Evolution of the Great Lakes

Kevin Kincare
United States Geological Survey

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Kevin Kincare and Grahame J. Larson

Introduction

The Great Lakes are one of Michigan's most distinctive geographic features. The familiar shape of the Great Lakes is however, a recent phenomenon. Their present form is the result of a number of geologic factors, such as glacial erosion and deposition, isostatic depression and subsequent rebound due to glacial-ice load (see FOCUS BOX on page 177), distribution of glacial meltwater, and changing lake outlets. In this chapter, we will examine the formation of the Great Lakes, as well as the many changes they went through before becoming the familiar features we know today.

Origin and development of lake basins

The shape and location of each Great Lake has been largely determined by its underlying geology. Most of the bedrock beneath each lake basin is easily-eroded Paleozoic sedimentary deposits (Plate 6), with the exception of Lake Superior, which is a structural basin underlain by complex Middle Proterozoic rocks (Chapter 12). Therefore, prior to Late Cenozoic Ice Age, each lake basin was probably already a river valley draining to an ancestral St. Lawrence River (Fig. 13.1). The structural bedrock highs of the Kankakee arch south of Chicago, and the Findley arch on the west side of Lake Erie (see FOCUS BOX, Chapter 4), would have effectively prevented the pre-glacial rivers in the Great Lakes region sea-floor from draining to the Mississippi River.

Recent climatic evidence from cores of sea-floor sediment and ice caps indicate that at least 40 separate glaciations occurred during the last 2.75 million years (Fig. 6.1). Each glaciation that was extensive enough to reach Michigan probably further eroded the lake basins and altered the Great Lakes. Most of the glacial deposits in Michigan are from the last two of these glaciations (the early and late Wisconsin advances, both <80,000 years ago) and the majority of these deposits are from the final phase of deglaciation (<20,000 years ago). Older glacial deposits, found to the south of Michigan, in Illinois, Indiana, and Ohio (Fig. 6.2), show that earlier glaciations advanced well to the south of Michigan. However, we have not found any evidence yet for older glacial deposits in Michigan, although this may change with advances in dating techniques and additional field research. Although the glaciers were agents of erosion each time they covered Michigan, our state still has some of the thickest glacial deposits in North America, up to 365 m thick in northern Lower Michigan (Rieck and Winters 1993, Soller 1998; Fig. 17.8). Given this fact, our discussion will center on deposits of the late Wisconsin glaciation.

Glaciers and lakes

The Laurentide ice sheet is a term used for the continental glacier whose center was in the Hudson Bay area and spread out, to eventually cover much of Canada and the northern United States. A second ice sheet, the Cordilleran

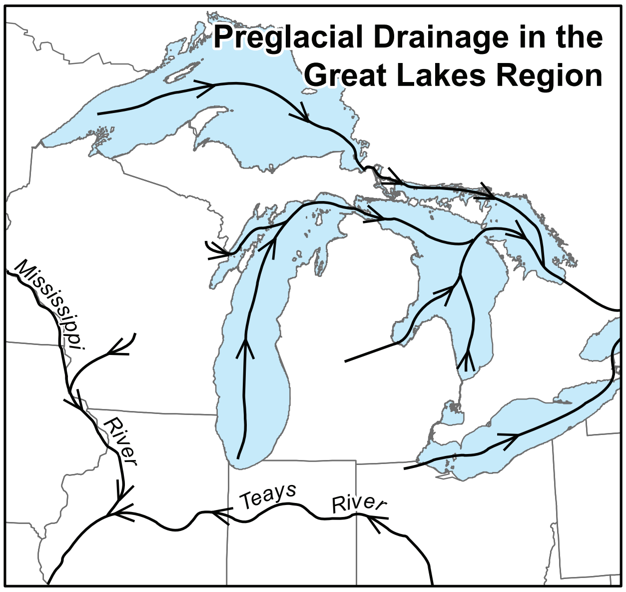


FIGURE 13.1 Preglacial drainage in the Great Lakes region during the Late Cenozoic, prior to glaciation. Probable river patterns are based on known bedrock structures and geophysical reconnaissance of buried valleys. After Hough (1958) and Flint (1971).

