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Proceedings of the Great Lakes Paleo-Levels Workshop: The Last 4000 Years

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NOTICE

The Workshop was held at NOAA's Great Lakes Environmental Research Laboratory in Ann Arbor, MI, April 13-14, 1999, and was sponsored by GLERL and the US Army Corps of Engineers, Detroit District.

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INTRODUCTION

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The Great Lakes, Figure 1, are one of North America's largest water resource systems with a basin area of about 770,000 km^2, of which about one third is lake surface. It is one of the most intensively used freshwater systems in the world, serving multiple interests including navigation, hydropower, recreation, water supply, food supply, and riparian. The outflows from Lakes Superior and Ontario are regulated by regulatory works in the St. Marys and St. Lawrence Rivers, respectively. The remainder of the system is naturally regulated through the large surface areas and limited outflow capacity. Great Lakes water levels change slowly due to the large lake surface areas and constricted outlet channels, which integrate short-term climate fluctuations. Because of the relatively small range in lake levels, about 1.8 m, significant uses have become dependant upon small changes in water levels and outflows, resulting in system sensitivity to relatively small changes in climate variability and change.

The Great lakes water levels constitute one of the longest high quality hydrometeorological data sets in North America with master gage records beginning about 1860 with other sporadic records back to the early 1800's. However, from a longer term perspective under the current system hydraulic regime, which has been in place for about 3000 years, we have only observed/measured about 5 percent of the time series. There is a highly likely probability that we could experience runs of high and low lake levels with extremes significantly higher than our present measurements would indicate. For example, few if any envisioned in the 1960's that we would have a 30 year run of well-above-average high lake levels and set two record highs within 13 years. A longer term lake level perspective can only be obtained by examining paleo information from the geologic record, or in a more limited fashion, from stochastic hydrologic analysis based upon the historic record.

The workshop was convened to place the historical lake level measurements in a longer term perspective through a series of papers and discussion by highly qualified experts in the specialized field of Great Lakes paleo lake level reconstruction. This perspective will serve a broad range of uses from evaluation of potential Great Lakes shoreline damages to the development of more robust lake level regulation and water resource policy.

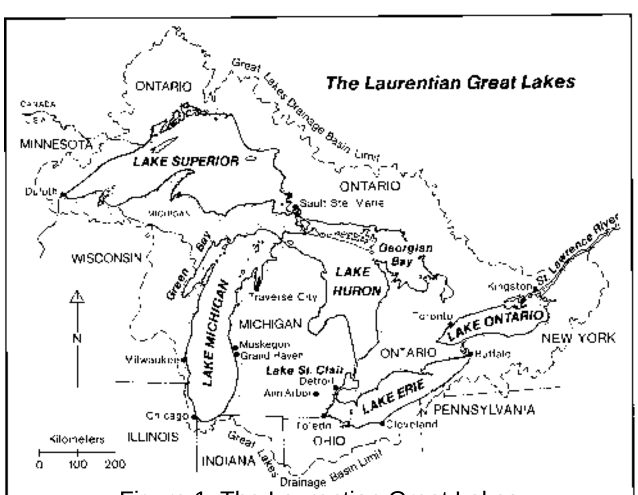

Figure 1. The Laurentian Great Lakes.

~ Paleo Lake Levels — The Last Four-Thousand Years ~

BACKGROUND

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Human perception of flooding and erosion hazards along the shores of the Great Lakes is often limited by the length of human memory and recorded past conditions. Yet, records of Great Lakes water level fluctuations cover only a little more than a century (1860 to the present). This is only a little more than 3 percent of the period that the Great Lakes system has existed with its present drainage system.

During the last 4 or 5 millennia, Great Lakes water levels have responded to geological, atmospheric, and anthropogenic changes. There have been subtle, yet progressive tectonic changes to the basin as the earth's mantle and crust have adjusted to the disappearance of the massive Pleistocene continental glacier. These geotechnical changes have both tipped and warped the basins; modifying the relative elevation of the outfalls and levels of water impoundment in each lake. The Earth's atmosphere exhibits poorly understood long and short period changes in the balance between polar and tropical air masses, the location of major pressure system lows and highs, and the jet stream. As the atmosphere evolves, meteorological cycles and trends materialize resulting in periods of greater or lesser precipitation for different regions of the earth. Within recorded time there have been seasonal, annual, and decadal alternating cycles of droughts and floods in the Great Lakes system resulting in periods of low and high water levels. In more recent times, man has modified the Great Lakes basin and drainage system through land use practices and modifications to the drainage system, including some changes to the capacity of lake inflows and outflows. Finally, there are the effects of atmospheric pollution and the unknown future associated with global warming.

Geological evidence developed by independent researchers studying different evidence in different locations throughout the Great Lakes suggest the presence of water level change trends associated with basin tectonics and longer period atmospherically-induced cycles than those captured in the recent record. Future stewardship of the Great Lakes system needs to look beyond the documented record in predicting future water level changes and potential extreme values.

In response to the 1993 International Joint Commission Great Lakes Levels Reference Study, the U.S. Army Corps of Engineers is developing a basin-scale Flooding and Erosion Evaluation System. This system includes a lake-by-lake relational database of geological, land use and hydraulic information; coastal recession, sediment and flood models; and computer-based linkages to analyze the erosion and damage consequences of changed water level and sediment supply scenarios. This interactive data base and analysis system provides those responsible for Great Lakes coastal zone planning, public policy, engineering actions, and operating water level control elements with a powerful regional management tool. The goal of this workshop is to develop rational future Great Lakes water level scenarios for use in the development and application of this regional management system.

~ Paleo Lake Levels — The Last Four-Thousand Years ~

HOLOCENE LAKE LEVELS AND CLIMATE, LAKES WINNIPEG, ERIE, AND ONTARIO

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Significant change in postglacial lake levels in areas formerly covered by the Laurentide Ice Sheet is mainly a function of ice-margin position during ice retreat, land topography, and differential glacial rebound (Hough, 1958), and climatic or hydrological change (Larsen, 1985). Additional causes of lake level change have been related to increases or decreases in drainage basin area (Lewis and Anderson, 1989), and by erosional downcutting of outlet sills (Pengelly et al., 1997). Potential neotectonic adjustments (Sanford et al., 1985; Wallach et al., 1998) are another possible source of change in relative lake levels. Of these factors, climate change is likely the most important (and perhaps the most uncertain) for consideration of future lake levels in the next few centuries.

In this presentation, I will first show results of a nearly completed study of the geological evolution of Lake Winnipeg, a large lake in southern Manitoba, about 550 km northwest of Lake Superior, that is comparable in area with the Laurentian Great Lakes. Lake Winnipeg's surface area of 24,530 km^2 is 31 percent larger than that of Lake Ontario and 4.6 percent smaller than that of Lake Erie. The Lake Winnipeg team discovered significant changes in water level owing to the effects of a mid-Holocene switch from dry to moist climate conditions. These changes were superimposed on progressive tilting of the basin by ongoing postglacial rebound (Todd et al., 1996, 1999; Lewis et al., 1998a,b, 1999). These findings have sensitized me to the importance of water balance in drainage basins for lake levels; in extreme situations, negative water balance may cause lakes to switch from an open (overflowing) state to a closed state with levels drawn down by evaporation below the basin outlet sill. Even in an open state, lake levels will fluctuate moderately with changes in water balance, as in the present hydrological regime of the Great Lakes.

From Lake Winnipeg, I will move to Lake Erie and show that similar dry climate effects (closed basin conditions) likely occurred there in the early to mid-Holocene. Then, I will identify sedimentological work by others in the Ontario basin which suggests climatically-driven oscillatory changes in lake level over the past 4000 years (Flint et al., 1988; McCarthy and McAndrews, 1988; Dalrymple and Carey, 1990). Finally, I will explore recent paleoclimate findings in the Ontario basin that late Holocene climate shifts may be explained as changes in air mass circulation, drier periods occurring because of more frequent westerly flows of dry Pacific air (Yu et al., 1997). Further research to better determine the relationship of paleoclimate to former lake level change could prove helpful in understanding the past and future regime of change in water balances and levels of the Great Lakes.

Lake Winnipeg

Lake Winnipeg consists of a small South Basin separated from a large North Basin by a constricted central area with the long axis of the 400 km-long lake oriented approximately south to north. Its catchment area (982 900 km^2) (National Atlas of Canada, 1985) includes most of the Canadian prairies and adjacent US, and extends eastward to the Lake Superior drainage basin on the Canadian Shield (Figure 1a). The bathymetry of the lake is flat and shallow ranging from about 11 m (South Basin) to 19 m (North Basin). Lake Winnipeg is located in the former basin of glacial Lake Agassiz (Figure 1a). Throughout its history, the Lake Winnipeg basin has been differentially uplifted by postglacial rebound in a northerly to northeasterly direction towards the centre of former maximum glacial loading in Hudson Bay. As a result, computations of theoretical former levels of Lake Winnipeg show a progressive increase in water surface area and southward transgression, under the assumption of open lakes in component basins overflowing their outlet sills to the north (Figure 1b).

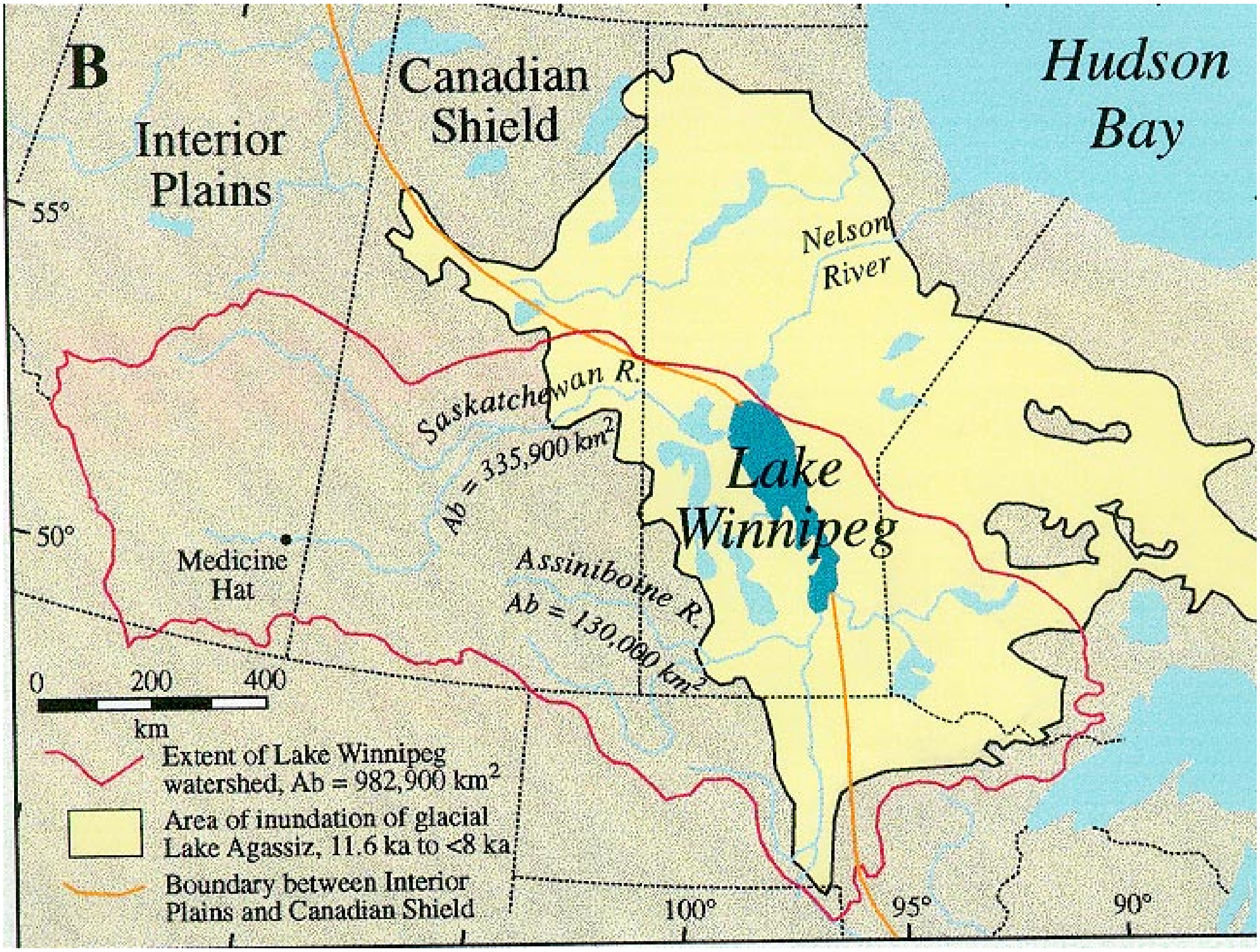


Figure 1a. Lake Winnipeg setting, showing the Lake Winnipeg watershed abutting that of Lake Superior at its eastern margin.

The level of Lake Winnipeg (217.4 m asl) is controlled by a hydroelectric power dam 79 km north of Warren Landing at Jenpeg, site of the former natural control. The lake presently overflows northward to Hudson Bay via Nelson River (Figure 1a). Bedrock lake floor now submerged 3-5.5 m at Warren Landing was an earlier control sill for North Basin. The Hecla-Black sill served as a control for an early separate impoundment in South Basin (Figure 1b).

The sedimentary sequences throughout the lake were studied with high-resolution seismic profiles and long cores (Figure 1c). Twenty-five cores fully penetrated Lake Winnipeg fining-upward muddy sediments into the underlying glacial Lake Agassiz silty clay, and 5 reached paleo-beach or nearshore sands under offshore mud, evidence of former low lake stands. Both seismic profiles and cores revealed a widespread inconformity between the Lake Winnipeg sediments and Lake Agassiz deposits, signifying a gap in time and a period of erosion. The surface of the clay-rich Agassiz sediments beneath the inconformity commonly displays a decimetre-scale zone of higher shear strength, higher bulk density, and a crumbly dry texture, indicating a period of soil formation and desiccation before deposition of the overlying Lake Winnipeg sediments. The desiccation zone is absent in the northern, deeper part of North Basin which is an area of continuous inundation since the basin was isolated before 7.7 ka (7700 radiocarbon years BP) by the recession and drainage of glacial Lake Agassiz.

The age-spans of subaerial exposure are severely underestimated by the computed open-lake model according to the ^{14}C ages of basal Lake Winnipeg sediment, especially in South Basin (compare Figs. 1c and 1b). The South

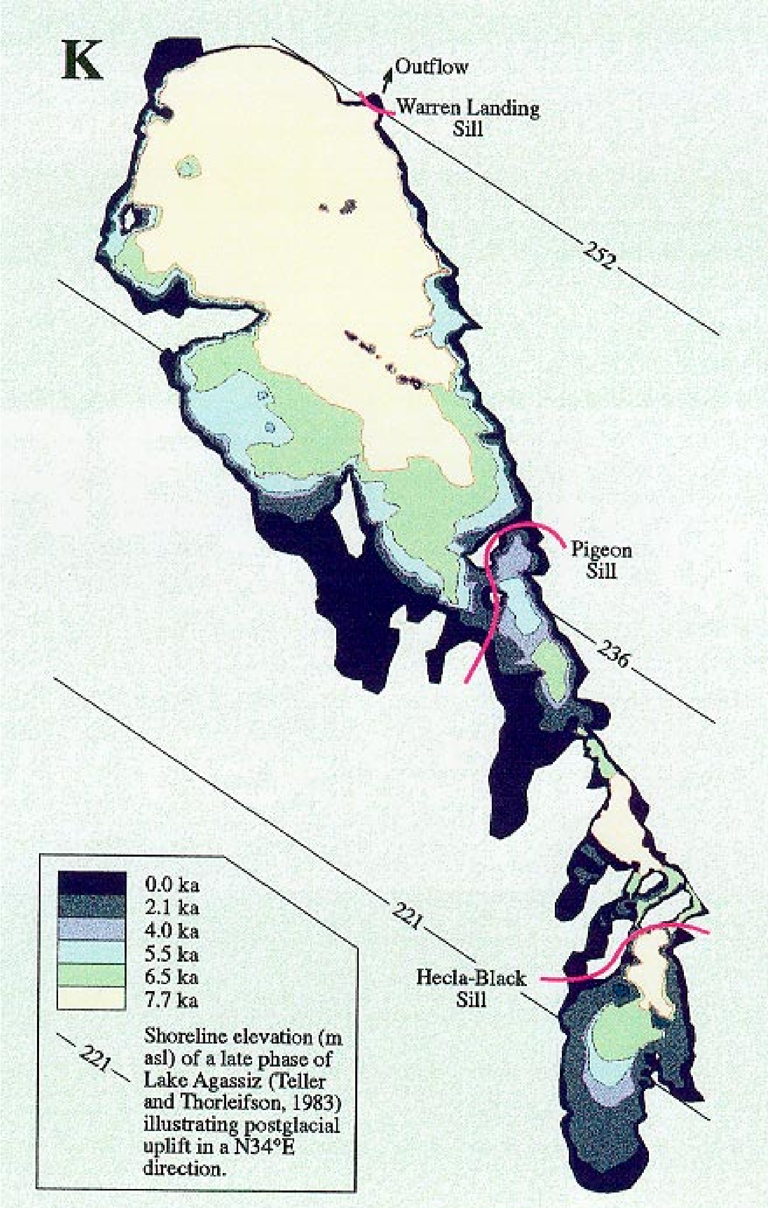


Figure 1b. Computed postglacial evolution of lake areas in the component basins of Lake Winnipeg assuming the lakes were open and overflowed their outlets to the north in differentially rebounding basins.

