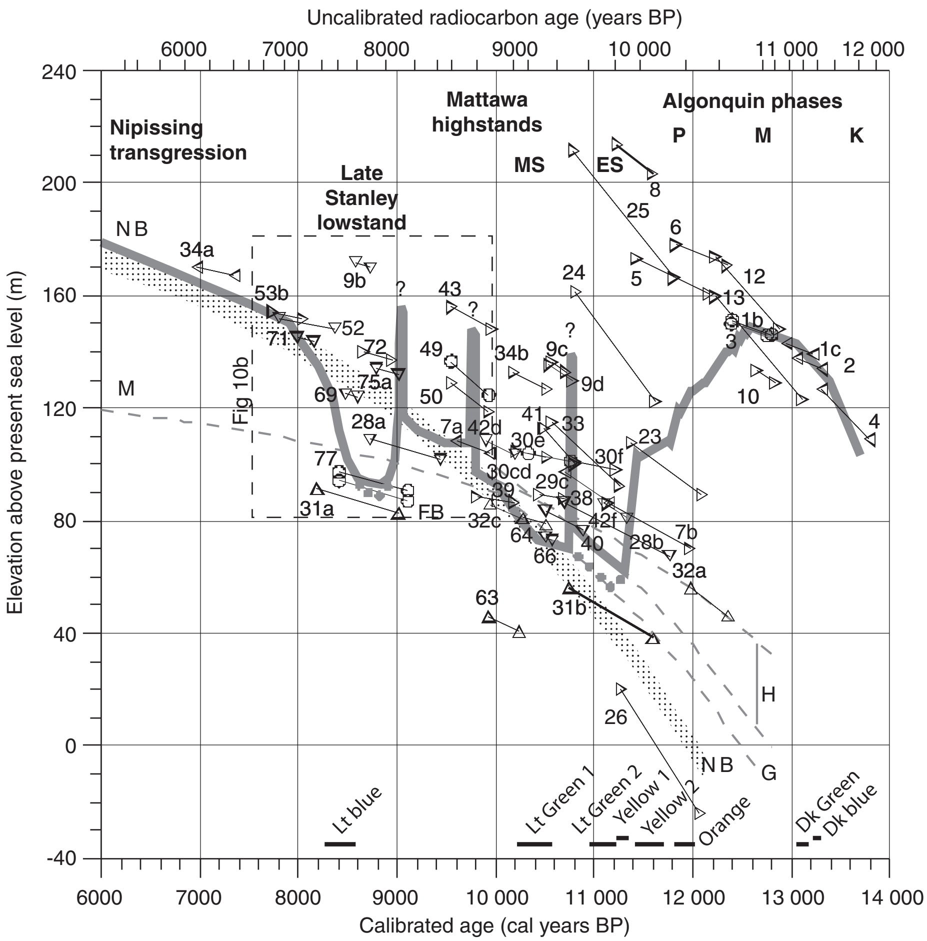

FIGURE 10A. Plot of original elevations of numbered water-level indicators (Table II) in Huron and Georgian Bay basins relative to area B (Fig. 5), inferred lake surface, and potential overflow outlet elevations (Table III) vs. age in calibrated and radiocarbon years BP. Downward-pointing triangles indicate data implying a lake level below the plotted site, as for *in situ* tree stumps. Upward-pointing triangles are for data implying a lake level above the plotted site, as for shallow water mollusks or driftwood. Right-pointing triangles indicate data implying isolation of a small basin by a falling large lake level. Left-pointing triangles indicate data implying a transgression by a large lake. See Figure 10B for a plot of all data in the 10 000-7500 cal BP interval. The inferred lake level for Huron basin is shown by the thick continuous line (dotted for Georgian Bay basin where different). The elevation history of potential overflow outlets (Table III) is portrayed by a dotted band (North Bay–NB) or by dashed curves (Huron basin sills–H, Georgian Bay basin sill–G, and Michigan basin–M). The Algonquin lake phases are Kirkfield Algonquin (K), Main Algonquin (M), and Post-Algonquin (P). MS and ES are middle and early Stanley lowstands in Huron basin, respectively. In Georgian Bay basin, the equivalent lowstands are named Hough. The horizontal lines above the X-axis, named by colours, indicate age and duration of lowstand sequence boundaries inferred from seismic reflection and core data from Moore *et al.* (1994) and Rea *et al.* (1994a).