FIGURE 13.2 Farthest extent of Late Wisconsin ice advance, which occurred about 23,000 yrs ago in Illinois and Ohio, but later at other points along the ice margin.

ice, formed in the northern Rocky Mountains and, in places, merged with the Laurentide ice sheet. Three major lobes of the Laurentide ice sheet moved across and covered Michigan from out of the overdeepened lake basins and are named after those basins, the Lake Michigan, Saginaw, and Huron/Erie lobes (Figs. 17.4, 17.12). The farthest extent occurred during the Last Glacial Maximum (LGM) and was well to the south in Indiana and Ohio (Fig. 13.2).

The ice at the LGM began to retreat around 21,500 years ago. Evidence from the cross-cutting relationships of former ice margins and their related outwash shows that the lobes did not move synchronously (Mickelson et al. 1983, Kehew et al. 1999). The thinner, weaker Saginaw lobe probably retreated from south-central Michigan first, leaving fine-grained deposits in the St. Joseph River valley, from a lake that was trapped by the Lake Michigan and Huron/Erie lobes which were still much farther south (Fig. 13.3). At this time, before the Great Lake basins opened up, meltwater drainage was down the Wabash and Kankakee Rivers (Fig. 13.4).

The pull-back of the ice from the LGM did not proceed as a long, steady retreat, but was uneven and punctuated by a series of rapid retreats and readvances (Chapters 6, 17). This complicated (and sometimes confusing) series of events led to numerous proglacial and post-glacial lakes within each Great Lake basin. At times, these lakes were separate, while at other times they were confluent or connected by river channels.

Lakes are often associated with glaciation, particularly during glacial retreat, because drainage can easily become blocked by ice on one side and by landforms of a previous ice advance on the other. Therefore, knowing the record of glacial retreat (Chapter 6) is important for understanding the sequence of glacial lake levels. Leverett and Taylor (1915) produced the most detailed compilation of field data on glacial lake phases in the Great Lakes basin—almost 100 years ago!

Lakes can produce a definitive set of features based on particular types of depositional and erosional patterns. The low-energy environments of lakes generally lead to fine-grained deposits, e.g., silt and clay, on their beds. Two major exceptions to this are sandier shoreline areas (high energy due to wave action) and river mouths (where deltas are typically built).

A lake, by definition, has the same surface elevation at all points. On this basis, shorelines from ancient lakes can be traced from place to place, even if intervening parts of the shoreline no longer exist due to erosion or burial. Deltas can also show a former lake-surface elevation where a river entered a lake, as well as showing the off-shore connection to sediments of deeper, quiet-water deposits. Lakes have the power to erode as well. Many shorelines are not traced by beach deposits, but by wave-cut bluffs (strandlines) eroded into adjoining headlands, as well as spits, which may contain the eroded material, transported into the spit (Fig. 13.5). River channels that are now dry

FIGURE 13.3 Map of the ice margin about 19,000 years ago, showing the Saginaw lobe reentrant.

FIGURE 13.4 Map of the ice margin about 18,000 years ago, showing meltwater drainage overflowing to the Wabash and Kankakee Rivers.