Basin was essentially dry for several thousand years, and then inundated by shallow brackish water, evidenced by fossil thecamoebian assemblages. The onset of Lake Winnipeg sedimentation did not progress in the expected southerly pattern, but began abruptly throughout the South Basin about 5000-4000 years ago, indicating inundation by a climatically-driven hydrological water surplus.

Inundation in the North Basin was progressive, shown by the southward younging sequence of basal Lake Winnipeg sediment ages, suggesting the climatic effect was less severe. The sediment sequence in the northernmost area is more complete, and changes in lake status are recorded, based on algal fossil remains and especially stable isotope ratios. At a core site in central North Basin, a long period of closed lake conditions (excessive evaporation relative to precipitation and inflow) from about 6.5 to 4.7 ka is indicated by consistently enriched ^{18}O ratios in cellulose compared with sediment pore water values. The former recorded primary biological productivity in evaporatively stressed surface waters, whereas the latter are entrapped bottom waters exhibiting average lakewater ^{18}O values.

The dry South Basin and closed North Basin are shown to be a consequence of the known warmer and drier mid-Holocene climate by evidence from vegetation history and analogue climate parameters for precipitation, evaporation and runoff. Ritchie's (1976) synthesis of paleo-pollen assemblages for the western Canadian prairies places southern and central Lake Winnipeg in the driest vegetation zone (grassland) at 6.5 ka (Figure 1a). The grassland zone is limited today to the southwestern Canadian prairies where q = 5 mm/yr (runoff per unit area), ep = 900 mm/yr (potential or lake evaporation), and p = 300 mm/yr (precipitation) (National Hydrological Atlas of Canada, 1978). In a simple space-for-time approach, variants of this modern grassland climate are applied in computing water balances of mid-Holocene Lake Winnipeg.

The maximum supportable lake surface area, A_m, in a drainage basin of total area A_b, is related to the mean annual climate parameters for the basin by $A_m = A_b * q/(q+ep-p)$ (Bengtsson and Malm, 1997) where q = specific runoff (runoff per unit area), ep = potential or lake evaporation, and p = precipitation. Lake basins are in closed status

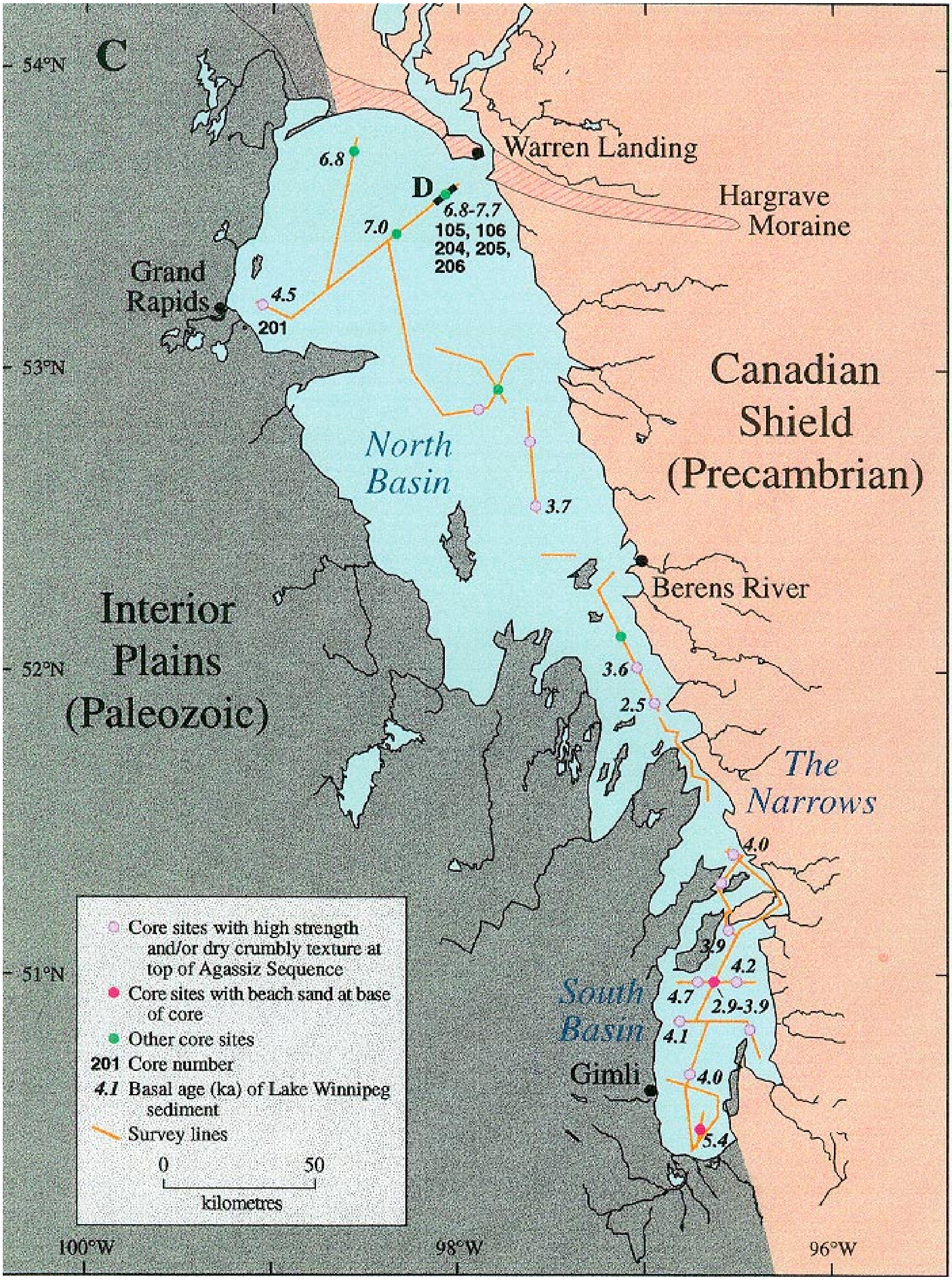


Figure 1c. Locations of seismic profiles and cores showing radiocarbon ages of basal Lake Winnipeg sediment.

when their computed open-lake areas exceed water surface areas supportable by their basin climate. Supportable lake areas for a moister version of the grassland climate accord with the geological observation of a closed North Basin (Figure 1d). The mid-Holocene climate apparently closed and dried the South Basin in accordance with the geological observations of widespread desiccation and delayed inundation (Figure 1e). These basins were separate hydrological entities until about 1.7 ka. The ^{18}O evidence shows North Basin switched to overflowing status about the same time the Saskatchewan River was diverted to it (4.7 ka) by rebound. Similarly, water balance for the South Basin was augmented by diversion of the Assiniboine River from Lake Manitoba about 4 ka. The generalized likely climate history experienced by Lake Winnipeg since 7.5 ka started with a dry grassland climate, less arid in the north, which created a closed North and a dry South Basin. This was followed by progressively moister climates (with short dry episodes) and a switch to open lake conditions starting about 4.7 ka in North Basin and 4.5-4 ka in South Basin.

Lake Erie

In 1985, Coakley and Lewis reported subsurface erosion, interpreted as wave erosion in former low-level shore zones as low as 39 m below present lake level (bpl) in the Long Point area of eastern Lake Erie (Figure 2a,b). These shore zones sloped upward in an easterly direction but projected below the sill of the lake outlet to Niagara River. Although it was considered possible that the shore features related to an earlier low-level phase which predated early Lake Erie, it was concluded more information was needed before their significance would become clear (Coakley and Lewis, 1985). A reevaluation of these observations with information and the example of climate-driven drawdown (closed-basin conditions) from Lake Winnipeg provides the needed clarity.

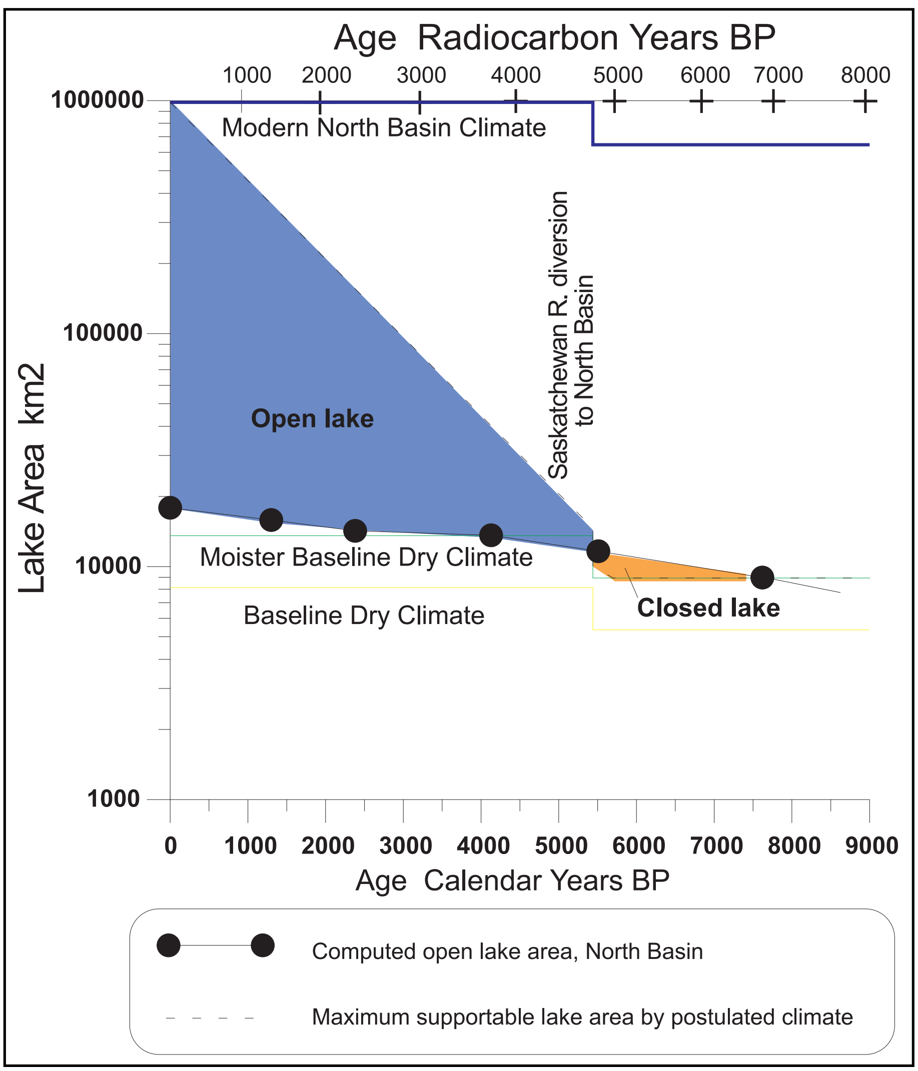


Figure 1d. Lake Winnipeg North Basin lake area vs age plot for both computed open-lake conditions and areas supportable by the basin climate.

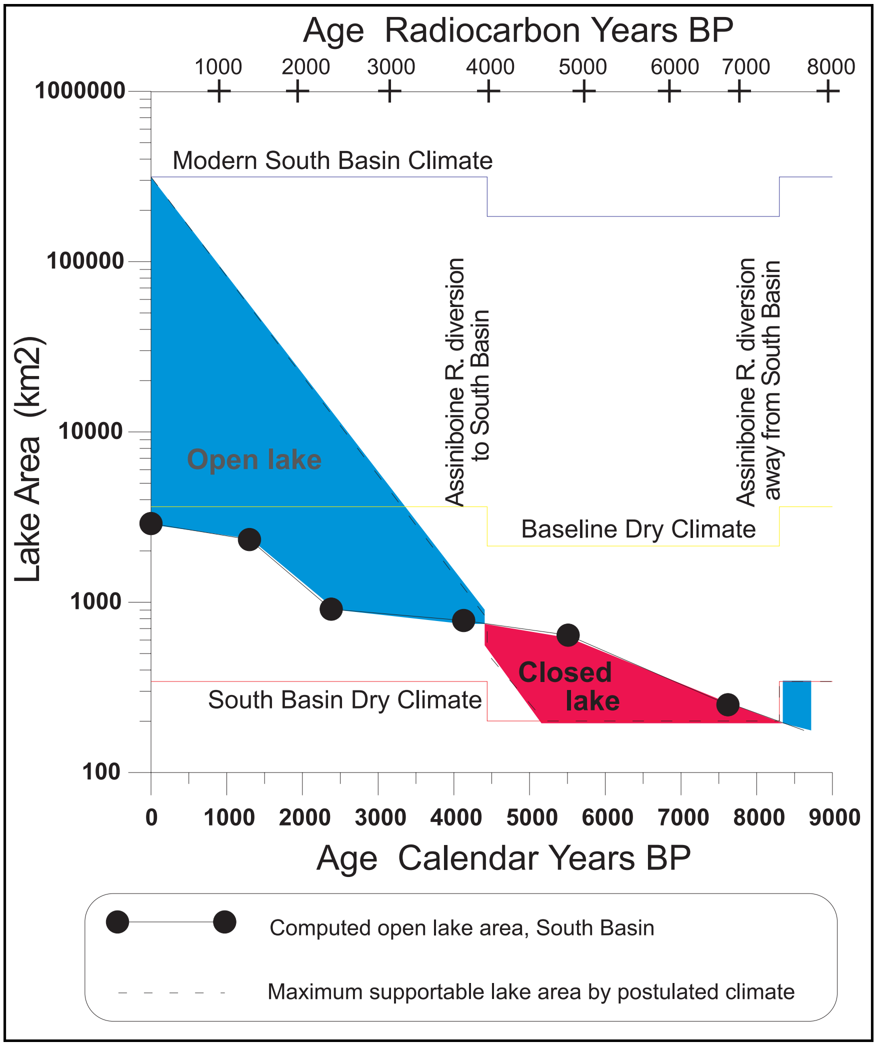


Figure 1e. Lake Winnipeg South Basin lake area vs age plot for both computed open-lake conditions and areas supportable by the basin climate.

The erosion surface is readily traced in seismic profiles and echograms beneath the mid- and late Holocene offshore muddy sediments in Lake Erie (Figure 2b). It is also apparent in a borehole near the western end of Long Point (Coakley and Lewis, 1985) and in piston cores of sediments on the north flank of eastern Lake Erie east of Long Point (Cameron, 1991). This inconformity is the eroded upper surface of glaciolacustrine sediments, and thus must postdate the termination of inflow of glacial sediment plumes from the upper Great Lakes about 10.5 ka when that drainage was diverted from Erie basin by glacier retreat and opening of northern outlets at lower elevation. The eastern Lake Erie inconformity recorded in the seismic profile (Figure 2b) extends down to about 27 m or more bpl, using an estimated sound velocity of 1500 m/s. The deepest erosion is interpreted as a paleo-shoreface with shorelines inferred approximately 4-7 m higher where a sandy terrace (beach) is found. Eastward, this shore zone projects to about 10 m± below the elevation of possible outlet sills to the Niagara River, indicating the existence of past closed-basin conditions.

Differential rebound progressively uptilted the Lake Erie basin in a northeastward direction throughout its early Holocene and much of its later history. This movement was a result of glacio-isostatic adjustment of the earth's crust to the former load of the Laurentide Ice Sheet centered over Hudson Bay. Hence, Lake Erie's outlet across the Niagara Peninsula to Lake Ontario was uplifted faster than other parts of its basin, and the capacity of the basin increased with time as the basin tilted upward to the northeast. Under climatic conditions of a water surplus, the lake would overflow its outlet, and the water level in this "open lake" would rise with the outlet elevation, causing the lakewater to backflood its basin and increase the lake area and volume. The rate of rise decreased with time, and the rising outlet elevation is generally expressed mathematically as a negative exponential function of shoreline age, as for the Lake Winnipeg basin. Using this function, the degree of basin uptilt for any desired age is

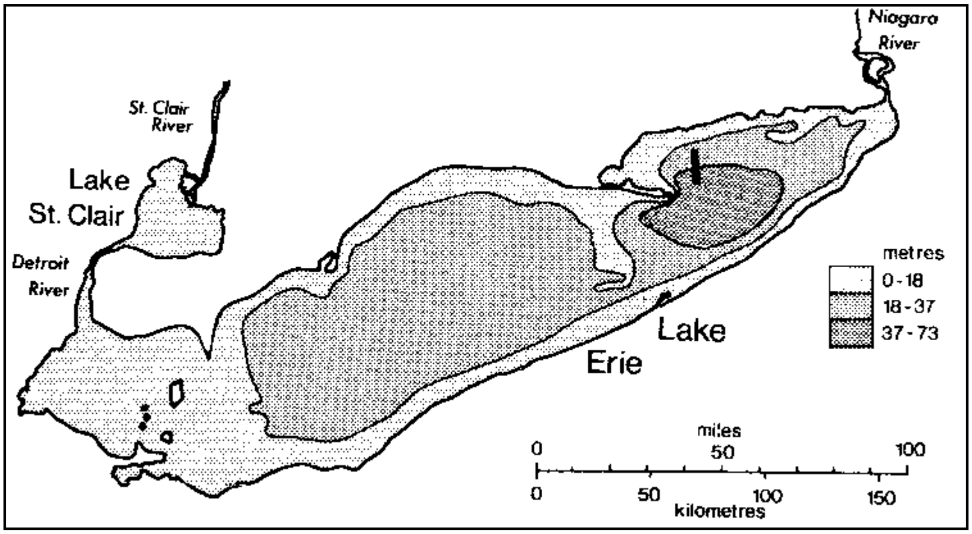


Figure 2a. Map of Lake Erie bathymetry showing the location of the seismic profile (thick line) in Figure 2b.