Diagramme de l'altitude initiale des indicateurs des anciens niveaux lacustres (tabl. II) pour les bassins du lac Huron et de la baie Géorgienne relativement à la région B (Fig. 5), où les points donnent l'altitude des exutoires (tabl. III), l'âge en années étalonnées et conventionnelles. Les triangles pointant vers le bas font référence à un niveau lacustre inférieur au site reporté, comme dans le cas des souches d'arbres in situ. *Les triangles pointant vers le haut font référence à un niveau lacustre supérieur au site reporté, comme dans le cas des mollusques en eau peu profonde ou de bois flottant. Les triangles pointant vers la droite font référence à l'isolement d'un petit bassin suite à baisse du niveau d'un grand lac. Les triangles pointant vers la gauche font référence à la transgression d'un grand lac. Voir la figure 10B pour le diagramme des données comprises dans l'intervalle 10 000-7500 cal BP. Le niveau lacustre reconstitué du lac Huron est représenté par un trait gras continu (en discontinu pour la baie Géorgienne lorsqu'il diffère). L'historique de l'altitude des exutoires potentiels (tabl. III) est représenté par une bande en pointillé (North Bay–NB) ou par les courbes en tireté (seuils du bassin du lac Huron–H, seuil de la baie Géorgienne–G et bassin du lac Michigan–M). Les phases du Lac Algonquin sont les suivantes : Kirkfield Algonquin (K), Algonquin principal (M) et Post-Algonquin (P). MS et ES identifient respectivement les niveaux moyen et ancien de Stanley du lac Huron. Dans le bassin de la baie Géorgienne, les niveaux équivalents sont nommés Hough. Les lignes horizontales au-dessus de l'abscisse, nommées par une couleur, indiquent l'âge et la durée des limites déduites à partir de données de réflexion sismique et de forage issues des travaux de Moore* et al. *(1994) et de Rea* et al. *(1994a).*

B

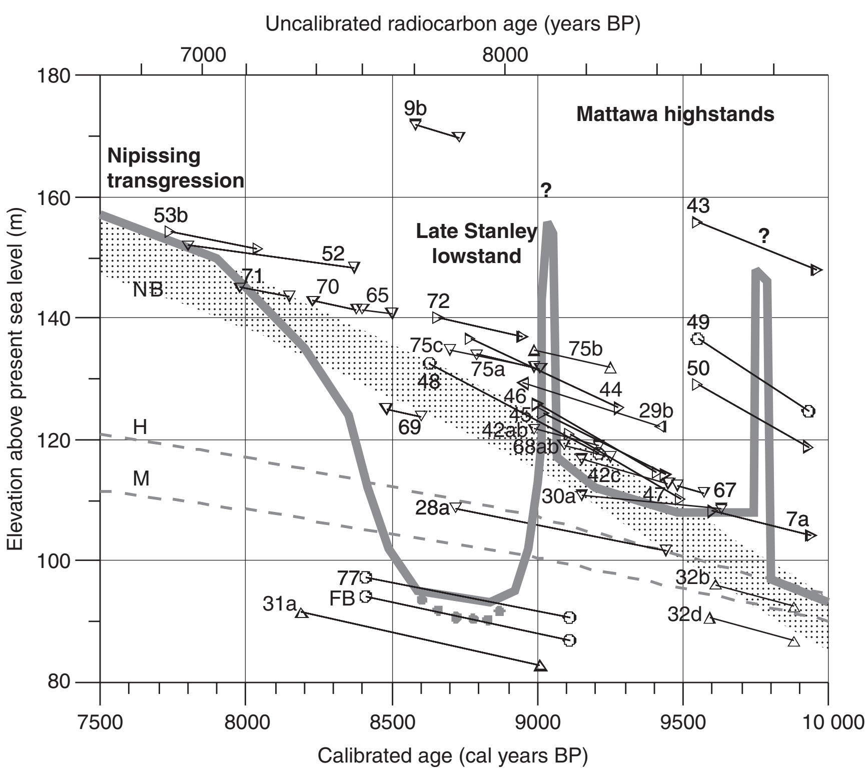

FIGURE 10B. Expanded plot of original elevations of water-level indicators for the 10 000-7500 cal BP period in Huron and Georgian Bay basins relative to area B (Fig. 5), inferred lake surface, and potential overflow outlets vs. age in calibrated and radiocarbon years BP. Symbols as described in Figure 10A.

Diagramme de l'altitude initiale des indicateurs des anciens niveaux lacustres pour la période 10 000-7 500 cal BP pour les bassins du lac Huron et de la baie Géorgienne relativement à la région B (Fig. 5), où les points donnent l'altitude des exutoires, l'âge en années étalonnées et conventionnelles. Les symboles sont les mêmes qu'en A.

is documented by onsets of organic sedimentation at Wye Marsh, southern Georgian Bay (72), gyttja deposition in small basins northeast of Georgian Bay (44, 45, 46, 47), and tree growth at low elevations now under water in Huron basin (75a, 75c) and possibly in the Straits of Mackinac (28a).

9000-7000 cal BP

Lake-level decline from the last high Mattawa phase continued down to levels defined by the Stanley unconformity (77) in Huron basin (Lewis *et al.*, in press) and the Flowerpot beach (FB) in Georgian Bay basin (Fig. 10). This lowstand phase, which continued for several centuries, formed the Light Blue reflector and sequence boundary in deep water sediments (Moore *et al.*, 1994; Rea *et al.*, 1994a), and allowed tree growth on the Deane-Tovell saddle between Huron and Georgian Bay basins (65, 69, 70). Lake-level rise from this low phase by 7.5 ka BP (8.3 cal ka BP) is suggested by the chronology of the Light Blue reflector in northern Huron and Georgian Bay basins (Moore *et al.*, 1994).