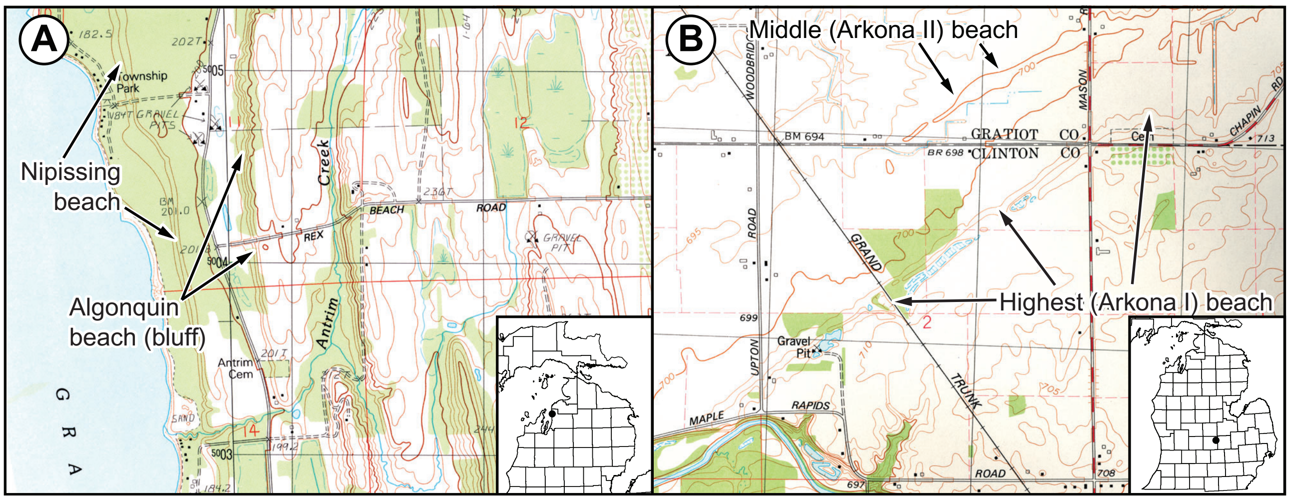


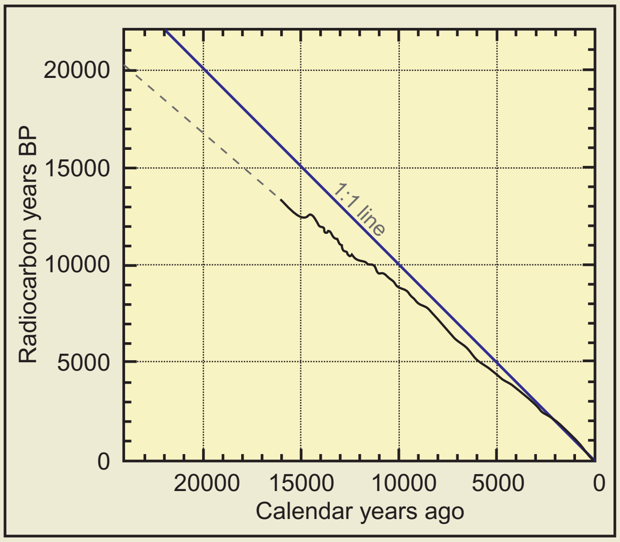
FIGURE 13.5 Topographic maps showing strandlines and/or beach ridges of some former glacial lakes in Michigan. A. Well-developed shorelines inferred from wave-cut bluffs. Two shorelines are visible on this map, as long, steep banks that were cut by prolonged wave action. The lower bluff is from the Nipissing phase; the higher one is from Glacial Lake Algonquin. The bluff east of Antrim Creek in not a shoreline, its base rises upstream and it lacks a flat platform on its shoreward side. Note the prominent drumlins on the east edge of this map. Source: Atwood 1:25,000 quadrangle, 5 m contour interval. B. Shorelines inferred from more subtle beach ridges. Two beaches (highest and middle Lake Arkona) are seen here as linear ridges, often with interconnected, closed contours, on the topographic map. These features are accumulations of sand, formed due to wave action and longshore transport at the former shoreline. Sandy beach ridges are preferred building locations, due to better drainage than adjacent clayey lake bottom sediments. In this case, the beaches are being used for a cemetery (a high, dry, sandy location on the otherwise wet lake plain) and, in places, it is being mined for sand and gravel. Source: Ovid West 1:24,000 quadrangle, 5 ft contour interval.

FOCUS BOX: Radiocarbon years vs. calendar years

In the scientific literature, one may encounter ages for events given in radiocarbon years BP (before present), as well as in calendar years ago. To make things more confusing, these two methods seemingly express different ages for the same geologic event. To make sense of this, a little history and explanation are needed.

The discovery in the early 20th century of a radioactive isotope of carbon (^{14}C) was important for the study of recent deposits because its relatively short half-life (5730 yrs) made it ideal for determining the ages of organic material younger than about 40,000–50,000 years old. It forms as a result of cosmic rays interacting with nitrogen in the atmosphere. Most importantly, because ^{14}C is in the atmosphere, all living things have ^{14}C incorporated into their tissues. When an organism dies, it no longer assimilates ^{14}C, so its ^{14}C mass begins to decrease by radioactive decay. Determining the radioactivity that remains in a dead organism allows us to approximate the elapsed time since it died.