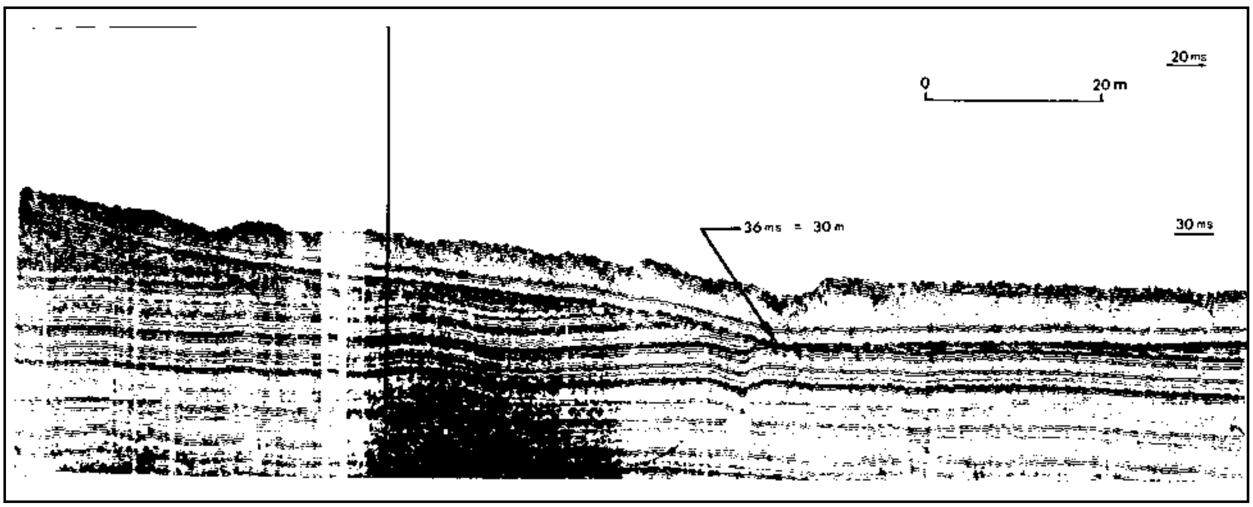


Figure 2b. High-resolution seismic profile showing a mud-buried erosion surface more than 36 ms (about 27 m for a sound velocity of 1500 m/s) below the present lake surface. From Coakley and Lewis, 1985; Figure 5, courtesy Geological Association of Canada.

readily calculated from the evidence of uptilted former glacial lake shorelines (Barnett, 1979; Coakley and Lewis, 1985; Calkin and Feenstra, 1985). The theoretical open-lake areas have been recently computed by Gareau et al. (1999) for the period 11.3 to 7.7 ka and are shown in Figure 3.

Even in the present-day climate, the water balance of the Erie basin is barely positive if the inflow from the upper Great Lakes were absent. Water balance calculations, based on the driest modern conditions between 1948 and 1995 indicate the basin was nearly closed in 1963 (q = 16.2 cm/a; e = 86.3 cm/a; p = 59.6 cm/a) (Figure 4a). If the driest values in the same period had occurred together in one year (1963 values for q and p, and e = 110.1 cm/a in 1991), the climate-supportable water surface area would have fallen to 25,200 km^2. This is 1,600 km^2 less than the current combined area of Lake Erie (25,700 km^2) and Lake St. Clair (1,100 km^2). The basin would have then experienced a water deficit.

Water supply to the basin was likely reduced in the early Holocene by enhanced evaporation owing to higher insolation, and different atmospheric circulation patterns while upper Great Lakes discharge was diverted away from the Erie basin as mentioned above. More evaporative lake conditions than at present are possibly also indicated by relatively heavy ^{18}O isotopic composition of offshore ostracodes and molluscs about 10-7.5 ka (Fritz

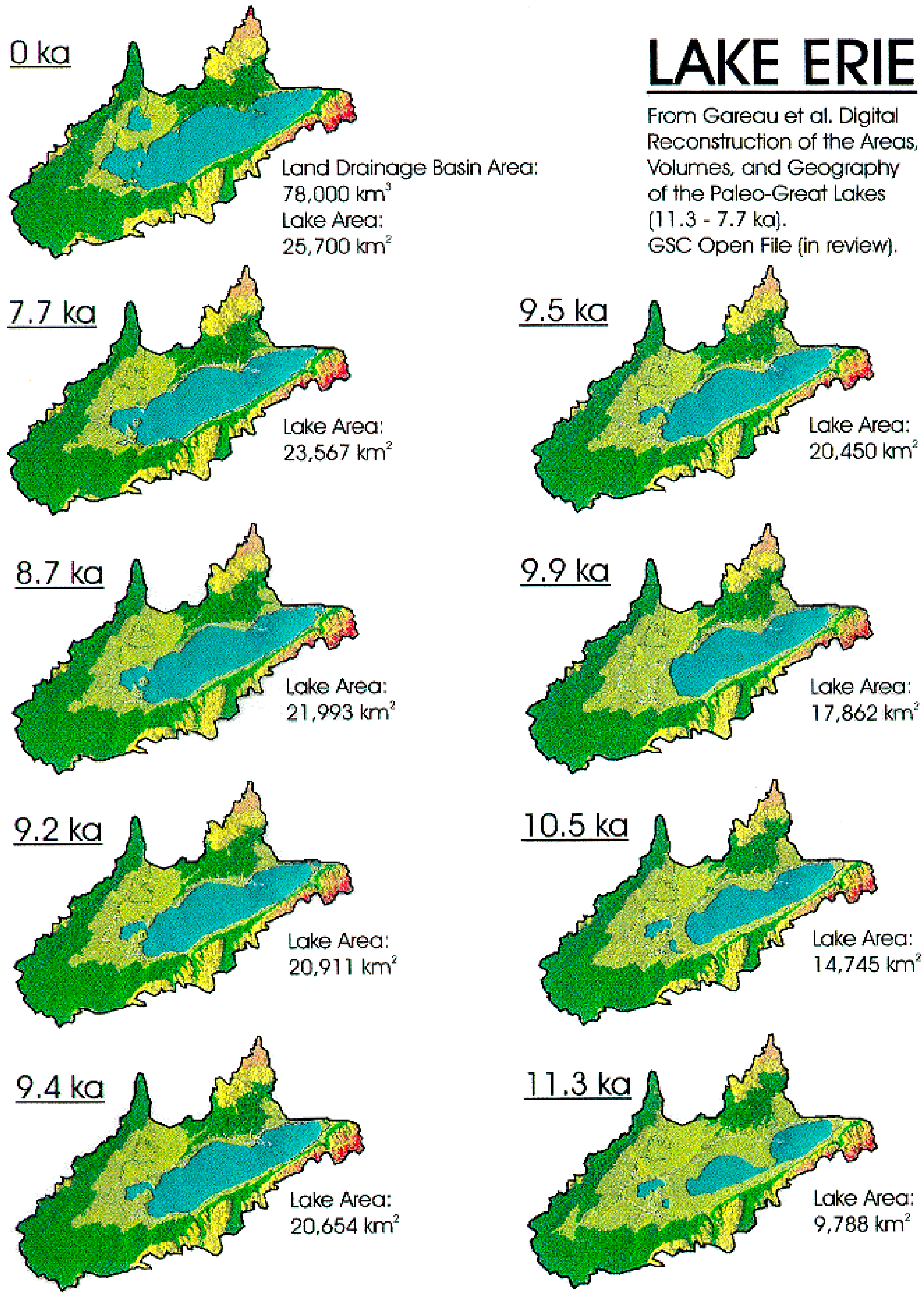


Figure 3. Shaded-relief images of the Lake Erie drainage basin showing computed open lakes (overflowing the outlet sill towards the north east) at eight stages of Lake Erie from 11.3 to 7.7 ka. Open lake areas are given for each stage. Drainage basin colors change at 100 m elevation intervals; dark green denotes areas between elevations of 200 and 300 m asl. From Gareau et al. (1999).

Figure 4a. Estimates of present-day supportable lake area in the Erie-St. Clair basin by its annual climate for 1948-1995, assuming there was no inflow from Lake Huron. The supportable lake surface area equals $A_b * q/(q+ep-p)$ where A_b = total area of drainage basin and lake, and *q, ep,* and *p* are mean annual specific runoff (runoff per unit area), lake evaporation, and precipitation, respectively (Bengtsson and Malm, 1997). The average parameters for the whole period (given in the figure) would support a lake area of 100,000 km^2, almost four times the present Lake Erie. However, in 1963, the driest year, a lake area of only 39,200 km^2 would have been supported, much closer to the 25,700 km^2 area of present Lake Erie.

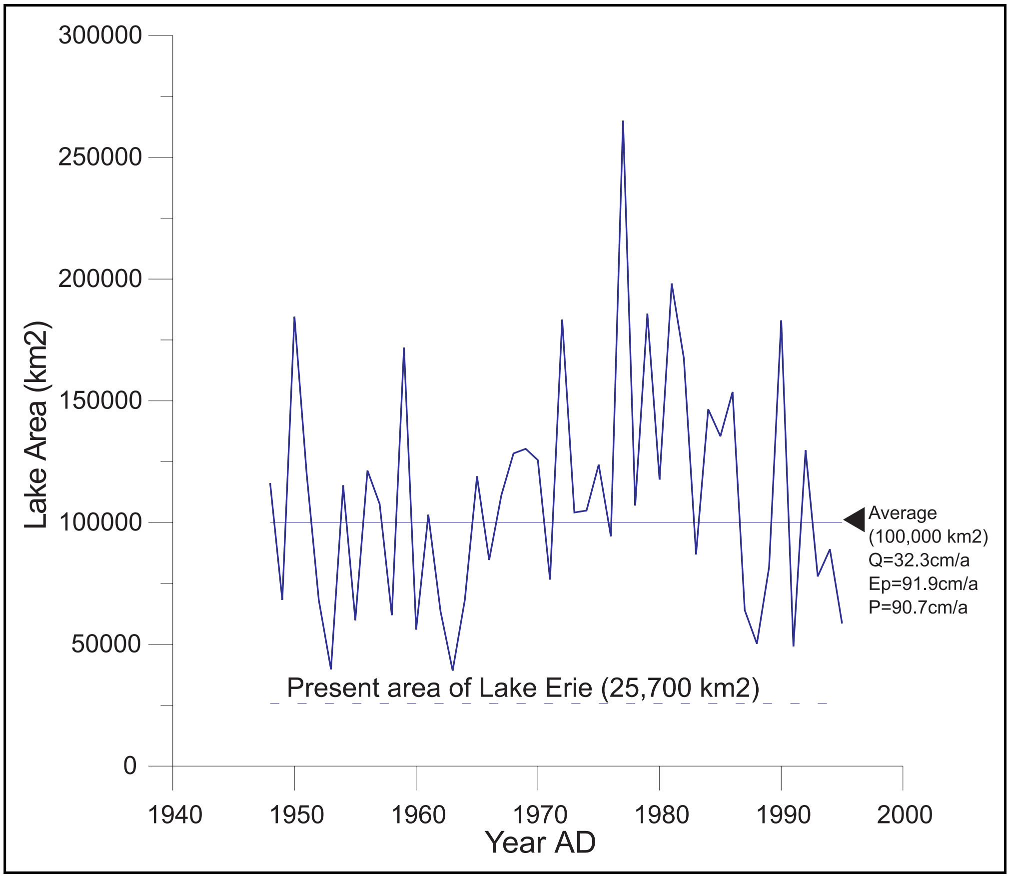
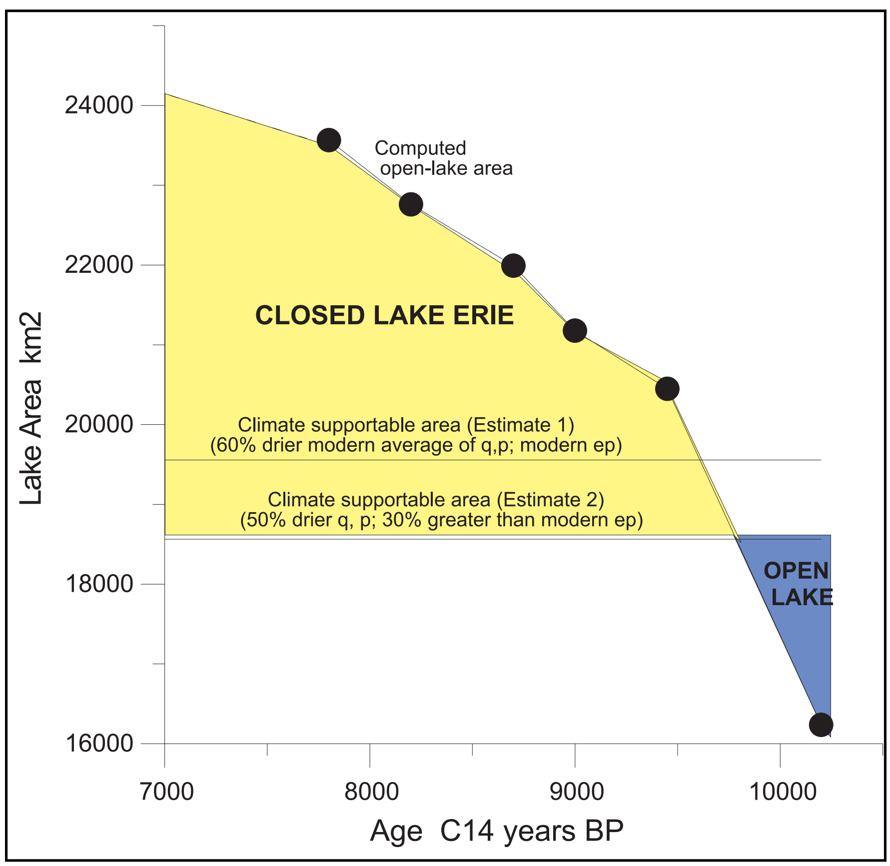


Figure 4b. Comparison of climate-supportable lake area and theoretical computed open-lake area in the Erie basin for the early Holocene. The basin was closed when climate-supportable lake area was less than that of the computed open-lake. The early Holocene climate is estimated to 60 percent drier than modern climate based on pollen transfer functions of soil moisture (Webb et al., 1993) and on stable isotopic proxy records of temperature and humidity (Edwards and McAndrews, 1989). Two estimates of early Holocene lake-supportable area, 19,600 and 18,600 km^2, are shown as thin horizontal lines. (The area of the much smaller Lake St. Clair, probably less than 1000 km^2, is ignored.) The open-lake areas were computed for water surfaces at the level of the basin outlet to the northeast (Fig. 3); these areas increased with time as the basin was uptilted to the northeast by ongoing differential glacio-isostatic rebound. The plots suggest that the climate-supportable lake areas became less than the computed open-lake areas between 10 and 9.5 ka. After this age, evaporation would tend to lower lake level below the outlet sill to reduce water surface area and evaporative loss; this would convert Lake Erie into a closed-basin lake. Closed conditions and reduced lake levels would continue until climate became moister or until inflow from the Upper Great Lakes began again during the Nipissing Great Lakes phase.



et al., 1975; Lewis and Anderson, 1992), as well as from the isotopic evidence in wood cellulose (Edwards and McAndrews, 1989). The erosion surface signifying the lake lowstand is coeval with the early Holocene pine pollen maximum in the Erie basin (F. McCarthy in Cameron, 1991; T. Anderson in Lewis and Anderson, 1992). This period is marked by a soil moisture reduction of 60 percent near eastern Lake Erie at 9 ka compared with modern values based on published analyses of regional pollen-inferred paleoclimate (Webb et al., 1993). Two climate estimates for the early Holocene are explored here, given the above paleoclimatic setting. Estimate 1 retains the same average evaporation as in the 1948-1995 period, but runoff and precipitation are reduced by 60 percent. Estimate 2 is 50 percent of modern runoff and precipitation with evaporation 30 percent greater than average values between 1948 and 1995.

Supportable lake areas for the early Holocene climate estimates 1 and 2 (19,600 km^2 and 18,600 km^2, respectively) are compared with computed open-lake areas in Figure 4b. After 10-9.5 ka, the open-lake areas are larger than can be supported by the early Holocene climate. Hence, it was quite likely the early Holocene water level was drawn down by net evaporative water loss into a closed state to form the lake lowstands observed in the sedimentary stratigraphic record. It is possible the lake existed in a closed state into the mid-Holocene until rebound diverted the Upper Great lakes outflow again through the Erie basin.

Lake Ontario

When water supply is sufficient to maintain lakes in an open overflowing state, climate fluctuations would still affect water level but to a much lesser degree than when lakes are closed (Bengtsson and Malm, 1997). This was the probable situation for Lake Ontario (and the other Great Lakes) during the late Holocene. Lake levels over the past 4000 years have been interpreted from the sedimentary record in three Lake Ontario lagoons—at Toronto (McCarthy and McAndrews, 1988), west of St. Catharines (Flint et al., 1988), and at Kingston (Dalrymple and Carey, 1990). All records show or imply fluctuations in lake level with amplitudes of about 1-2 m and durations of about 1000 years (Figure 5a,b). These results have been correlated by the last authors with higher than average levels occurring prior to 4000 BP, from 3100 to 2500 BP, and after 2000 BP. Dalrymple and Carey (1990) also suggest the Lake Ontario water level oscillations correlate with climate changes in northeastern North America, rather than with climate changes elsewhere in the Great Lakes basin as in the US midwest and in Lake Michigan (Larsen, 1985).