The final increase in lake levels from the low phases of late Lake Stanley and late Lake Hough in Huron and Georgian Bay basins, respectively, probably occurred about 8000 cal BP, late enough to allow tree growth in eastern (71) and western (52) Huron basin, but early enough to transgress the Rains Lake site (53b) in northwestern Huron basin. At this elevation, the water surface had risen to the level of the Nipissing beach at North Bay, indicating that the lake in Michigan-Huron-Georgian Bay basins was overflowing the North Bay outlet at full discharge. As this outlet uplifted faster than other parts of the lake basin, relative lake levels rose throughout most of the upper Great Lakes basins as the well-known Nipissing transgression, manifested in the present data by transgression of the Rains Lake (53b) and Smoky Hollow Lake (34a) sites (Fig. 10A).

INTERPRETATION OF WATER-LEVEL HISTORY OF THE MICHIGAN BASIN

Because of their connection via the Indian River lowland and Straits of Mackinac (Fig. 1), high water levels were always at common elevations in the Huron and Michigan basins, as at present, following retreat of ice from their northern regions (Eschman and Karrow, 1985; Hansel *et al.*, 1985). Consequently, the history of lake-level variation above the Michigan basin sill defined in the Huron basin is applicable to the Michigan basin as shown in Figure 11. Only evidence of lowstands and other unique indicators of Michigan basin lake elevations are shown and discussed here.

Submerged *in situ* tree stumps in southern Lake Michigan (the Olson Forest of Chrzastowski *et al.*, 1991) have been traditionally interpreted as having been drowned in the Nipissing transgression (Chrzastowski and Thompson, 1992, 1994; Colman *et al.*, 1994b). However, in this analysis, the Olson tree stumps (sites 74a, 74b at 8.2-8.4 ka BP, 9.2-9.4 cal ka BP, in Figs. 1 and 11) appear tens of metres above the North Bay

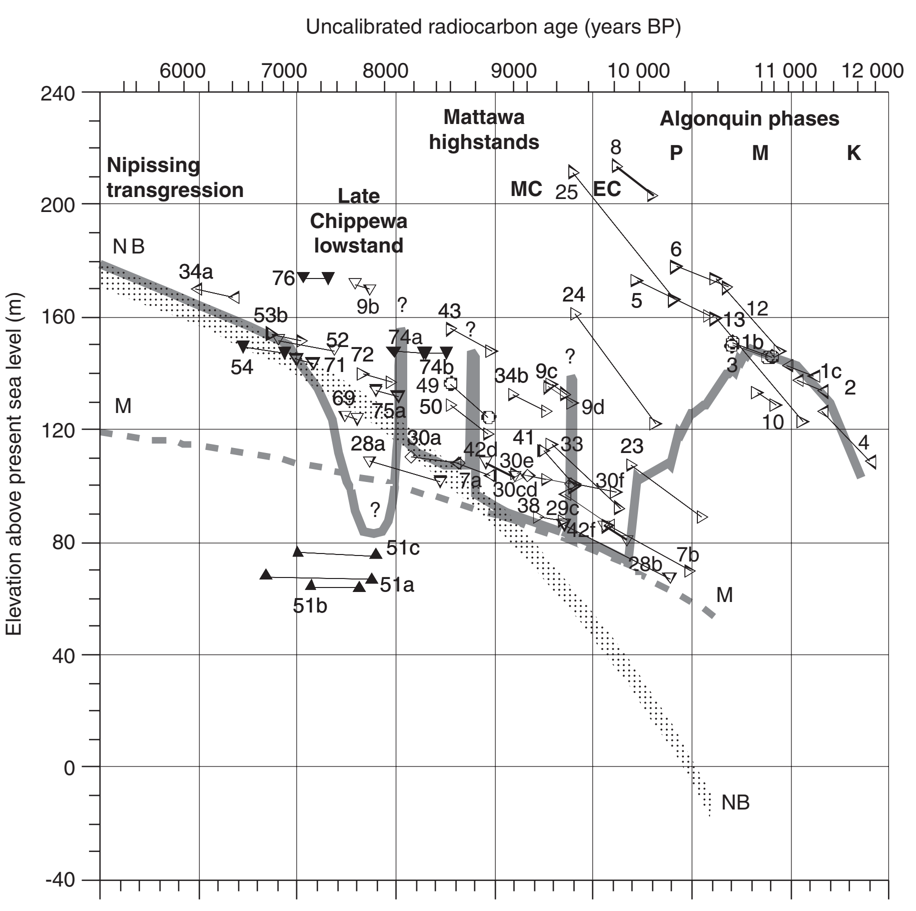

FIGURE 11. Michigan basin plot of original elevations of water-level indicators relative to area B (Fig. 5), inferred lake surface, and potential overflow outlets vs. age in calibrated years BP and radiocarbon age in ka BP. Water-level indicators for the Huron basin above the Michigan sill are applied to the Michigan basin as these highstand lakes were confluent in both basins. Black symbols indicate data specific to the Michigan basin. MC and EC represent middle and early Chippewa lowstands, respectively. Otherwise symbols and lines as for Figure 10A.