The study of *dendrochrononlogy* also allowed scientists to determine the age of trees, and ancient wooden materials, like beams and logs, by matching patterns of their annual growth rings. Unfortunately, the tree-ring chronology, which goes back around 8000 yrs (Stuiver 1978), did not match the ^{14}C chronology because the ^{14}C chronology was based on the assumption that the initial concentration of ^{14}C in the atmosphere has always been constant. Not only has the concentration of ^{14}C in the atmosphere changed over time, but more ^{14}C is produced at higher latitudes, and there are biological and chemical energy pathways that favor uptake of lighter isotopes (^{12}C and ^{13}C) over ^{14}C (Faure 1986). Therefore, a calendar year is constant, but a "radiocarbon year" can vary over time (see Figure). Study of these problems has led to the development of radiocarbon year corrections (or calibrations), not only by using tree ring data but also with corals, ice cores, and marine-sediment cores that are applied to ^{14}C ages to calibrate them to calendar years. Ages reported in the literature now specify which is being used. Radiocarbon ages are typically reported in years BP. The "present" is assumed to be 1950—before the open-air testing of atomic weapons, which sent the ^{14}C content of the atmosphere off the scale. Currently, all but the oldest ^{14}C dates can be readily calibrated to calendar years. All radiocarbon dates mentioned in this chapter have been converted from radiocarbon years BP to calendar years ago, using the calibration curve of Fairbanks et al. (2005).



Graph of calendar years vs. radiocarbon years. By reading the graph one can see, for example, that a sample from a tree that died 12,500 radiocarbon years BP is approximately 15,000 calendar years old. After: Quaternary Isotope Lab, Univ. of Washington.

or have rivers too small to have cut its channel provide evidence of old glacial lake outlets. Each bit of evidence, separately or in combination, is used to map ancient lake margins, and their cross-cutting relationships tells us the order in which they occurred.

Early lakes of the Lake Erie/Huron basin

The history of the lakes in the Erie and Huron basins are so closely tied together that they must be discussed together. Retreat of the Huron/Erie lobe from the Fort Wayne moraine around 17,400 years ago formed the first important proglacial lake of the Great Lakes basin—Glacial Lake Maumee (Fig. 13.6, Table 13.1). There were three phases of Lake Maumee (I, II, and III) which initially filled the lowland west of Lake Erie and spilled over the drainage divide

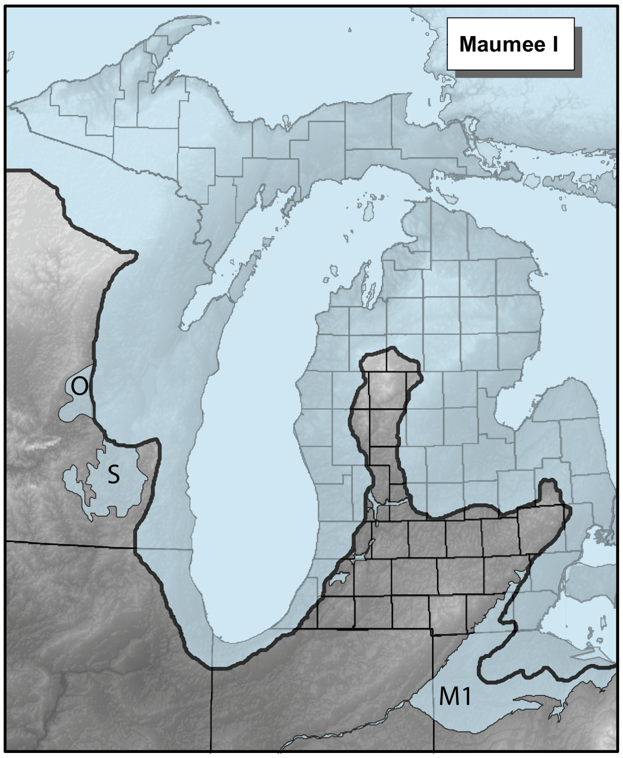


FIGURE 13.6 Map of the ice margin about 17,500 years ago, showing Lake Maumee I (M1). Glacial Lakes Scuppernong (S) and Oshkosh (O) are shown in front of the Green Bay lobe in Wisconsin (Clayton 1997).