Recently, however, Yu et al. (1997) presented independent evidence of atmospheric control of effective moisture in the Ontario basin. Reductions in effective moisture and reduced water levels from 4800 to 2000 BP were detected in the lithostratigraphic and mollusc isotopic record within the sediments of Crawford Lake, a small isolated, atmosphere-controlled lake basin on the Niagara Escarpment, 25 km northwest of western Lake Ontario. They found a co-occurrence of lowered lake level (implying reduced moisture supply) and lighter ^{18}O ratios. These observations are most readily explained by a change in air mass circulation which brought drier air (with a lower ^{18}O signature) into the region more frequently than before. The alternate explanation of lowered levels in this lake, increased evaporation, is not tenable as heavier ^{18}O ratios would result, and these were not observed. In the US midwest, Baker et al. (1992) have suggested that variations in the relative presence of three air masses (warm dry Pacific air, warm moist tropical air from the Gulf of Mexico, and dry Arctic air) could account for shifts in the position of the prairie-forest border during the Holocene. Thus, it seems probable that if shifts in air mass distribution could change moisture-sensitive vegetation zones and small-lake water supply in the past, it could also change the hydrological water balances and water levels in the Great Lakes basins. These studies suggest that additional research of the paleo-record is warranted to better determine the mechanisms of the links between climate and lake level for application to achieving better understanding of potential future levels.

Summary

Lake Winnipeg. The paleogeography of Lake Winnipeg, inferred from glacial rebound, indicates the early Holocene basin was downtilted to the north such that separate impoundments existed in at least southern and north-

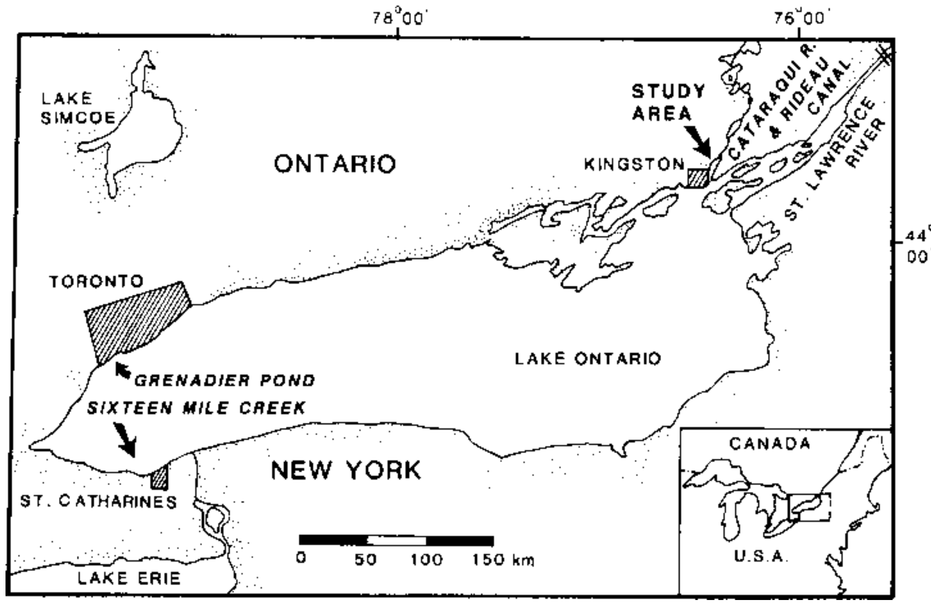


Figure 5a. Map of Lake Ontario showing three sites at which paleo-lake levels over the past 4000 years have been determined by geological study: Grenadier Pond in Toronto by McCarthy and McAndrews (1988); Sixteen Mile Creek near St. Catharines by Flint et al. (1988); and Cataraqui River in Kingston by Dalrymple and Carey (1990). From Dalrymple and Carey (1990), courtesy *Canadian Journal of Earth Sciences*.

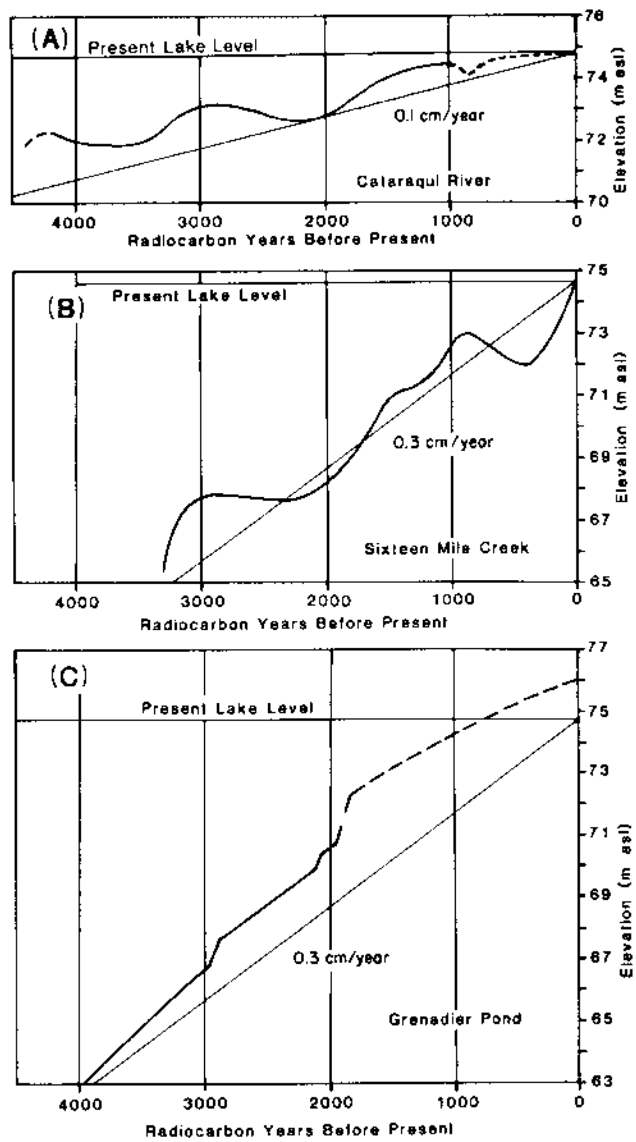


Figure 5b. Lake Ontario surface elevation versus radiocarbon age for the three study sites located in Figure 5a. The straight inclined lines are the computed long-term trends of lake level imposed at each site through tilting of the basin by differential glacio-isostatic rebound. From Dalrymple and Carey (1990), courtesy Canadian Journal of Earth Sciences.

ern sub-basins with outlets over their northern rims from about 7.7 to 2 ka. After 2 ka the northern basin level had transgressed and backflooded southern basins to form the present large lake. However, observations of sediment cores indicate that the long-term water level is not fully explained by a theoretical, open (overflowing) lake model based on rebound. Instead, the evidence shows that North Basin was closed from about 6.7-4.7 ka, and that the South Basin was dry from about 7.5 to 4 ka. The closed and dry conditions reflect a regional deficit in the water balance of the lake basin that is consistent with a drier-than-present paleoclimate in the mid-Holocene.

Lake Erie. Sediment stratigraphy in the eastern basin indicates an early Holocene lowstand, evidenced by an inconformity and mud-buried shore-zone morphology. This paleo-lake level which lies below the basin outlet indicates closed conditions and a regional deficit in the water balance of the lake basin as for mid-Holocene Lake Winnipeg. This condition is consistent with a dry pine paleoclimate and the lack of inflow from the upper Great Lakes at the time. Calculations based on estimates of early Holocene paleoclimate show that supportable lake areas were all less than computed open-lake areas, thereby confirming the basin was in a closed condition during the early Holocene, and possibly into the mid-Holocene.

Lake Ontario. A review of studies of shallow-water sediment stratigraphy in three lagoons around the lake indicate that lake-level oscillations of 1-2 m amplitude with durations about 1000 years occurred during the last 4000 years. New data from Crawford Lake near western Lake Ontario suggest that oscillations in Ontario levels could be linked to climate/hydrology changes, and that these changes could be driven by changes in atmospheric circulation and air mass humidity. Additional research is warranted to better understand the connections between paleoclimate and former lake levels as a step towards improved forecasting of future levels.

Acknowledgements

The Lake Winnipeg results would not be possible without the cooperative work of the many participants of the Lake Winnipeg Project led by the Geological Survey of Canada. The NOAA Great Lakes Environmental Research Laboratory kindly provided the modern hydrological parameters for comparison with evidence of closed conditions in early Holocene Lake Erie. Discussions with Thane Anderson (Canadian Museum of Nature) and Zicheng Yu (Natural Resources Canada) were helpful concerning early-mid Holocene paleoclimate and the atmospheric circulation model, respectively.

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~ Paleo Lake Levels — The Last Four-Thousand Years ~

RECONSTRUCTING HOLOCENE LAKE-LEVEL FLUCTUATIONS IN THE LAKE SUPERIOR BASIN

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Lake Superior is the largest lake of the five Great Lakes and overflows directly into Lake Huron. The mean water levels of the two lakes are at approximately 183.4 m and 176.3 m above sea level, respectively. Isostatic uplift is active across the entire Great Lakes basin and with rates of vertical movement increasing from southwest to northeast across the region (Coordinating Committee, 1977); thus, differential tilting continues to raise the bedrock-floored outlet channel from Lake Superior above the level of Lake Huron. The current rate of uplift at the outlet is about 0.33 m/century relative to the southern shore of Lake Michigan (Larsen, 1994). Between 5,000 and 2,100 years ago, Lake Superior was joined with Lake Huron and Lake Michigan and formed a single lake supplied by precipitation and evaporation over its drainage basin and governed by overflow through a single outlet channel at the south shore of Lake Huron (Hough, 1958; Farrand and Drexler, 1985). About 2,100 years ago, differential tilting began to raise the outlet channel from Lake Superior above the level of Lake Huron (Larsen, 1994). The present separation between lake surfaces was attained over the subsequent 21 centuries. Since separation from Lake Huron, Lake Superior has fluctuated due to its own hydroclimatic mass balance, with its mean level controlled (prior to construction of the Soo Locks) by the threshold altitude of the St. Marys rapids at about 182 m. The range of lake-level variation over the past century has been $\pm$ 0.5 m, reflecting variations in precipitation, evaporation and the outflow through the locks and St. Marys River at the eastern end of the lake.

Both isostatic tilting and climate change continue to significantly affect the relative levels of Lake Superior as they have over the past 5,000 years as evidenced by raised and drowned paleo-shorelines. Differential tilting of the Lake Superior basin has rearranged the attitude of the lake shores relative to the mean level of the lake at its outlet. Between 5,000 and 2,100 years ago, the mean level of Lake Superior was maintained by its previous outlet, the St Clair River, at the south shore of Lake Huron. Beaches formed simultaneously, when the combined three upper lakes were joined, are now uplifted relative to the level at that outlet. But, uplift of the St. Marys rapids above Lake Huron 2,100 years ago changed the pattern of relative shoreline positions and their land forms. Beach features deposited along the shores of Lake Superior since separation of the lakes continue to be uplifted along the north shore of the lake but relative to the new and present outlet (Larsen, 1987, 1994, 1999a). The northern shore of Lake Superior rises more rapidly than does the outlet due to tilting, while the southern shore experiences progressive submergence as the outlet rises more rapidly than this shore. Duluth harbor, for example, has undergone about 5.8 m of submergence over the past 2,100 years, evidenced by submerged tree stumps and the drowned channel of the St. Louis River. Correlated coastal landforms track the progressive submergence through time. Beach ridges deposited by contemporaneous lake levels are found near present mean lake level in the east, but descend in altitude to the west to where they are submerged near the Apostles Islands and at Duluth harbor. The earliest landforms from the previous three-lake system—the Nipissing phase shore deposits—now exposed at an altitude of 195 m near the outlet, are near or below lake level near Duluth (Larsen, 1999a, Larsen et al., 1999).

A new suite of radiocarbon dates from basal peat deposits lying on submerged beach deposits near the Apostles Islands defines the rate of uplift of the St. Marys rapids relative to Chequamegon Bay during the past 2,100 years. That rate is between 0.18 and 0.21 m/century, consistent with the observed historic rate of uplift of the outlet relative to this site. Other radiocarbon ages from stratified organics near the Apostles Islands outline broad fluctuating episodes of higher and lower climate-related lake levels superimposed on the rising mean level controlled by uplift at the outlet (Fig 1). High levels occurred at about 1,100, 400, and 200 radiocarbon years ago with low-level episodes at about 650 and 350 years ago. The implied range of lake level variation is greater than ± 1.0 m about the long term mean level of the lake. Dated sediments in vibracores from beach ridges on Long Island detail the most recent lake-level history. Lake level was high during the early 18th century and then fell to below the present mean level for periods between 1770 and 1825. The episodes of the 18th and early 19th centuries were also on the order of ± 1.0 m about the mean level. Since the mid 19th century the levels of Lake Superior appear to have fluctuated within the historic range of ± 0.5 m. The high-level episodes of the past tend to correspond with a high episode in evidence for eastern Lake Superior as well as southern Lake Michigan (Larsen, 1994). The highs preceding the historic period appear to reflect climate change related to the "Little Ice Age". A major low episode in lake level corresponds with a 13th century maximum of solar activity commonly associated with the "Medieval Warm Period" of northern Europe (Lean and Rind, 1998). The low-level interval of 350 years ago took place between periods of higher lake level. The low level interval tends to relate temporally with the "Maunder Minimum" sunspot interval recorded in Europe between 1614 and 1741 which was associated with cooler summer temperatures in Europe (Lean and Rind, 1998). The connection with historically-documented climate intervals in Europe is still unclear. We are cautioned that these events seem not to have been synchronous or in phase elsewhere.

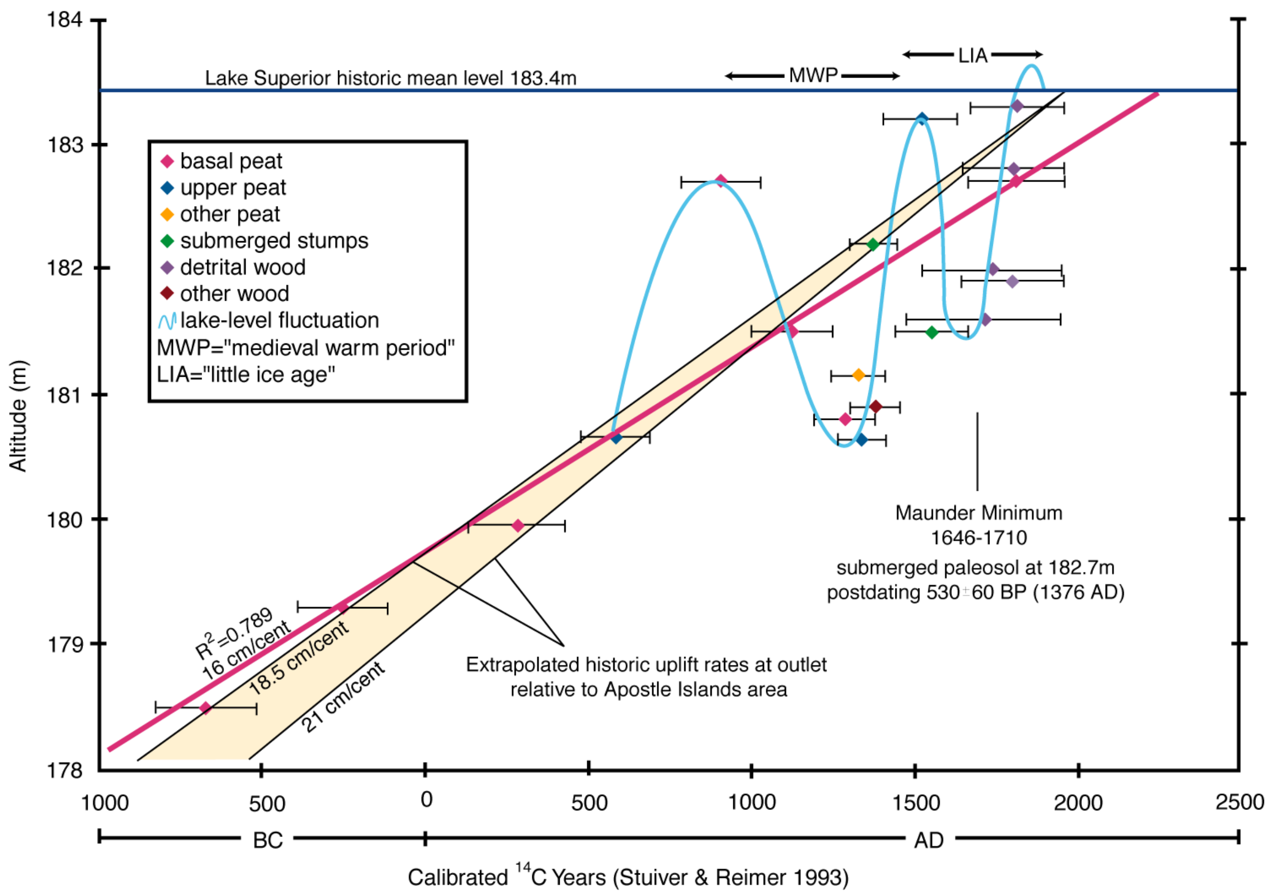

Figure 1. Paleo lake level record for the Apostles Island area, Lake Superior. Climate-related fluctuations in Lake Superior level are superimposed on a rising mean level caused by uplift of the St. Marys Rapids relative to the Apostles Islands. Red line shows a linear regression calculated on basal peat dates and suggests an uplift rate of at least 16 cm/century at the outlet. The historic rates of uplift of the outlet relative to the Apostles Islands (Coordinating Committee, 1977) are shown for comparison. Together these illustrate the rising mean level at this location. High and low level episodes on the order of 1-1.5 m above and below their contemporaneous mean levels are shown. Lows occurred during the "Medieval Warm Period" and within the "Little Ice Age." High levels of similar magnitude occurred during the "Little Ice Age" and preceeding the "Medieval Warm Period."