Diagramme de l'altitude initiale des indicateurs des anciens niveaux lacustres pour le bassin du lac Michigan relativement à la région B (Fig. 5), où les points donnent l'altitude des exutoires (tabl. III), l'âge en années étalonnées et conventionnelles. Les indicateurs du niveau lacustre du lac Huron situés au-dessus du seuil Michigan sont appliqués au lac Michigan puisque ces plans d'eau élevés étaient connectés aux deux bassins. Les symboles noirs font référence aux données spécifiques du lac Michigan. MC et EC représentent les niveaux moyen et ancien de Chippewa. Les autres symboles et les lignes sont identiques à ceux de la figure 10A.

outlet, and could not be affected by the Nipissing transgression. A rise of water level to the final highstand of Lake Mattawa follows the Olson Forest after 8.2 ka BP (Fig. 11). This flooding event seems to have been the cause of forest drowning.

As in the Huron and Georgian Bay basins, an extreme decline in lake level following the last Mattawa highstand is suggested in the Michigan basin by the occurrence of shallow-water sediments and molluscan fauna in the deepwater environment of central Lake Michigan (51a, 51b, 51c) (Lewis *et al.*, in press). A final recovery of lake level to an overflowing condition at the North Bay outlet probably occurred about 8.1-7.8 cal ka BP, allowing for tree growth in northern Michigan basin (54). From this time forward, relative lake level rose under control of the North Bay outlet as the Nipissing transgression.

CLOSED LOWSTAND CONDITIONS

During the period 8.95-8.3 cal ka BP (8.05-7.4 ka BP) lake levels (late Stanley in Huron basin, late Hough in Georgian Bay basin, and late Chippewa in Michigan basin) were below the sill of the North Bay outlet, the lowest possible overflow outlet at the time (Figs. 10B-11). Thus the late Stanley, late Chippewa, and late Hough phases were hydrologically closed lakes at their lowest level, and as such may have resulted from the impact of a severe dry climate in which evaporative water losses exceeded water inflows by precipitation and runoff. The inference of dry climate is supported by the presence of thecamoebians that indicate a more saline lake environment, and by climate transfer function analysis of pollen assemblages which suggests less precipitation and warmer temperatures at 7.7 ka BP than today in the Georgian Bay area (Blasco, 2001). Other lowstands around 9.3 and 9.8 ka BP (10.5 and 11.2 cal ka BP) may have also been closed briefly.

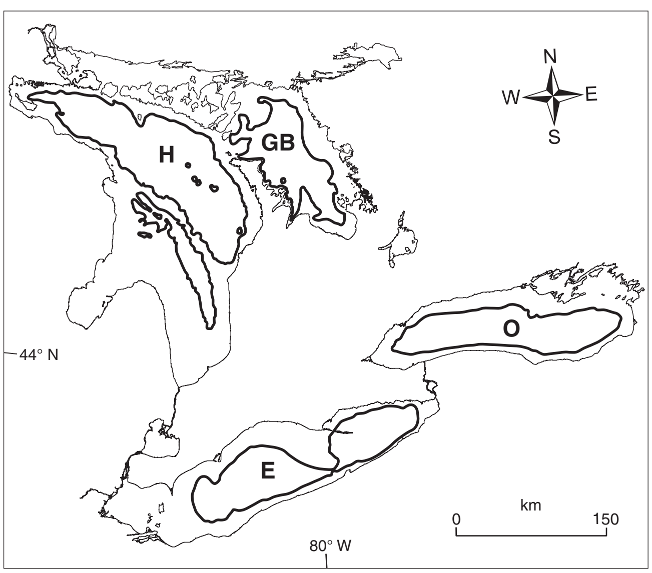

FIGURE 12. Paleogeographic lowstand map of Huron, Erie and Ontario basins at about 7.8 ka BP (8.6 cal ka BP) showing the reduced area of inferred closed (terminal) lakes whose shorelines are well offshore from the present lake shores. E = Erie basin, GB = Georgian Bay basin, H = Huron basin, and O = Ontario basin.

Carte paléogéographique du bas niveau lacustre à environ 7,8 ka BP (8,6 cal ka BP) pour les bassins des lacs Huron, Érié et Ontario. La carte montre la superficie réduite des lacs fermés dont les anciens rivages étaient bien au large des rivages actuels. E = bassin du lac Érié, GB = bassin de la baie Géorgienne, H = bassin du lac Huron et O = bassin du lac Ontario.

Possible shorelines for the dry climate-induced closed lowstands at 7.8 ka BP (8.6 cal ka BP) are illustrated in a paleogeographic reconstruction of the Huron, Erie and Ontario basins on Figure 12. The northern Huron lowstand shore is tied to the Stanley unconformity beneath northwestern Lake Huron (Hough 1962; Lewis *et al.*, in press). The Georgian Bay lowstand shore is tied to the low-level Flowerpot beach in the entrance to Georgian Bay (Blasco, 2001). Water bodies in these basins and those beneath southern Lake Huron were isolated closed lowstands, well offshore from the present lake boundaries. The Erie lowstand shore is tied to a submerged beach in eastern Lake Erie (Fig. 9E; Coakley and Lewis, 1985; Lewis *et al.*, 2004) at a lake level that would have extended into the central Lake Erie area before that sub-basin became mostly infilled with sediment (Sly and Lewis, 1972). The Ontario basin is assumed to have been impacted by the same dry climate that affected the other basins. Under these conditions, its lowstand shoreline was inferred at a level that made its basin area-to-lake area ratio equal to that of the Huron basin (Bengsston and Malm, 1997).

DISCUSSION

GEOPHYSICAL MODELS OF ISOSTATIC ADJUSTMENT

With knowledge or estimates of the elastic and viscous properties of Earth's lithosphere and mantle, respectively, geophysical models compute the crustal isostatic response following a known or inferred history of ice sheet loading and pattern of deglaciation. Clark *et al.* (1994) used this approach to evaluate isostatic movements during deglaciation of the Great Lakes basin. They successfully illustrated the increasing amplitude of postglacial isostatic rebound towards the north and northeast in the direction of past thicker ice and ice retreat in accordance with the evidence of deformed paleo-lake shorelines. They applied an innovative approach to constrain and calibrate their model by using geological knowledge of large-lake drainage transfers from northern to southern outlets as basins were differentially tilted during the Algonquin and Nipissing phases. Lake-level history was derived by tracking the upward movement of overflow sills through time. Lake-level indicators were not tracked separately from the sills with the result that possible episodes of hydrologic closure could not be detected.

Gravitationally self-consistent models of glacio-isostatic adjustment on a global scale have been progressively developed and improved over the past few decades (Peltier, 1998). These models are constrained and calibrated to relative sea-level histories at ocean-continent boundaries. Model estimates of isostatic adjustment are produced for continental interiors, and one of these, the ICE-3G model (Tushingham and Peltier, 1991), has been compared favourably with short-term evidence of tilting of the Great Lakes basins derived from trend analysis of lake-level gauge records (Tushingham, 1992).

However, over the longer term, the geophysical model results are not in complete agreement with past events in the watershed. The relative performances of two geophysical models and the empirical model were assessed by comparing the elevations of the northern and southern outlets from the upper Great Lakes basins during the Algonquin and Nipissing drainage transfers which are known to occur at about 10.5 ka BP and 5 ka BP, respectively, from independent geological evidence (Karrow *et al.*, 1975; Karrow, 1980; Monaghan *et al.*, 1986). Outflows were transferred from the Kirkfield outlet to the Port Huron and possibly Chicago outlets during the Algonquin transfer (Fig. 7A), and from the North Bay to the same southern outlets during the Nipissing transfer. Estimates of the elevations of the overflow sills and timing for these drainage transfers when northern outlets rose isostatically above southern outlets were provided for the ICE-4G glacio-isostatic model (Peltier, 1995, 1996) courtesy of W.R. Peltier. Differences between Port Huron and the other outlets which should have been near zero at the times of the drainage transfers are shown in Figure 13. The empirical model performs best with northern and southern sills being within two metres of each other during transfers. The Clark *et al.* (1994) model is similar, although the best agreement for the Algonquin transfer was obtained at 10 ka BP, rather than the expected 10.5 ka BP, and the Nipissing-age transfer at Chicago differed by 3 m. Variances in sill elevations during transfers for the ICE-4G model are larger, and ranged from 3.5 to 10 m.