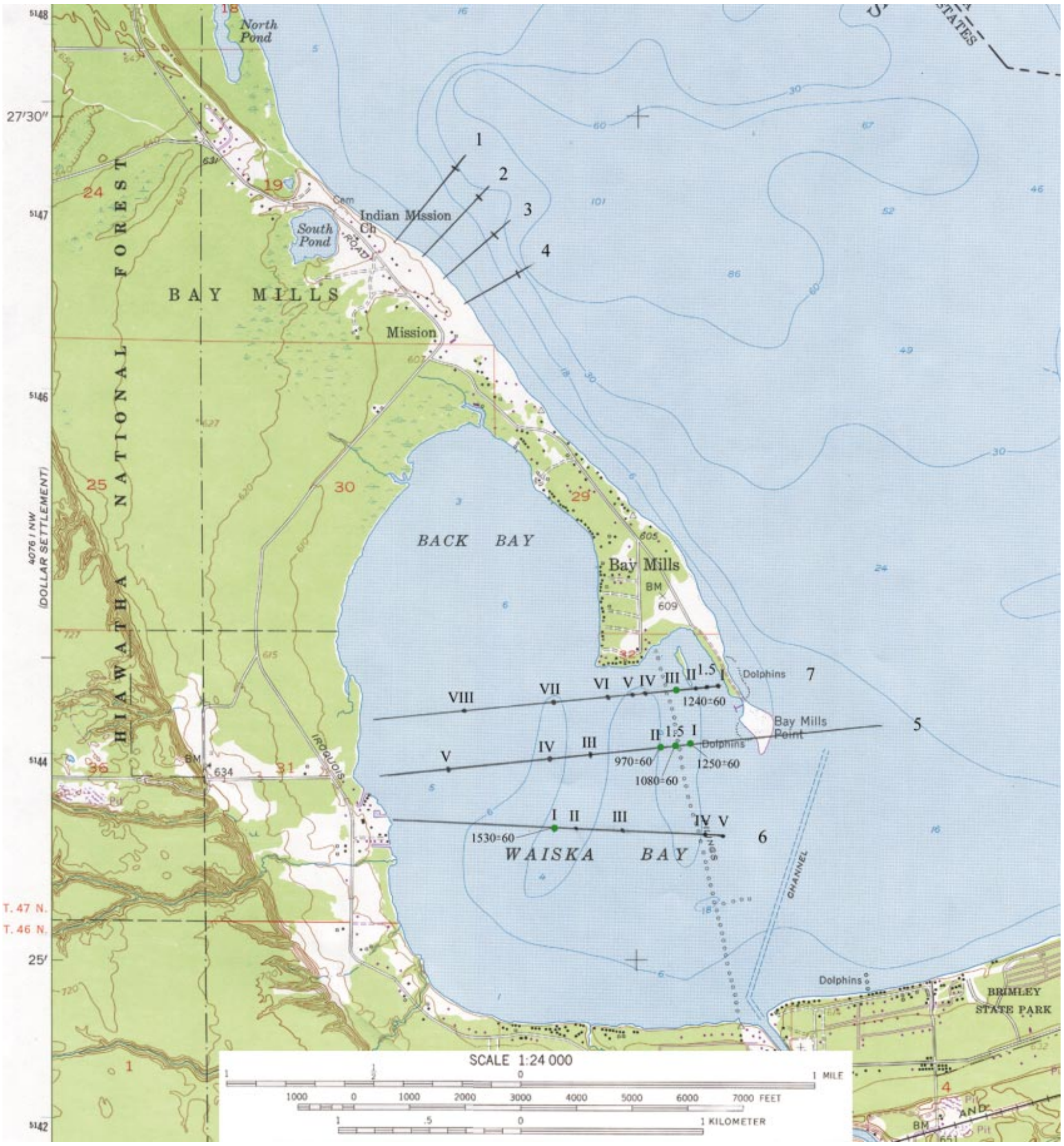

Figure 2. Submerged spit sequence at Bay Mills, MI. The map shows the positions of bathymetric profiles and core locations within the bay. The shaded area of the map outlines two spits, now submerged, that were deposited between 1500 and 1000 years ago. The sample locations and radiocarbon ages of detrital wood are also shown. A submerged offshore bar system associated with the lakeward edge of the middle spit along Line 6 shows it had an active shore face at a lake level 12.5 m below the present mean level between 1000 A.D. and the 18th and 19th centuries when the modern spit was destroyed.

Submerged land forms at the opposite end of the lake at Bay Mills, MI indicate anomalous low lake-level stands of Lake Superior (Larsen, 1999b). Here, near the outlet, the measurable effects of differential uplift are minimal. A spit complex at Bay Mills has prograded eastward across the mouth of Waiska Bay, built by longshore drift supplied by the erosion of bluffs to the west (Figure 2). The present spit (and the most recent in the sequence of spits) is clearly shown on early 19th century navigation charts prepared by the Royal Navy. Two earlier recurved spits were deposited across the bay mouth, but now lie submerged below lake level. Radiocarbon ages from detrital wood recovered from the bases of the submerged spits show limiting ages of 1530 ± 60 years BP (before present) for deposition of the earliest spit and 1240 ± 60, 1250 ± 60, 1080 ± 60, and 970 ± 60 years BP for deposition of the younger of the submerged spits. Bathymetric profiles across the bay show a preserved offshore bar system lying in 3 m of water on the lakeward (north) flank of the younger submerged spit. A comparison with an analogous offshore bar system along the active shoreface of the modern spit at Bay Mills shows that the preserved bar system was deposited at a water depth 1.5 m below the present mean level of Lake Superior (183.4 m). This same depth corresponds with the bedrock spillway of the St. Marys rapids that formerly controlled the level of the lake. The limiting ages of the submerged spits and the earliest map of the present active spit show that the preserved offshore bar system was deposited between 970 + 60 years BP and the early 19th century. Spit-building activity is thought to correspond with episodes of higher lake level when bluff erosion supplied greater sediment loads to the littoral system. The drowned offshore bar system relates to a low-level episode that occurred between 1,000 AD and 1824 AD that allowed reworking of the active face of the spit. Subsequent high levels and concomitant bluff erosion contributed sediment to the present spit which then served as a protective barrier to the previous spits.

The indicated low level 1.5 m below the modern mean level of Lake Superior suggests that the lake fell to the threshold depth of the spillway to the St. Marys River and that little or no outflow from Lake Superior took place at that time. Given the range and timing of the low lake-level episodes recorded near the Apostles Islands, either the Medieval Warm Period low or the low stand mid way through the Little Ice Age could account for the anomalous submerged bar system at Bay Mills.

These reconstructions of past lake levels in Lake Superior are derived from independent data sets. As the uppermost lake in the Great Lakes system, Lake Superior is the lake most directly controlled by changes in precipitation and evaporation in the region. Assuming constant outflow, high and low episodes of the past most certainly reflect climate change on a century and sub century range.

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~ Paleo Lake Levels — The Last Four-Thousand Years ~

LAKE LEVELS IN THE ERIE BASIN: DRIVING FACTORS AND RECENT TRENDS

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Introduction

Three major factors control lake levels in the Erie basin. They are: postglacial isostatic rebound of the crust at its outlet, regional hydrological and climatic changes, and to a less dramatic extent, long-term neotectonic movements. Under the influence of the above factors, Lake Erie has undergone a complex history of rises, declines, and stability since the close of the glacial period some 12 000 years ago. From a low-level stage more than 50 m below present levels, the lake has risen to its present stage. The heavy solid line in Figure 1 represents a reconstruction of postglacial lake levels in the Lake Erie basin. It is based on over 50 radiocarbon-dated strandline indicators collected by the author and others at sites around the lake basin (see Table 1 in Coakley and Lewis, 1985). The upper limits of the original envelope of probable levels proposed by Lewis (1969) are shown as a light dotted line. A useful discussion of lake levels in the Erie basin over the past 12 000 years is also presented in Pengelly et al. (1997).

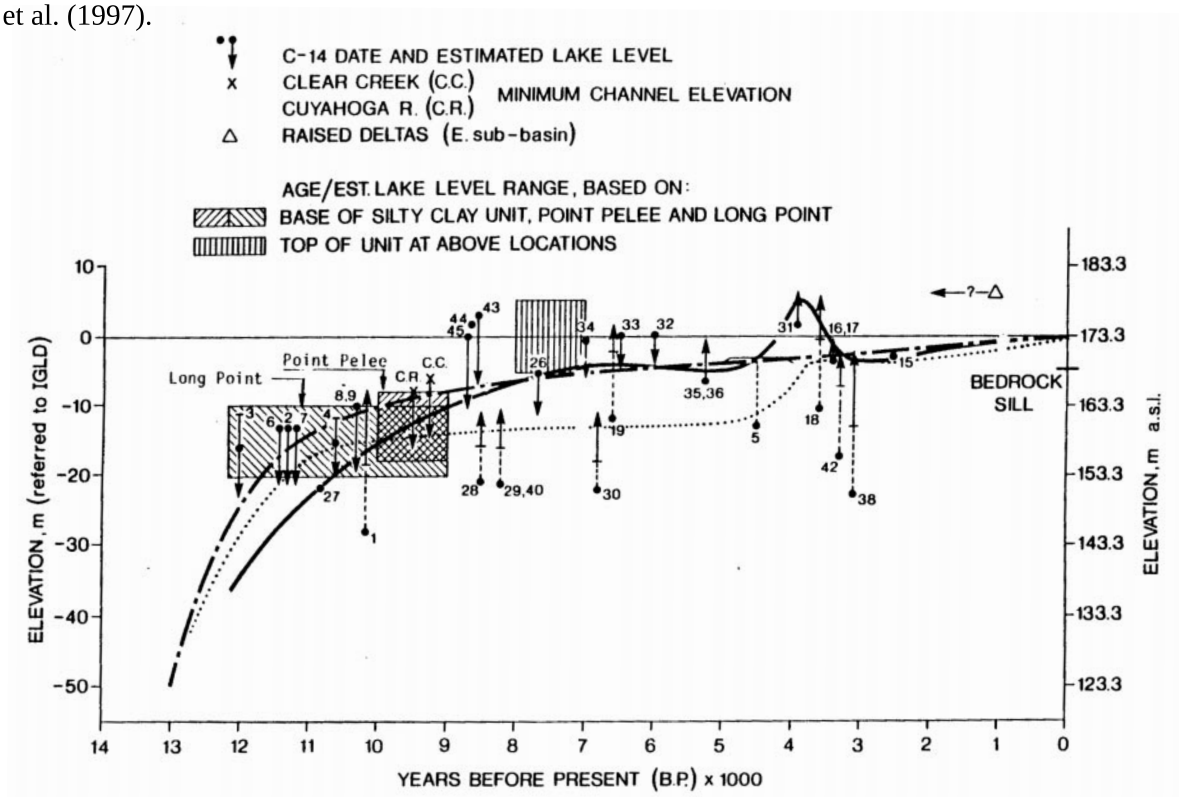

Figure 1. Reconstructed history of postglacial Lake Erie levels (heavy solid line) based on over 50 radiocarbon-dated level indicators reproduced from Coakley and Lewis (1985). The heavy dashed line is an hypothesized level curve assuming a composite model of uplift and overflowing conditions at the outlet; the light dotted line is the upper fit envelope in Lewis (1969). The curve is applicable to the central and eastern basins only. The radiocarbon dated references used in the curve are listed in Table 1 in Coakley and Lewis (1985).

Postglacial isostatic rebound. Lake Erie rose from its minimum levels (Early Lake Erie) due to crustal rebound of the outlet at the Niagara River from its glacier-depressed state. The recovery rate decreased exponentially with time. Reconstruction models based on exponential rebound alone (Andrews, 1970) grossly overestimate the initial rise in levels, compared to that based on dated strandline indicators. However, the curve based on a composite model of exponential uplift, originally proposed by Mörner (1980) and elaborated on in Coakley and Lewis (1985) (shown in Figure 1 as a heavy dashed line) provided a much more reasonable fit to the data points. Both curves represent lake levels in the central and eastern basins assuming confluence of these basins and overflowing conditions at the outlet. Two important discrepancies are evident from the two curves:

- Most of the indicators in the western basin lie above the reconstructed level curve indicating that this part of the lake was subaerially exposed or contained perched bodies of water.

- The difference in elevation between the composite model curve and that based on dated strandline indicators suggests that the rapid initial rebound, together with negligible upper lakes inflows and a relatively dry climate at the time, resulted in a period of closed conditions in Lake Erie. In other words, drainage within the Lake Erie watershed had no outlet except evaporation. Although such a scenario remains to be confirmed in the sediment record, it has been cited by Tinkler et al. (1994) as an explanation for anomalous radiocarbon age sequences in shells from the Niagara Whirlpool site.

The less-than-ideal fit of the lake level curve to the data set prompted the idea that postglacial uplift might not have been as simple as predicted on the basis of simple or composite rebound model utilizing a fixed hinge-line (Hough, 1966). Geophysical complexities such as lower mantle rheology and the presence of a "peripheral forebulge" in front of the ice margin have been cited as perturbing factors in the rebound dynamics of a glacier-depressed crust (Pelletier, 1986). Also, preexisting crustal movement might play a role (see section below). Nevertheless, one assumes that with time, the effect of crustal movements related to rebound would decline, and that the other factors mentioned at the beginning of this paper would gradually assume a larger importance. These are discussed briefly below.

Hydrological and climatic factors. A number of significant hydrological events, involving major changes in inflow, outflow, and evaporation-precipitation, have occurred in postglacial Lake Erie. Evidence for changes in lake levels caused by such effects is derived largely from postglacial sediment records, such as those from below the three major cuspate forelands along the north shore (Coakley, 1976; 1985). Lewis and Anderson (1989) found evidence in sediment cores from a deep borehole on Long Point that waters from glacial Lake Algonquin discharged briefly into the Erie basin around 11,000 years Before Present (BP). Lewis and Anderson (1989) and Tinkler and Pengelly (1995) also concluded that catastrophic releases from Glacial Lake Agassiz, a huge glacial lake covering most of the prairie provinces of Canada, reached Lake Erie. The hypothesized result was significant, but transient, peaks in Erie levels around 8,000 years BP. The possibility of such a scenario was questioned by others (Rea et al., 1995), but the matter is still a subject of debate. The best documented episode of a singular excursion from the predicted exponential model is that now referred to as the Nipissing event or "flood". This event was initially proposed by Hough (1966) and expanded upon by Lewis (1969), Coakley and Lewis (1985), and others. Climatic change was also postulated by Lewis and Anderson (1989) and by Lewis (1999, this volume) as a reason for water level changes in the Erie and Huron basins. The evidence for climate-induced changes in lake levels is most often found in dated profiles of stable isotope (carbon and oxygen) concentrations in lake sediments (Fritz et al., 1975; Tevesz et al., 1997) and in pollen profiles (Lewis and Anderson, 1992). However, as changes in isotopic composition can be related to both climate and hydrological events, it is difficult to separate their individual effects.

Sedimentary evidence for the above dramatic lake level events is sparse in the Lake Erie basin, especially when compared to the excellent records preserved in the Lake Michigan beach ridges (Thompson and Baedke, 1999, this volume). Because of the location of the lake outlet at the northeast end of the basin where uplift was greatest, the lake gradually backfilled from east to west. This resulted in the sequentially drowning or erosion of all low-level shorelines or in their burial below tens of meters of postglacial sediments. High level stages, i.e. above the present lake level, are also difficult to trace in the present landscape because intense shoreline recession during the Holocene over most of the basin has obliterated such shorelines. Furthermore, any such high-level stands might have been too brief to leave durable shoreline indicators. In areas where any might have survived, such as in the bedrock-dominated shores in the eastern parts of the basin, they are ill-defined and controversial. For that reason, the most useful evidence of lake evolutionary changes might be found in the sediment record preserved below the large cuspate forelands of the Erie basin (Coakley, 1985) and in other isolated occurrences (Barnet et al., 1985; Barnett, 1985). Furthermore, because of the erosive environment of the lake, material capable of providing good radiocarbon dates (e.g. trees in growth position) are hard to find, and there is a problem with transported organics and other error-producing effects. These deficiencies are discussed in detail in Coakley and Lewis (1985).

Role of background neotectonic tilting of the Lake Erie basin. Finally, it is becoming clearer that tectonic factors unrelated to glacial rebound, referred to here as neotectonic factors, play an increasing role in recent lake level trends. According to most exponential rebound models (excluding those that are based on complex factors such as a "peripheral forebulge" responding in a manner opposite to that of the area of rebounding crust), substantial rebound of the Niagara outlet at Buffalo should have been reduced by now to virtually zero. Still, recent trends deduced from water-level gauge monitoring at various parts of the lake, and from regional geodetic surveys show that there is an ongoing tilting of the Erie basin. Because of the known neotectonic history of this region (Sanford, 1993; Mohajer, 1993) there is no reason to doubt that these trends represent ongoing linear crustal adjustments near the outlet that might predate postglacial rebound. The bold dashed curve in Figure 1 is based on a combination of such linear uplift with the standard exponential model of crustal rebound.