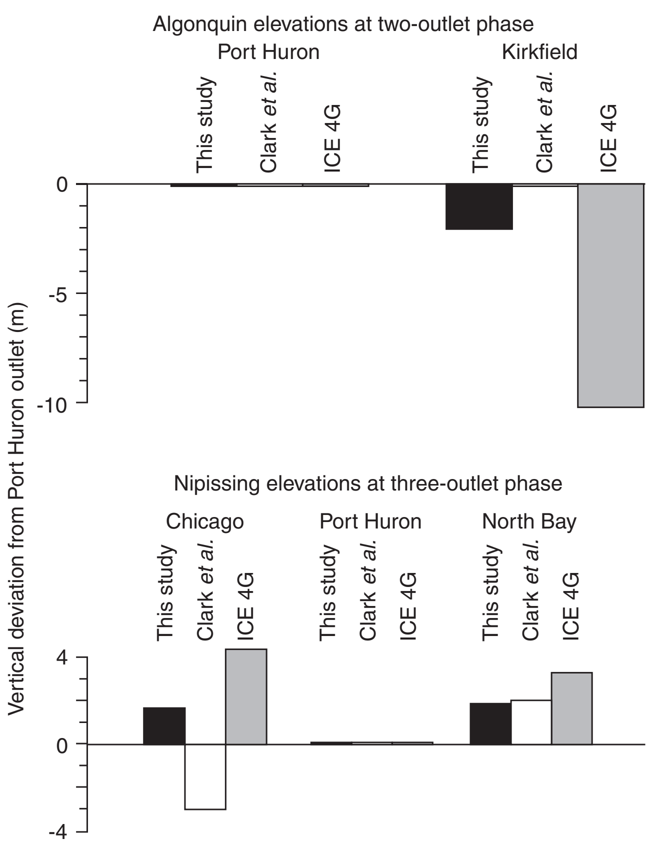

FIGURE 13. Deviations of sill altitudes when they should have been of equal elevation during drainage transfers from northern to southern outlets of the Algonquin and Nipissing lake phases for the Clark *et al.* (1994) and ICE-4G geophysical models and the empirical model used in this paper. Sill elevations, reduced relative to Port Huron at the southern end of the Huron basin, show that the empirical uplift model used in this study best describes (lowest deviations) the relative elevations of outlets during drainage transfers.

Écarts entre les altitudes des seuils lorsqu'elles devraient être égales aux autres durant les transferts du drainage des exutoires septentrionaux aux exutoires méridionaux des phases lacustres Algonquin et Nipissing pour les modèles géophysiques de Clark et al. *(1994) et de ICE-4G et le modèle empirique utilisé dans cette étude. Les altitudes des seuils, ajustées à celles du Port Huron à l'extrémité sud du bassin du lac Huron, montrent que le modèle empirique de soulèvement glacio-isostatique utilisé dans cette étude rend mieux compte (écarts plus faibles) des altitudes relatives des seuils durant les transferts de drainage.*

Near and beyond the maximum margins of the ice sheets, the geophysical models predict crustal uplift as a forebulge of modest relief compared with the amplitude of depression beneath the centre of the ice load. This has been demonstrated in ice-marginal Atlantic and Arctic coastal regions (Barnhardt *et al.*, 1995; Dyke, 1998). The forebulge effect has not been recognized from empirical evidence in the continental region adjoining or south of the Great Lakes basin. However, ice marginal areas in this region could have undergone uplift and subsidence associated with the growth, migration and decay of a glacio-isostatic forebulge (Colman *et al.*, 1994a), and some evidence for subsidence exists, for example, anomalies in modern tilting of the Erie and southern Michigan basins (Mainville and Craymer, 2005).

Benefits of geophysical models are their ability to compute both vertical and horizontal earth movements associated with the isostatic process, and to relate these estimates to an absolute datum such as present sea level. Confidence in predictions of past crustal movements in the Great Lakes region by geophysical models, especially deformation over several millennia, will be increased markedly when these models are calibrated and constrained by the empirical observations of glacio-isostatic adjustments within the same region.

PREVIOUS RECOGNITION OF LOW LAKE LEVELS IN HURON BASIN

An extreme lowering of lake level in the Huron and Michigan basins after the Algonquin phases was predicted by Stanley (1936) when the ice sheet margin receded from the isostatically-depressed lowland drainage route past North Bay, Ontario, to the Mattawa and Ottawa river valleys. Sediment unconformities, discovered by Hough (1955, 1962), confirmed lake-level lowstands in each of the Michigan and Huron basins; these lowstands were named Chippewa and Stanley, respectively. Hough (1962) inferred a delay in isostatic adjustment so that, in his model, the North Bay sill remained at its depressed Algonquin level, and Lake Stanley could be interpreted as an open overflowing water body (Lewis *et al.*, in press).

Although a regional advance (Marquette readvance) of the Laurentide Ice Sheet about 500 km wide reached the southern coast of the Superior basin at 10 ka BP (11.5 cal ka BP) (Lowell *et al.*, 1999), and ice retreat was slowed during the Younger Dryas in eastern Ontario and western Québec (Simard *et al.*, 2003), there is no evidence of ice advance or of delayed isostatic recovery in the Georgian Bay-North Bay outlet region, about 300-500 km east of the Superior basin (Dyke *et al.*, 2003). As a result, it is highly unlikely that rebound was delayed significantly and, at the time of late Lake Stanley (Fig. 10), the recovered outlet was above water level, as was also interpreted by Lewis *et al.* (in press). Closed lake conditions for late Lake Stanley are also consistent with the preliminary paleoecological findings based on thecamoebian analysis of somewhat higher salinity in late Lake Hough, the equivalent closed water body in the Georgian Bay basin, as reported by Blasco (2001).

WORKING HYPOTHESIS FOR CLIMATE CHANGE AND THE HYDROLOGICALLY CLOSED LOWSTANDS

Major shifts in Great Lakes water levels have long been understood in terms of overflowing lakes. Lake elevations change as a result of shifting ice dams during glacial retreat or advance, outlet erosion, or by differential glacio-isostatic adjustment of the outlets (Eschman and Karrow, 1985; Hansel *et al.*, 1985; Larsen, 1987; Barnett, 1992; Larson and Schaetzl, 2001). These mechanisms apply to the Algonquin highstand and the transgression to the Nipissing Great Lakes. The intervening Mattawa high phases (Lewis and Anderson, 1989) are also considered to be overflowing lakes, but with variable levels controlled by resistance at hydraulic constrictions downstream of North Bay to variable high-discharge flows from upstream Lake Agassiz or subglacial drainage.

The discovery of low lake levels below the lowest possible overflow outlet in their basins implies a controlling factor of excess evaporation (water loss) over precipitation and runoff (water supply) during a period of dry climate. These lowstands, the late Chippewa, late Stanley, and late Hough lakes in the Michigan, northern Huron, and Georgian Bay basins, respectively, are thought to reflect a phase of climatically-driven hydrologic closure that has not been recognized previously. The earlier pre-9.5 ka BP (pre-10.9 cal ka BP) lowstands, middle and early lakes Chippewa, Stanley and Hough, are inferred to be close in level to, or slightly below, their basin outlets, and thus were possibly also subject to enhanced evaporative losses of water. During these early, low lake phases, sustaining upstream inflows were not available, as Lake Agassiz discharge was diverted away from the Great Lakes by advances of ice in the Superior and Nipigon basins (Thorleifson and Kristjansson, 1993; Lewis *et al.*, 1994). Dry air from the glacial atmospheric circulation over the nearby Laurentide Ice Sheet (Bryson and Wendlund, 1967; David, 1988; Anderson and Lewis, 2002; Wolfe *et al.*, 2004) probably exerted continuous evaporative stress on nearby water surfaces, causing drawdown of these lakes during phases of reduced inflow.

By the time of the latest and longest-duration lowstands (late Chippewa, Stanley and Hough) the effects of glacial atmospheric circulation would have diminished greatly owing to the small size of the remaining Laurentide Ice Sheet, and to its more distant location relative to the Great Lakes basin. At this time (8 ka BP, 8.9 cal ka BP), meltwater drainage from the merged glacial lakes Agassiz-Ojibway began bypassing the upper Great Lakes basins directly into the Ottawa Valley (Veillette, 1994; Teller *et al.*, 2002; Teller and Leverington, 2004). Thus, the Great Lakes watershed suddenly lost a significant source of water supply and was susceptible to the early Holocene dry climate (Edwards *et al.*, 1996).

Atmospheric water supply today can be thought of as a function of the relative time spent over the Great Lakes basin by three major air masses over North America. These masses are the Arctic air from the north (dry and cold), the Pacific air from the west (dry and warm), and the Maritime Tropical air bringing moist, warm air north from the Gulf of Mexico (Bryson and Hare, 1974; Bradbury and Dean, 1993). Once glacial lake drainage began bypassing the Great Lakes, and as the Laurentide Ice Sheet downwasted over Hudson Bay, southward incursions of dry Arctic air, previously blocked by the high Laurentide Ice Sheet, were likely becoming increasingly frequent (Yu and Wright, 2001), and this dry air may have initiated or intensified draw down of lake levels to the late Chippewa, Stanley, and Hough lowstands. The closed lowstands may have been maintained by enhanced evaporation into increasingly strong flows of dry, warm Pacific air from the west. These flows are indicated by abundant evidence of vegetation shifts to drought-resistant plant taxa west of the Michigan basin (Baker *et al.*, 1992; Wright *et al.*, 2004). By 7 ka BP (7.8 cal ka BP), increasing incursions of the Maritime Tropical air mass were delivering sufficient precipitation, indicated by the appearance of mesic forest species in pollen diagrams of the Great Lakes region (Webb III *et al.*, 1998) to convert the Michigan, Huron and Georgian Bay water bodies to open, overflowing lakes, as at present. These changes are consistent with paleovegetation maps which show a rapid northward migration through the Great Lakes region of the mixed-boreal forest biome boundary between 8.0 and 7.0 ka BP (8.9 and 7.8 cal ka BP) (Dyke *et al.*, 2004).

The foregoing hypothesized climatic history is consistent with changes inferred for small Elk Lake in Minnesota, based on a comprehensive study of proxy climatic and limnological indicators (Bradbury and Dean, 1993). Similarly, this climatic history is supported by a coeval depletion in the ^{18}O composition of precipitated carbonate in Deep Lake, Minnesota, attributed by Yu and Wright (2001) to the blocking of southern air masses by more frequent presence of dry Arctic air.