Lake Erie Levels: 4000 Years BP to Present

The Nipissing event. When levels in the Huron basin had reached the level of the Port Huron sill, drainage was shifted from the North Bay (Mattawa-Ottawa River) outlet to southern outlets at Chicago and Port Huron-Sarnia. The timing for this event, initiating an inflow of approximately 6000 $m^3.s^{-1}$ to the Erie basin, is estimated at between 5500 years and 3700 years BP (Lewis, 1969). Archaeological surveys at native campsites at Fort Erie, Ontario (at the entrance of the Niagara River) found no artifacts dating older than 4000 years, indicating that the site was inhospitable to human habitation (too wet or flooded) before then. Such a massive inflow is expected to have been enough to overwhelm the then-existing outlet channel in the Niagara River and to cause a sharp rise in Lake Erie levels. The duration of this event is open to debate, but it probably ended after a time-period of decades to centuries when the outlet channel was sufficiently eroded and increased in area to accommodate the increased inflow. Alternatively, lake level control might have shifted temporarily to another lower sill. In any event, levels then fell to near their former levels.

The maximum elevation of the Nipissing "flood" in the Erie basin remains uncertain. Barnett (1985) was the first to locate and describe small tributary deltas in the eastern end of the lake near Port Rowan at an elevation of 3-5 m above the present lake level. These delta features yielded radiocarbon dates of around 1000 BP, i.e. too young for the Nipissing event, but the dates, on fine-grained organics, might be contaminated by young material. In any event, this find supports the figure of 3 m above present levels proposed by Coakley and Lewis (1985) as the maximum for the Nipissing rise. Evidence of this event elsewhere in the basin is hard to confirm, but support may be interpreted in the drowned river sections at Clear Creek, Ontario (Barnett et al., 1985), Old Woman Creek, Ohio, and, indirectly, in the eroded platforms identified beneath Point Pelee and Pointe-aux-Pins, two of the three large cuspate forelands of the north shore (Coakley, 1976; Coakley et al., 1997). During 1999, the author plans to add to the data base by drilling a series of boreholes in the eastern end of the lake near the Wainfleet Bog. It is hoped that these new data would provide badly needed confirmation of Nipissing-age events.

Post-Nipissing trends. From the above levels, the lake resumed its rise under the influence of the tectonic uplift component described above. Estimates of the rate of this rise are based on the assumption that the relative depth changes noted in the gauges since the mid 1800's are caused by a rise at the Buffalo end of the lake and not a subsidence in the western end. The latter would not lead to a true rise in lake levels elsewhere in the lake. The most comprehensive treatment of the gauge data is presented in Clark and Persoage (1970), in recent reports of the Coordinating Committee on Lake Levels (1977), and in Tushingham (1992). In the latter, the rate of change in gauge readings between Buffalo and points in the western end of the lake (deepening with respect to the Buffalo gauge) varies between 9 $cm.century^{-1}$ (Buffalo-Cleveland and Buffalo-Kingsville) and 7 $cm.century^{-1}$ (Buffalo-Marblehead). However, rates for other comparable pairs of gauges, for example, 4.5 $cm.century^{-1}$ for Buffalo-Toledo, are anomalously low and thus raise questions as their accuracy. Nevertheless, a figure of 8 cm. $century^{-1}$ was used in Coakley and Lewis (1985) as the underlying linear component in a composite model of lake level rise (Figure 1).

Summary

The causes of postglacial lake level rise in the Erie basin are varied and thus, predicting future trends is difficult without more research and data. However, using past trends as a guide for the future, levels will continue to rise at rates of approximately 5-10 $cm.century^{-1}$, driven primarily by neotectonic processes. Such a prediction is valid only if all other factors, such as regional hydrology and climate remain unchanged. This is an important qualification, however, as there are increasing fears of climate changes taking place on a global scale. Another imponderable in the prediction of future lake levels is the effect of increased pressure on Lake Erie water resources for consumptive water uses, irrigation, and the drive for greater efficiency in navigation and shipping. All these activities have the potential for lowering lake levels, especially when coupled by trends to a warmer and drier regional climate. Of all possible scenarios, dramatic lowering of Lake Erie would have the most serious impact. The negative impact on coastal wetlands, shipping tonnages, and the increased potential for resuspension of shallow-water contaminated sediment deposits by wave action and increased navigational dredging, could be devastating. The flip side of the coin is the economic impact of shore erosion and shoreline flooding if the lake rises above present levels.

As is evident from the list of references below, there is a scarcity of recent research into long-term lake level trends. Hopeful developments are the borehole drilling at Wainfleet by the National Water Research Institute, and plans for further archaeological investigations in the eastern part of the lake. These efforts, and others yet to be announced, will no doubt add considerably to our understanding of late Holocene lake levels in the Erie basin.

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~ Paleo Lake Levels — The Last Four-Thousand Years ~

STRANDPLAIN EVIDENCE FOR RECONSTRUCTING LATE HOLOCENE LAKE LEVEL IN THE LAKE MICHIGAN BASIN

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Establishing Past Lake Level

The reconstruction of late Holocene lake-level changes in the Lake Michigan basin requires information to be collected that indicates the elevation of the lake and the particular time that the water was at that elevation. One geologic approach is to collect paleo-elevation and paleo-age information from coastal and paralic (environments bordering the coast) features and sediments along the lake's margins (Table 1). Two methods are commonly employed to establish the lake's past elevation using coastal and paralic deposits and features. The first is geomorphic; it approximates the paleo-elevation of the lake by measuring the elevation of a geomorphic feature. Such features include wave and stream-cut terraces, delta platforms, stream and valley fills, spits, dunes, barrier beaches, and beach ridges. Examples of this type of approach are topographic surveys across a series of beach ridges (e.g., Johnson et al., 1990; Lichter, 1995; Petty et al., 1996), where elevation changes between the crests of the ridges are assumed to represent the magnitude of lake level change through time. A correction is then needed from the elevation of the dune crest of each ridge to where the elevation of actual lake level was when the ridge formed. The geomorphic approach is the most commonly used technique for estimating the elevation of past lake levels throughout the entire Great Lakes, its legacy extends back into the late 1800s (e.g., Leverett, 1897; Goldthwait, 1908).

Table 1. Types of elevation and age data that can be used to reconstruct past lake levels in the Lake Michigan basin.

Elevation Data	Age Data
Geomorphic - Terraces and wave-washed platforms - Fluvial - Coastal - Beach ridges, spits, barrier beaches, etc. - Delta platforms **Sedimentologic/Stratigraphic** - Facies - Facies contacts - Stratigraphic sequences	**Relative** - Geomorphic relationships - Stratigraphic sequences **Radiocarbon** - Wood, peat, bones, shells, etc. **Other Radiometric** **Incremental**

The second method of determining lake level elevation is sedimentologic and stratigraphic, where the elevation of the lake is determined by measuring the elevation of sedimentary deposits in coastal and paralic features and sediments. In this method, cores or outcrops are used to determine the elevation of the sedimentary deposits or the elevation of contacts between sedimentary deposits. Like the geomorphic method, a correction may be needed to yield the actual elevation of the lake from the elevation of the sedimentary deposits. Adding approximately 4 meters to the elevation of the contact between upper shoreface and lower shoreface deposits in a progradational or depositional regression sequence is an example of using this technique. Although imprecise, the base of the upper shoreface (fair-weather wave-base) is about 4 meters below lake level (Fraser and Hester, 1977).

Often sedimentary deposits do not have a clear-cut relationship to lake level. Sequences of sedimentary deposits, however, may be used to determine whether lake level was rising or falling while the deposits were accumulating. An example of this stratigraphic approach is determining lake level behavior from a sequence of palustrine (swamp) deposits landward of a beach ridge, barrier beach, or spit (e.g., Fraser et al., 1990). If the sequence indicates that the swale has become flooded through time and if a connection between the groundwater table and the lake can be established, then one could interpret that lake level was rising and causing the groundwater table to rise in the wetland. Regardless of which sedimentologic approach is used, both require a thorough understanding of facies characteristics within the myriad of coastal and paralic depositional environments that surround the lakes. Just as important, however, is a thorough understanding of how the entire depositional system behaves in response to lake level fluctuations.

Prior to the introduction of radiometric dating, the age of geomorphic features and sedimentary deposits was estimated by using their relative position to each other and the lake proper. For geomorphic features, these include elevation and distance landward from the modern shoreline. Elevation and distance landward can be also used for sedimentary deposits, but stratigraphic position and cross-cutting relationships are more useful. Relative positioning is normally the first technique used in interpreting age relationships, producing a relative time scale. Radiometric dating, commonly C^{14}, has transformed the relative time scale into an "absolute" one. C^{14} dating of organic deposits, however, is often opportunistic in that organic samples are dated when they are found. The age relationship between opportunistic organic deposits and a geomorphic feature or sedimentary deposit that can estimate lake level should be well known before it is dated.

Collecting Lake-Level Data in Lake Michigan

Beach ridges are shore-parallel ridges of sand that contain a core of water-laid sediment. They are formed in the final stages of a water-level rise in areas where there is a net positive supply of sediment to the nearshore system (Thompson and Baedke, 1995). They commonly occur in embayments as a series of ridges and wetland-filled swales — a strandplain. Strandplains may contain the most complete records of past lakelevel fluctuation of coastal and paralic features and deposits that surround Lake Michigan. Additionally, the beach ridges contain foreshore (swash zone) deposits that accumulate at or very near lake level. Of particular interest is the gravely plunge point at the base of the foreshore deposits (Figure 1). These sediments are the closest approximations of actual lake level that can be obtained within the nearshore system. By collecting these sediments within a strandplain, the upper limit of lake level during the development of the strandplain can be determined. Another advantage of studying strandplains is that the wetlands in the swales of beach ridges can be used to date the age of the ridges. That is, the age of the basal organic sediments in each wetland is a close approximation of the age of the ridge lakeward of each wetland. This relationship, however, is probably only valid where the strandplain has an elevated groundwater table, possibly from ground water focusing into the strandplain from surrounding uplands.

Five strandplains, differing in dimension, orientation, sediment supply, wave climate, and geographic location, were studied in the Lake Michigan basin (Figure 2) (Thompson and Baedke, 1997). Foreshore deposits were collected for most ridges by vibracoring the lakeward margin of each ridge. Foreshore deposits were identified by their grain-size characteristics and sedimentary structures. Their elevations were determined by leveling the core sites to the nearest 0.003 m, using the International Great Lakes Datum of 1985 (IGLD 1985) as a reference.

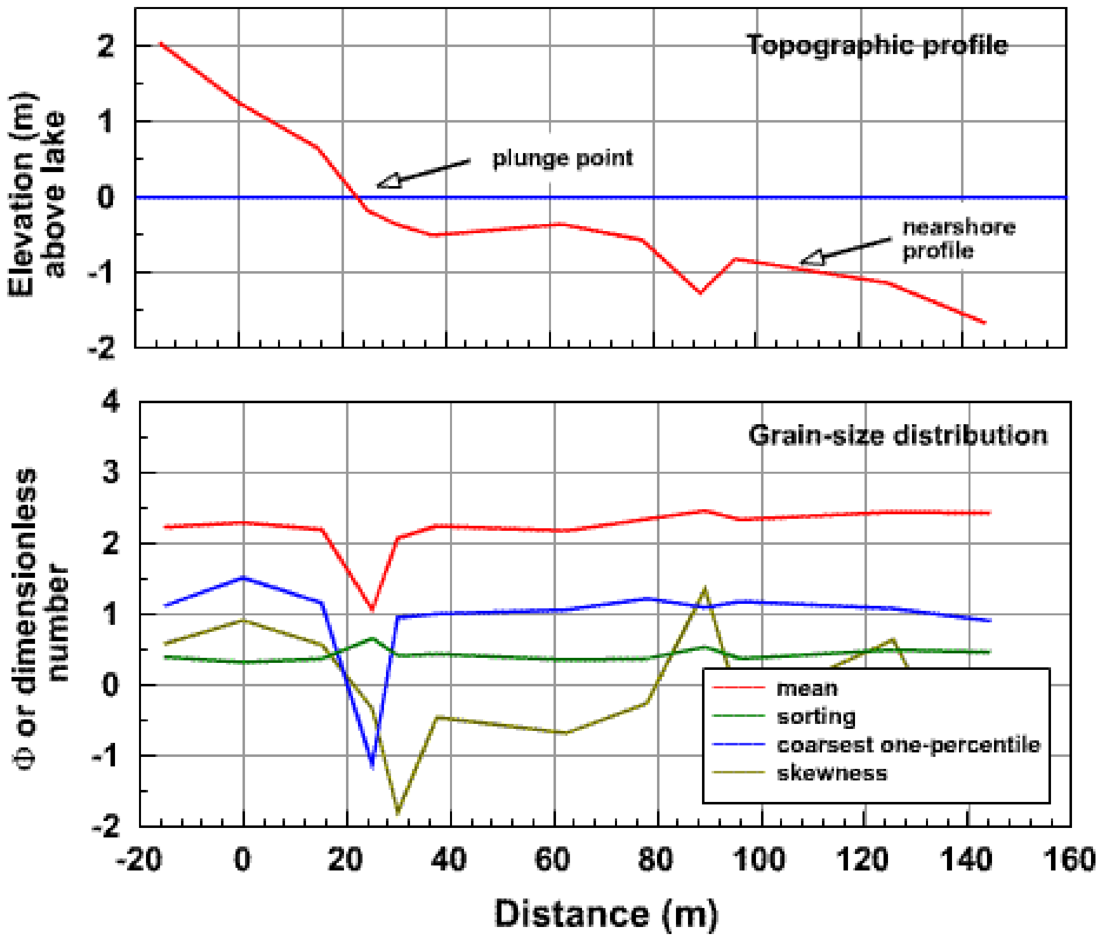

Figure 1. Graphs of beach profile (upper) and grain size (lower) for Lake Street Beach in the Indiana Dunes National Lakeshore. The graphs show that the coarsest and most poorly sorted part of the nearshore are at the plunge point (base of the swash zone). Similar relationships occur at all of the study areas. Note: Coarser sediments are more negative; and poorly sorted sediments are more positive.

Where suitable, basal organic sediments were collected from swales between the ridges. These samples were radiocarbon dated with a $*C^{13}$ correction and calibrated to calendar years (Stuiver and Reimer, 1993). Five relative lake-level curves were created (Thompson and Baedke, 1997).

Beach ridges at the five sites developed at an average rate between 29 and 38 years/ridge (33 yr/ridge average). A pair-wise t-test indicates that none of the average timings of beach-ridge development are statistically different than each other. The similarity in average timing suggests a periodic, or more likely quasi-periodic, lake-level fluctuation of about 33 years occurred in the Lake Michigan basin during the late Holocene (Thompson and Baedke, 1997). The five sites also contain groups of ridges that show a rise and fall in foreshore elevation across four to six ridges (commonly five). These groups of ridges indicate another quasi-periodic lake-level fluctuation at between 120 and 190 years (~150 years). Both of these quasi-periodic fluctuations are superimposed on long-term lake-level change associated with water volume changes in the basin and isostatic rebound. When rebound is removed from the five relative lake level curves and the data generalized using a Fast Fourier Transfer (FFT) smoothing, the resulting, “eustatic,” lake level curve shows the upper limit of late Holocene lake level for the Lakes Michigan/Huron basin that was observed at the Port Huron outlet and, for a short period of time, the Chicago outlet during the late Holocene (Figure 3).

The curve shows that lake level was very high 4,500 years ago. This high stage is often called the Nipissing II phase of ancestral Lake Michigan. Between 4,500 and 3,500 years ago, lake level fell about 4.5 meters to an

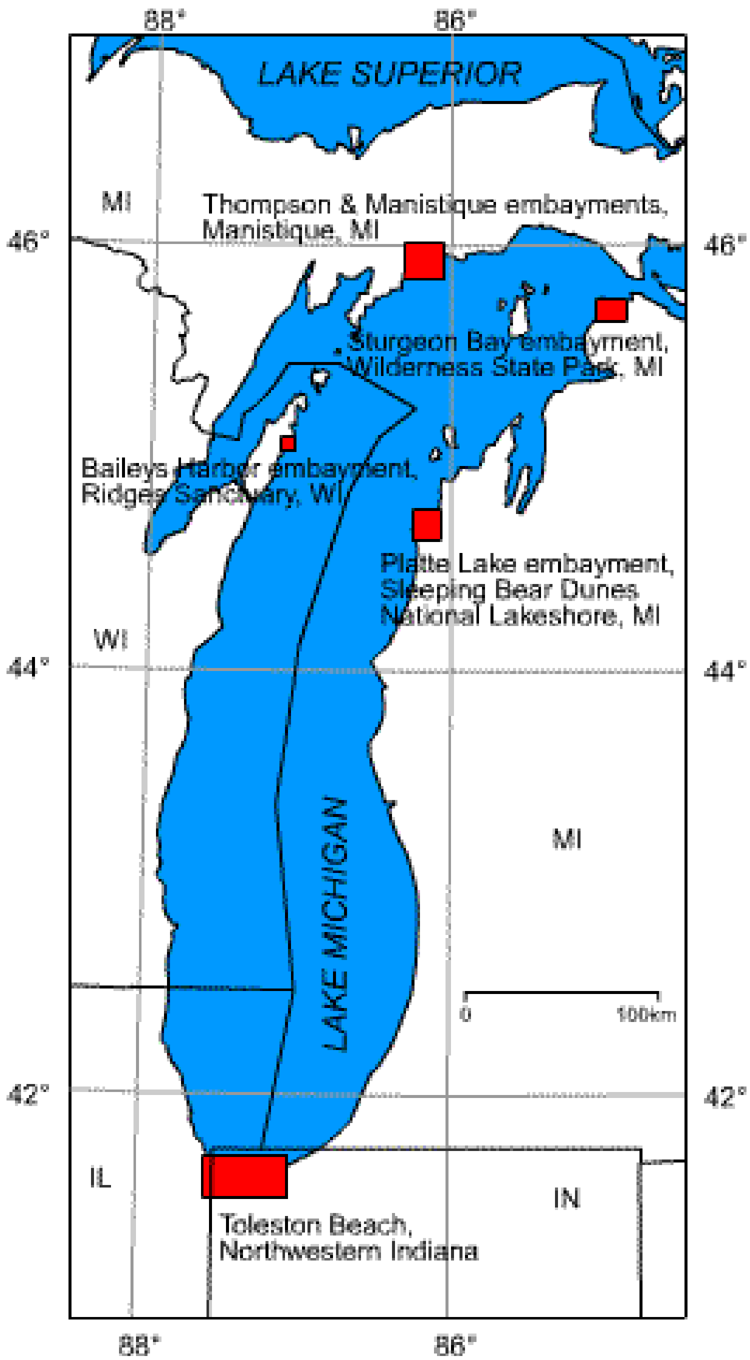

Figure 2. - Map of Lake Michigan showing areas studied (boxes). See Thompson and Baedke (1997) for detailed maps of each area.

elevation that is similar to the mean elevation of the historical record from 1819 to 1990. From 3,500 to the present, the upper limit of lake level reached elevations from .5 to 1 m above and about .5 m below the historical average. Two notable high phases occurred from 2,300 to 3,100 and from 1,100 to 1,900 calendar years ago. The older high lake level period corresponds to a lake level phase known as the Algoma phase of ancestral Lake Michigan. The younger high lake level period is currently unnamed.

Pervasive in the curve from 3,100 to the present are the 150-year quasi-periodic lake level fluctuations. Although present in the raw data, this fluctuation is not seen in the curve older than 3,100 because the FFT smoothing could not pick it out from the single data set that defines this part of the graph. It is important to note that the most recent 150-year fluctuation can be connected directly to the historical record, suggesting that the entire historical record also represents one 150-year quasi-periodic fluctuation. It should be noted that Figure 3 does not show the 30-year quasi-periodic fluctuations. The upper limits of these fluctuations are the points that define the 150-year events. The 30-year fluctuations, however, are observable in the historical record for Lake Michigan.

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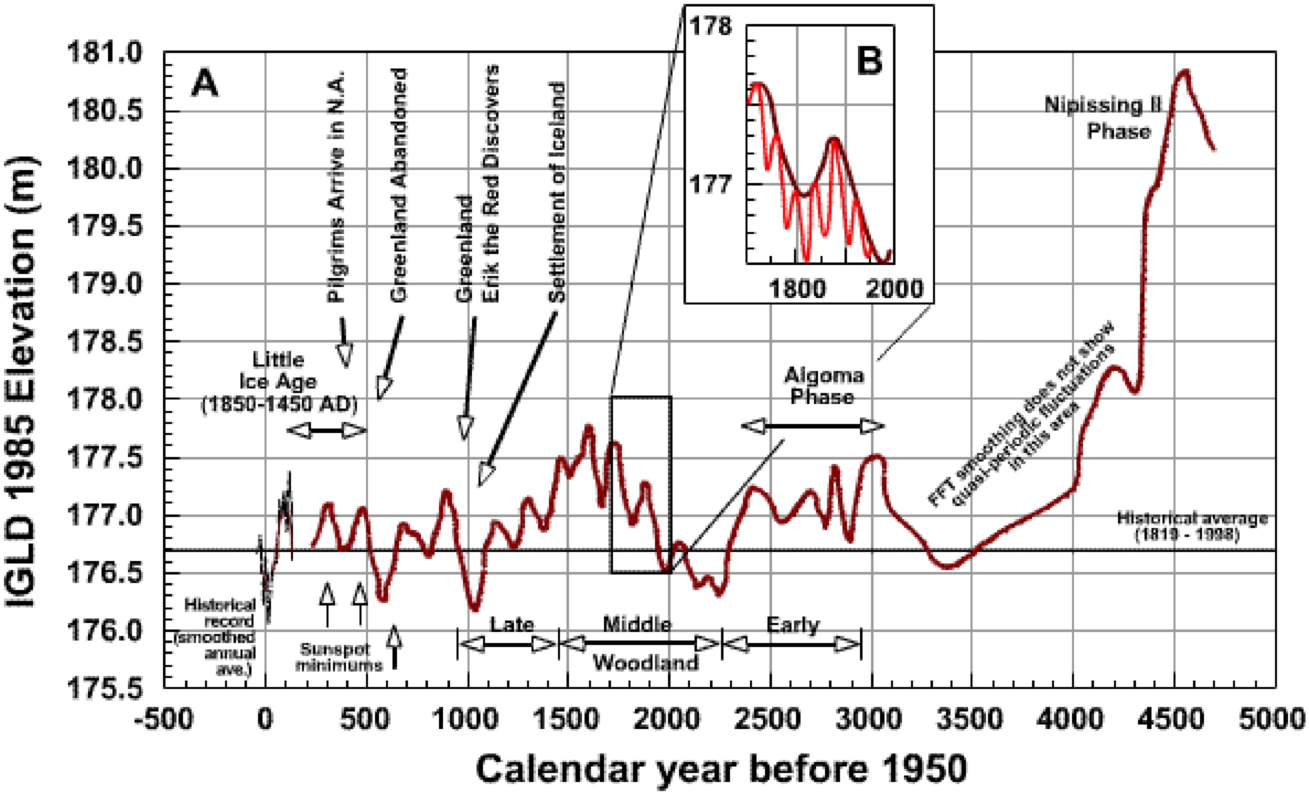

Figure 3. A. Graph of late Holocene and historical lake level for Lake Michigan. Historical events, solar sunspot minimums, and an archeological timescale are plotted for comparison to the curve. B. Expansion of part of the late Holocene lake level curve, showing the relationship of 150-year (dark red line) and 30-year quasi-periodic (light red line) fluctuations.

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DISCUSSION

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To identify a range of water level fluctuations for the present Great Lakes, it is important to understand the factors that affect these fluctuations. Postglacial factors that affect water level fluctuations include: (1) climatic conditions, (2) anthropogenic influences, and (3) geologic activities. For the purpose of this discussion the geologic and climatic processes will be discussed interchangeably since, in some cases, climatic processes cause geologic activities (advancement and retreat of Pleistocene glaciers was responsible for isostatic rebound), and geologic activities cause climatic processes (cooling of the atmosphere due to the injection of volcanic dust).

Climate Conditions

Past and present climates continually influence Great Lakes water level fluctuations. Table 1 shows prehistoric as well as historic changes in climate. Evidence of climate change is all around us. Over years, decades, and centuries, insects and birds have changed their migrating patterns, vegetation and landscape features have changed and, on a longer time scale, evidence of glaciers indicate cooler climates.

During the Pleistocene Epoch, temperatures were cooler and glaciers were present in the Great Lakes region. Then post-glacially, isostatic rebound was responsible for elevating land and lake surfaces and consequently changing the shape of the drainage basin. Isostatic rebound is based on the concept of isostasy, which states that land masses must maintain an equilibrium within the earth’s crust. A tremendous weight, such as a regional glacier, can theoretically force localized landmasses downward into the crust until a compensatory equilibrium is achieved. Once that weight is removed, as in a glacial retreat, the landmass will rebound upward to once again achieve equilibrium. The Laurentide glacier of the Pleistocene Epoch depressed the crust in the Great Lakes area differentially because the thickest ice was in the northeast. Geophysical measurements and comparison to modern ice caps indicate the ice was 3,000 meters thick beneath the center of the glacier with varying thicknesses throughout (Clark and Stern, 1968). Due to the varying thicknesses of the ice cover over the Great Lakes region, differential isostatic rebound rates are location specific. Figure 1 shows the vertical movement rates per century for the Great Lakes - St Lawrence River Basin (adapted from Clark and Persoage, 1970; Larsen, 1987). Rates of rebound vary from north to south in the Great Lakes region, with the Superior basin having the highest rate and Erie with the lowest rate.

Isostatic rebound affected the lake levels of the individual lakes as well as their drainage patterns. Larsen (this volume) found that Lake Superior was confluent with Lakes Michigan and Huron approximately 5000-2100 years BP. After 2100 years B.P., Lake Superior’s bedrock floored outlet channel isostatically rebounded above Lake Huron, thus separating it from the two lower lakes. Superior continues to uplift at a present rate of .33 meters per century. Both Lewis and Coakley (this volume) suggested that due to crustal rebound at the Niagara River outlet combined with post-Algonquin low water levels, Lake Erie was at one point a terminal lake (no outlet drainage) during the early Holocene, approximately 9000-11000 years ago. Presently, the eastern end of Lake Erie is rising at 9 cm per century with respect to the western end.

Although isostatic rebound continues to account for water level fluctuations of the present Great Lakes, progressive changes in climate is the dynamic force behind the present water level changes. Historic climate conditions in

Table 1. Paleo Climate (Modified from Thomas, 1968).

Pre-Historic & Historic Time	Years	Conditions
Pre-Pleistocene	6,000 ma - 1 ma B. P.	Several severe glaciating of the earth's surface lasting approximately 1 ma and occurring at 250ma intervals.
Pleistocene	1ma - 17,000 B.P.	At least 4 ice ages with interglacial ages occurring between the advances. Ended at 17,000 BP with a sudden warming of the Atlantic Ocean.
Postglacial and Historic	7600 - 4500 B.P.	Climatic Optimum. Temperatures average about 4 degrees warmer than they do now.
Postglacial and Historic	4500 B.P.	Warm Dry Conditions Prevailed
Asiatic Drought	3000 B.P.	Asiatic droughts were such that great migrations took place in Europe
Viking Period	1100 - 800 B.P.	Viking Period. Relatively warm, the coasts of Iceland and Greenland were ice-free. Agriculture was carried on in Greenland.
Little Ice Age	800 - 250 B.P.	1680 was known as the year of the Great Frost in England, and Iceland's climate deteriorated to the extent that authorities considered whether or not the island should be evacuated.

Ma = million years
BP = before present

the Great Lakes show two distinct precipitation regimes; lesser precipitation existed from 1900-1940 and greater precipitation dominates from 1940-1986 (Great Lakes Commission, 1986). Globally, the mean surface temperatures have shown a definite increase since 1940 (Figure 2). This trend is also apparent nationally (Figure 3).

Presently, in 1998 global temperatures were the warmest on record (National Climatic Data Center, 1998a). Nationally, the first two months of 1998 were the warmest and wettest for the contiguous U.S. (National Climatic Data Center, 1998b). The warmest on record took place in the Midwest and East. Warm states included Minnesota, Wisconsin, Illinois, Michigan, Ohio, and Pennsylvania (six of the eight Great Lakes states). Additionally, University of Massachusetts-Amherst researchers found from their studies of tree-rings and ice cores, that the decade of the 1990s was the warmest of the millennium (UMASS News, March 3, 1999).

Contrasting trends are noted globally by decreased precipitation (Figure 4) and nationally, by increased precipitation (Figure 5).

Although the underlying cause for these climatic trends is still under investigation, scientists identified astronomic and geologic phenomena as some of the driving forces behind these changes. All climatological processes are driven by the redistribution of solar energy. The sun, the heat engine that drives the circulation of our atmosphere, is a varying source of energy. Over its 11-year cycle, variation in the sun's base energy output has been measured up to .5 percent (Space Environment Center, 1999). Scientists believe that this variation is significant and can modify climate over time. Historically, various solar events have been associated with the Earth's climate; the most prominent event being the waxing and waning of sunspots. Sunspots appear with a periodicity of an 11-year cycle (one solar year). Sunspots, prominent dark spots on the solar surface, are made up of transient, yet concen-

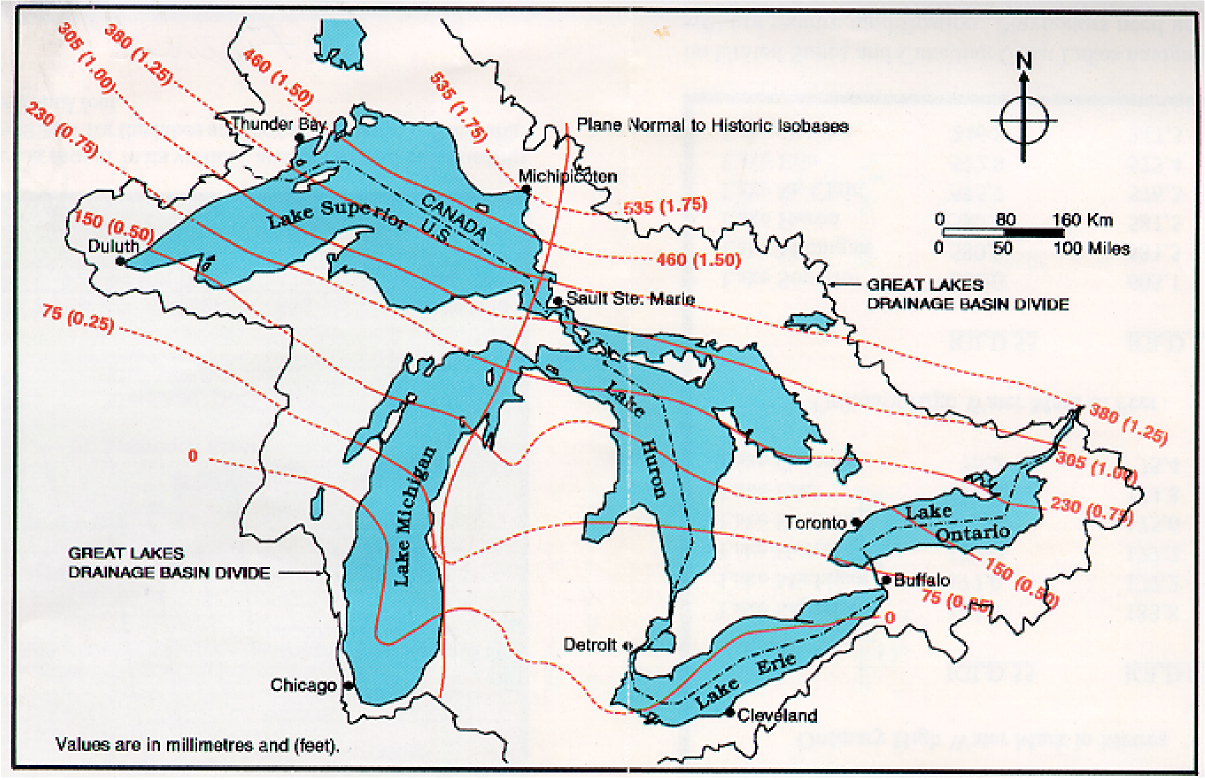

Figure 1. Vertical movement rates per century for the Great Lakes-St. Lawrence River Basin (adapted from Clark and Persoage, 1970; Larsen, 1987). For example, Michipicoten,Ontario is rising relative to Chicago, IL at a rate of approximately 535 mm (1.75 feet per 100 years.)

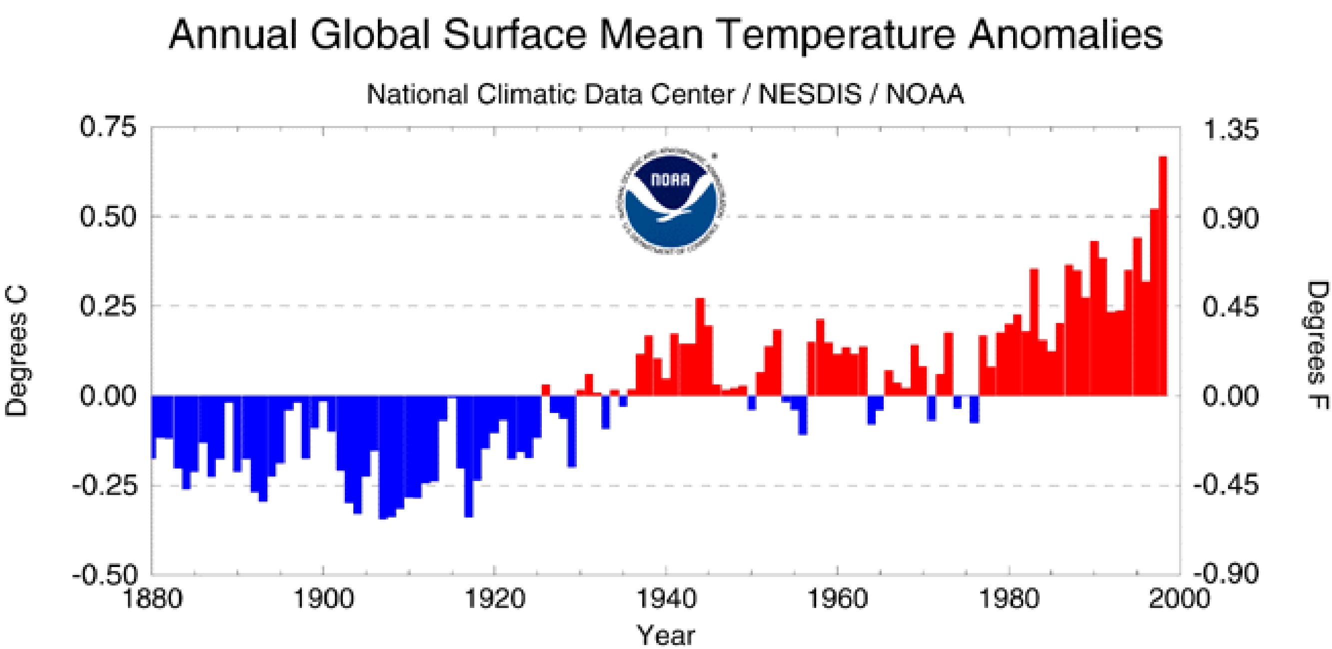

Figure 2. Annual Global Surface Mean Temperature Anomalies. NOAA, National Climatic Data Center (1999).