The phase of reduced and closed lakes at the onset of the present hydrologic regime offers a unique opportunity to evaluate the sensitivity of the Great Lakes system to high-amplitude climatic change. Understanding the sensitivity of the lakes to high-amplitude and long-duration change would be a distinct benefit in the light of the need to project and adapt to future changes under global warming which may drive the lakes below instrumentally-observed variability (Mortsch *et al.*, 2000).

ISOTOPIC COMPOSITION OF LAKE WATER

As the evaporation process favours concentration of water molecules containing the heavier ^{18}O isotope, previous findings of high concentrations of the lighter ^{16}O isotope, similar to that of glacial meltwater, in fossil valves of Huron basin benthic ostracodes at about 7600 ka BP (Rea *et al.*, 1994a, 1994b; Dettman *et al.*, 1995) appear to contradict the coeval presence of evaporatively-driven lowstands as postulated in this paper and in Lewis *et al*. (in press). Although surface lake water isotopic composition undoubtedly became concentrated in ^{18}O during seasonal periods of rapid evaporation, it apparently did not greatly influence bottom water composition, similar to the isotopic stratification found for glacial Lake Agassiz (Buhay, 1998; Birks *et al*., in press). Alternatively, adjustments in the chronology of isotopic events in the Huron and Michigan basins suggest that the low ^{18}O inflow occurred prior to the lowstands (Breckenridge and Johnson, 2005; Breckenridge, in press), and may have remained as bottom water during the evaporative phase. Additional research is needed to resolve the origin of bottom water in the Huron and Michigan basins.

SUMMARY

Postglacial isostatic adjustment in the Great Lakes region is described here in the time domain using an exponential decay expression constrained by the observed cumulative differential deformation of dominant paleo-lake strandlines in individual or groups of basins. Vertical earth movement throughout and following the last deglaciation into the middle Holocene is characterized as progressive, differential uplift relative to an area southwest of southern Lake Michigan, beyond the limit of the last glacial maximum. Rates and amplitudes of uplift increase towards the north-northeast in the direction of deglacial retreat and thicker ice. A mean relaxation time for the uplifting process of 3700 ± 700 years was obtained by averaging solutions for the decay time parameter at 20 transects throughout the basin where isobases (gradients) of two strandlines of different age were known. This relaxation time was used in the exponential expression to adjust the isobase gradients of dominant strandlines in the Great Lakes basins to 10.6 ka BP (12.6 cal ka BP), the approximate age of the well-known Algonquin phase.

Collectively, the Algonquin and adjusted isobases constitute a reference response surface for isostatic adjustment throughout the Great Lakes region. Interpolated values from this surface and the mean relaxation time were used in the exponential uplift expression to determine an ‘amplitude’ factor for uplift at any desired location. With these values and the exponential expression, uplift since any desired age could be computed. The original elevation of a site at any desired age could also be determined by subtracting the computed uplift from the present elevation of the site. This approach was used to transform values of pixels in a DEM for the present Great Lakes region to new DEMs at previous ages. A total of 12 paleogeographic reconstructions for the topography and bathymetry of the Great Lakes basins were prepared for ages between 11.4 and 5.0 ka BP (13.3 and 5.7 cal ka BP). These reconstructions showed that water surface areas ranged from +72% to -95%, and lake volumes from +200% to -97%, relative to the present lakes. Improvements in the estimation of glacio-isostatic effects can be expected from geophysical models when these are calibrated to the available observations of differential rebound in the Great Lakes basin.

The same empirical approach was applied to reconstruct the original elevations of dated indicators of former lake levels in the Michigan, Huron and Georgian Bay basins for the interval between 11.7 and 6.2 ka BP (13.5 and 7.1 cal ka BP). Original elevations for 79 dated indicators, comprising fossils from beach lagoon sediments, basal organic sediment from isolation basins, shallow-water fossils in unconformable zones within deepwater sediment, and submerged tree stumps in growth position, and others, were reconstructed to form the basis for interpreting lake-level history in these basins. In parallel fashion, the history of potential overflow sills was also reconstructed. Comparison of sill elevations with lake levels revealed a period (about 8.05 to 7.4 ka BP, or 8.95 to 8.3 cal ka BP) in which water surfaces were up to several tens of metres below lowest possible overflow outlets. The low lake levels are postulated to reflect the increased impact of early Holocene dry climate when upstream Agassiz overflow and/or subglacial floods were diverted around the Great Lakes basin, and water supply to the upper Great Lakes was reduced. Lake evaporation and draw down were likely enhanced by frequent incursions of dry cold Arctic air and, later, warm dry Pacific air, possibly related to atmospheric reorganization associated with the demise of the Laurentide Ice Sheet. This period of closed lakes, early in the present hydrological regime of the Great Lakes, offers an opportunity to probe and understand the sensitivity of the Great Lakes system to high-amplitude, long-duration climate change. Such information could improve confidence in projections of, and adaptations to, future levels of the Great Lakes under global warming.

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