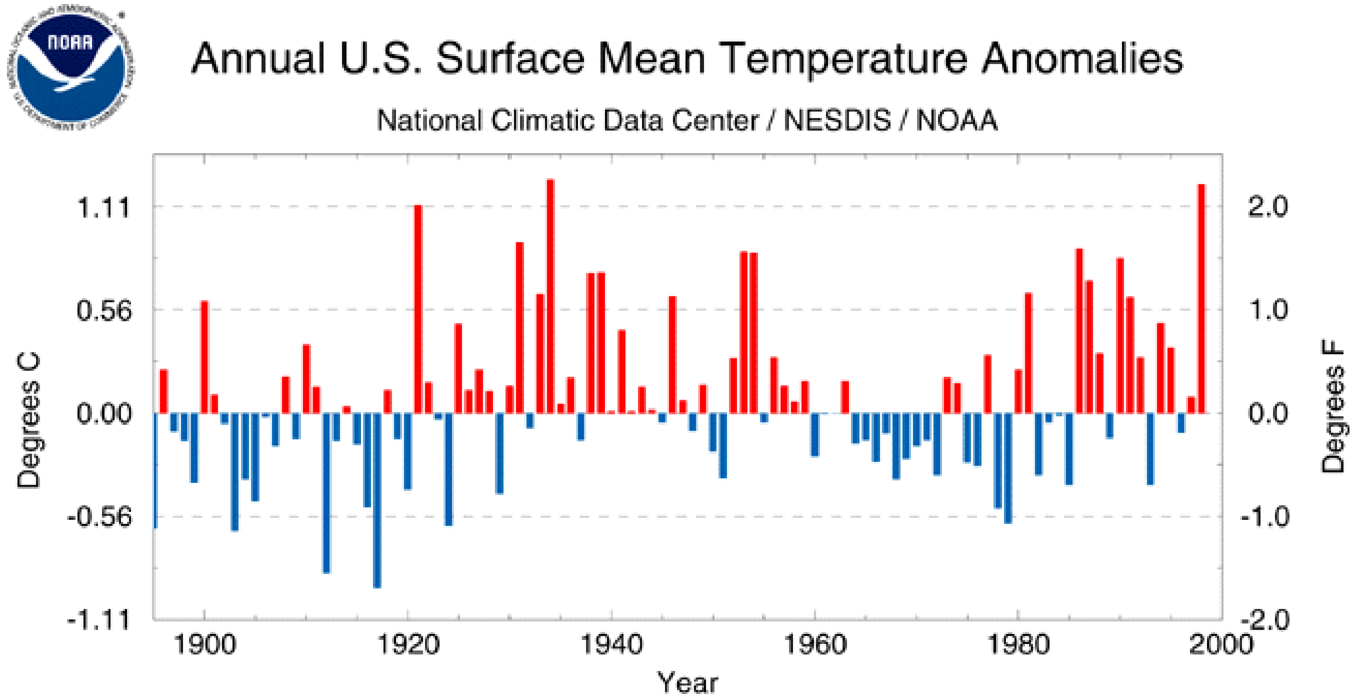

Figure 3. Annual U.S. Surface Mean Temperature Anomalies. NOAA, National Climatic Data Center (1999).

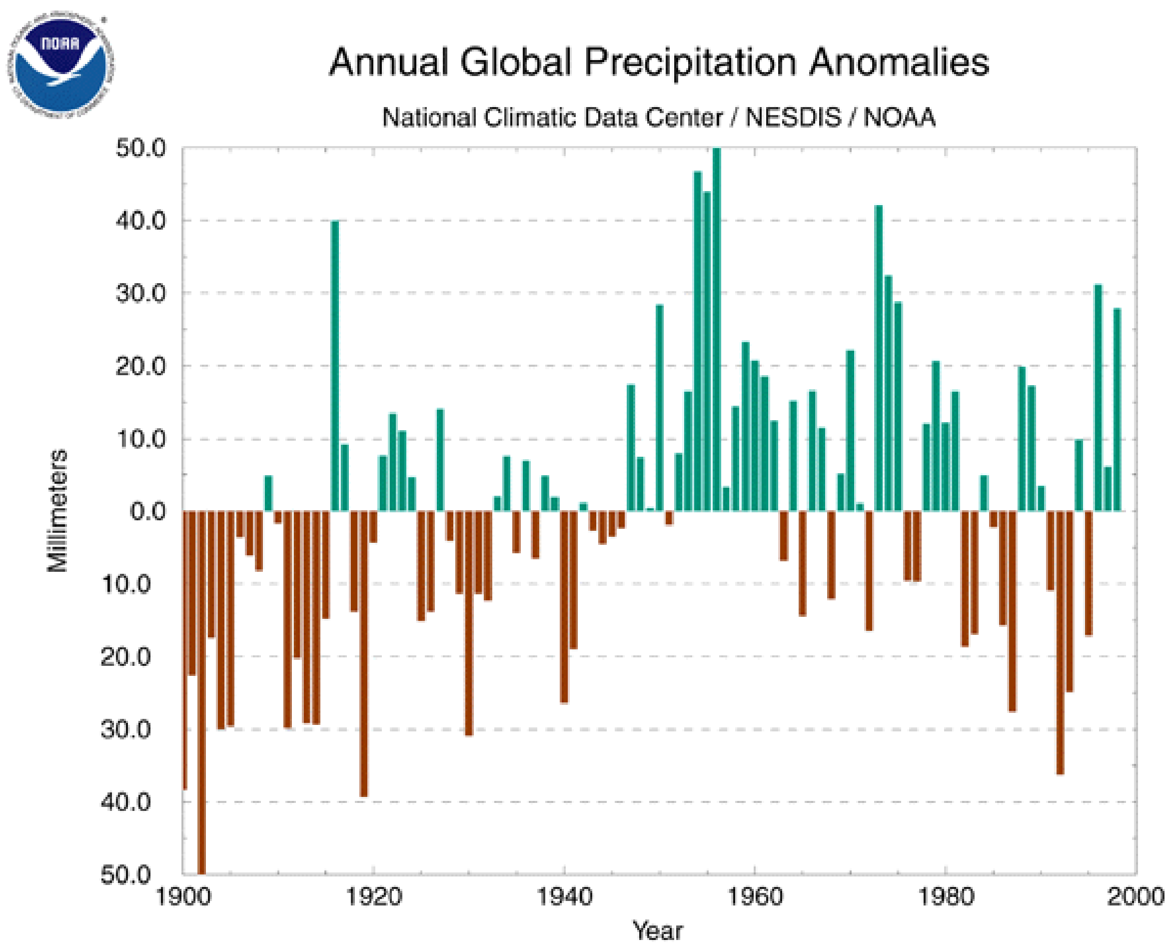

Figure 4. Annual Global Precipitation Anomalies. NOAA, National Climatic Data Center (1999).

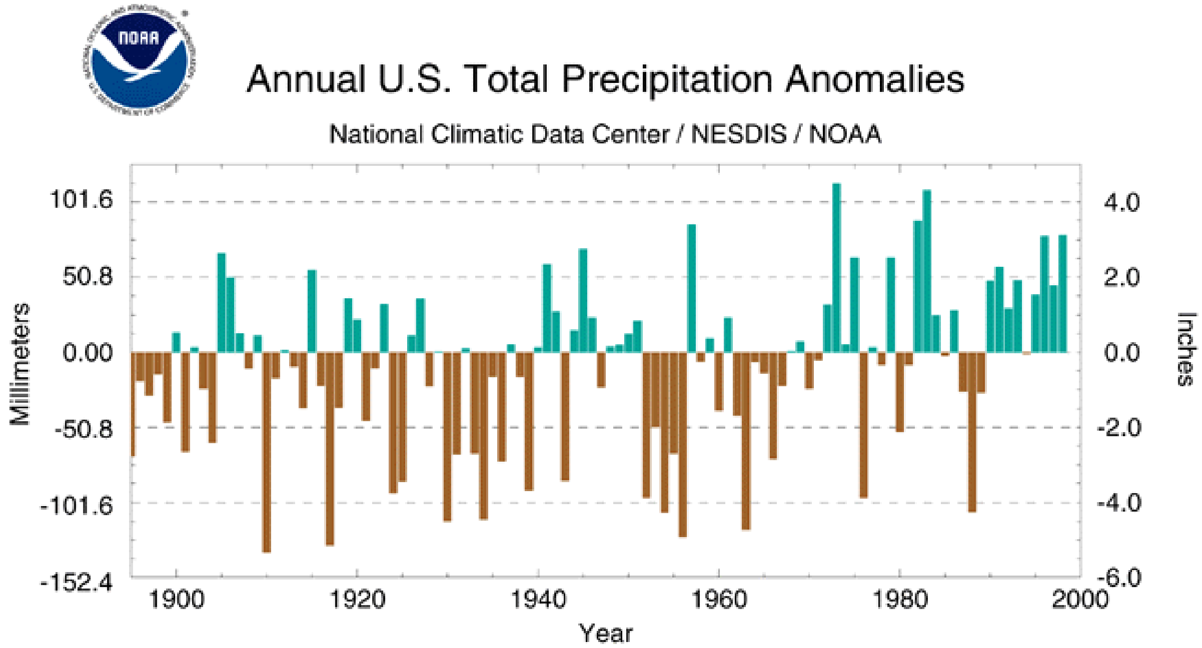

Figure 5. Annual U.S. Total Precipitation Anomalies. NOAA, National Climatic Data Center (1999).

trated magnetic fields. The amount of solar activities, such as solar flares and coronal mass ejections, which introduce massive amounts of solar materials into the solar system, are closely related to the typical number of sunspots that are visible (The Regents of the University of Michigan, 1999). Although the solar cycle has been quite regular for the last 300 years (Space Environment Center, 1999), in the 17th and 18th centuries there were fewer than average sunspots. This period coincides with "The Little Ice Age" in Europe. Alexander (1995) cited a correlation between the deviations of the mean annual river flow and mean annual sunspot numbers. He suggests that the perturbations in solar activity affect the phenomena that control the extreme events of floods. Morner (1993) contributes global climate change in part to four sunspot minima that occurred in the last 500 years. Presently, the last solar maximum was in 1989; the next will be in the year 2000.

Redistribution of solar radiation isn't only dependent on solar activity, but also the earth's atmosphere. The atmosphere shields us from solar radiation and variations in its transmissivity due to the relative presence or absence of dust, carbon dioxide, vapor pressure, ozone, and clouds, has an effect on the earth's climate. For example, in 1883, the eruption of the Krakatoa Volcano blanketed the upper atmosphere with volcanic dust and was blamed for the cooler conditions of the 1800s. Additionally, the presence of carbon dioxide in the atmosphere prevents the loss of terrestrial radiation into space, causing a rise in overall temperatures (the greenhouse effect).

Overall, the Holocene has been warmer than the majority of the past million years (Eichenlaub, 1979). However, some think that this warm period may be an interglacial period. Although the Holocene was globally marked by a warming of the Atlantic Ocean, in the Great Lakes region, it was identified when westerly air masses from the Pacific became more frequent, and July temperatures increased (Webb and Byron, 1972). The present climate of the Great Lakes is dominated by three air masses: *maritime tropical* (mT), *maritime pacific* (mP), and *continental polar* (cP), Figure 6.

Lewis (this volume) recently found that dry climates of the Holocene progressed eastward across the southern Great Lakes basin, appearing in the Ontario basin as late as 2000 years ago. He explains these climatic phases as changes in air mass circulation. Lewis further points out that the water balance is an important factor in the drainage basin. He states that a negative water balance can switch a lake from an open overflowing system to a closed one, driven by evaporation. For instance, Lewis indicates that the Lake Erie basin was closed in the early

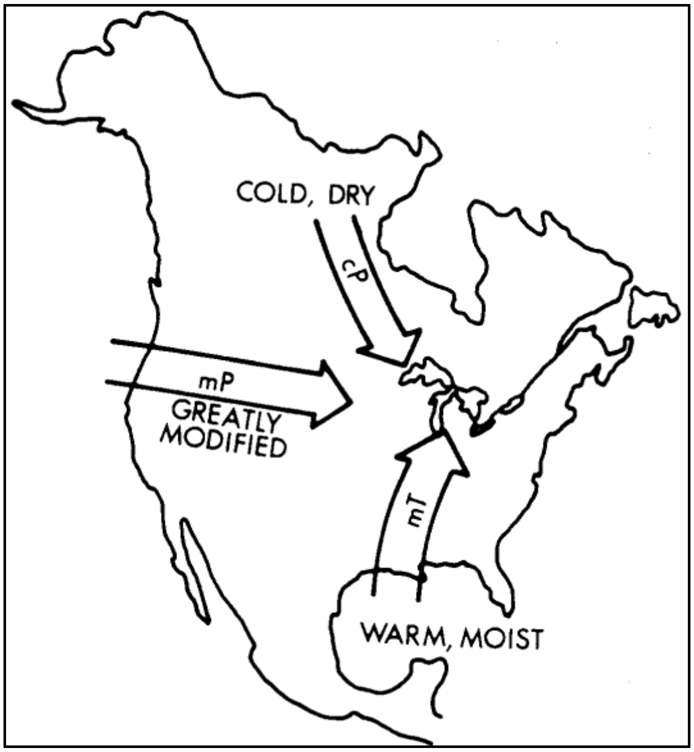

Figure 6. Air masses affecting the Great Lakes.

mid Holocene possibly due to no inflow from the upper lakes and high evaporation caused by a change in air mass circulation.

Larsen (this volume) attributes high and low prehistoric levels of Lake Superior and Michigan to changes in climate. Larsen cites periods of high levels that occurred about 1,100, 400, and 200 years ago with low level episodes about 650 and 350 years ago with a range of more than ±1.0 meter that is centered about a long term mean for Lakes Superior and Michigan. These high and low lake levels are further correlated with weather related events. Larsen associates the "Little Ice Age" with the highs preceding historic record, and one major low period is associated with the "Medieval Warm Period" of northern Europe.

Larsen (1973) compared the Camp Century Climate Trend (Dansgaard et al., 1971) to 5-year mean Lakes Michigan-Huron water levels. Results of this comparison suggested a projected cooler period (associated with higher lake levels) to take place from 1983-1988. Since Lakes Michigan-Huron levels reached record highs for a few months in 1986, this projection was quite accurate. Additionally, his results suggested lower lake levels by the year 2015 when the air temperatures are projected to reach a thermal maximum.

From the above discussion it is apparent that not only does climate play an important role in lake level changes, but also the scale of climate fluctuations are equally important. Therefore, in order to predict a range in water levels for the next 100 years, it is not only important to take into consideration the recorded historic data, but it is equally important to interpret climate on the mesoscale (previous 4000 years) to identify the overall trends. Accomplishment of this task requires scientists to interpret non-traditional data such as tree-rings, buried reefs, and sunspots, to name a few.

Anthropogenic Changes

Although nature has had major effects on lake level fluctuations, humans have also made changes in the lakes' regime. Changes that increased runoff to the lakes range from the clear-cutting of the State of Michigan to the urbanization of the northern Lake Erie shoreline. Changes such as deepening the navigation channels of the St. Clair and Detroit Rivers were responsible for lowering Lakes Michigan-Huron by .18 meter (Derecki, 1985). Additional changes imposed due to diversions, irrigation, hydropower, and water level regulations must be considered in the overall water balance of the Great Lakes.

Water Level Fluctuations and Range

Considering the forces acting on fluctuating water levels, it seems difficult to attribute cyclical patterns to these water level variations. However, through various methods researchers have attempted to associate these patterns with atmospheric and solar events. Thompson (this volume) created a long-term record of these events in which he discovered water levels fluctuating at a 33-year interval that was superimposed on a 150-year interval. Thompson summarizes that lake levels are at the peak of the 150-year fluctuation and are headed back to a low period. Thompson estimated Lake Michigan's water level range for the last 3500 years to be .5-1.0 meters above and .5 meter below the historical average. Larsen (this volume) estimated fluctuations on Lake Superior for the last 2500 years ranged from 1 to 1.5 meters. Lewis (this volume) said that Lake Ontario's water levels fluctuated between 1 and 2 meters for about 1000 years within the last 4000 years.

Future Research Needs

Paleo lake level research should be conducted on all Great Lakes to determine the physical limits and timing of past lake level fluctuations. It was agreed that there should be further research examining the relationship between long term climatic patterns and paleo-lake levels, the following research topics were suggested:

- Look at past air-mass movements with respect to environmental changes. Instead of focusing on the North-South shift, look at the East-West shift in Great Lakes climate patterns.
- Generate lake level responses to climate change.
- Develop global warming scenarios.
- Identify the climatic regime for each lake.
- Research submerged tree-stumps in the Straits of Mackinaw with respect to what they reveal about prehistoric precipitation patterns.
- Understand the ages of submerged reefs in Lake Superior.
- Determine the effects of deforestation on lake levels.
- Develop methods to accurately determine low lake levels.

Findings

- Evidence indicates that the range in historical water levels has been exceeded over the last 3000 years for both Lake Superior and Lakes Michigan-Huron.

- Paleo lake levels for Lakes Ontario and Erie are much harder to quantify but have probably varied around the historical range.

- The geologic record indicates runs of above average and below average lake levels with a quasi-periodic variation around 150 years. There is some belief that we may be moving from the high end of the range towards the low end of the range.

- The research into Great Lakes paleo water levels is very productive and useful for water resource studies. It should be continued with high priority given to Lakes Superior and Michigan-Huron.

- There may be a significant correlation between longer-term atmospheric circulation patterns and lake level fluctuations.

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