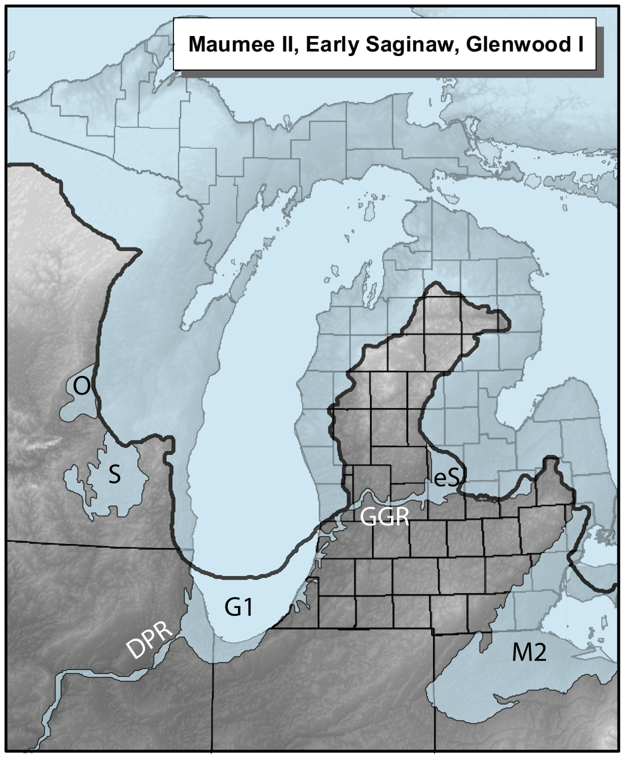


FIGURE 13.7 Map of the ice margin about 17,100 years ago, showing Lake Maumee II (M2), Early Lake Saginaw (eS), and the Glenwood I phase of Lake Chicago (G1). At this time, the Glacial Grand River (GGR) drained Early Lake Saginaw into Lake Chicago, and the Des Plaines River drained Lake Chicago to the Mississippi River drainage basin (via the Illinois River). Glacial Lakes Scuppernong (S) and Oshkosh (O) are shown in front of the Green Bay lobe in Wisconsin (Clayton 1997).

to the Wabash River near present day Fort Wayne, Indiana (Calkin and Feenstra 1985). The Maumee I shoreline (the elevation of the outlet and associated beaches) was at 244 m above sea level and, as the Huron/Erie lobe continued its retreat, extended from Fort Wayne into Michigan through Lenawee County to Macomb County. After the Huron/Erie lobe ice readvanced to the Defiance moraine, Lake Maumee was present only in northeastern Ohio. Ice retreat again allowed expansion of Lake Maumee into southeastern Michigan. But the retreat also uncovered a lower outlet, lowering Maumee I to the Maumee II elevation of 232 m (Fig. 13.7). Shoreline features of Maumee II are not well developed in Michigan and are known mostly from Ohio (Calkin and Feenstra 1985). There is no outlet at 232 m elevation across the Defiance moraine, so it is thought that the ice advance ending Maumee II (and building the Flint moraine) buried the outlet channel (Leverett and Taylor 1915). With the lower outlet buried, the previous and higher (244 m) outlet was used again. Ice retreat from the Flint moraine opened up an impressive channel near Imlay City (Lapeer County) allowing Maumee III to stabilize at 238 m and drain eastward into the Saginaw lowlands. This well-developed channel is clearly visible on a topographic map of Lapeer County, extending from the Lake Maumee plain to the Flint River.

Retreat of the Saginaw lobe margin from the Flint moraine also started opening up the Saginaw lowlands in central Michigan, forming Early Lake Saginaw (Fig. 13.7). This lake drained westward via the Glacial Grand River, through the Maple River valley at the village of Maple Rapids, to eventually empty into Glacial Lake Chicago, Glenwood I phase. The Allendale delta, just east of Grand Rapids, marks the location of the mouth of the Glacial Grand River at this time (Bretz 1953). Strandlines of Early Lake Saginaw exist at 225 m and 222 m in northern Shiawassee County. After the Huron lobe retreated out of Michigan's thumb, Maumee III merged with Early Lake Saginaw into

TABLE 13.1 Post-glacial lakes and levels from 17,000–2000 years ago in the Lakes Superior, Michigan, Huron, and Erie basins[1]

Years ago	Superior basin	Michigan basin	Huron basin	Erie basin
		Glenwood I (195)		Maumee I (244)
17000			Early Saginaw (225)	Maumee II (232)
				Maumee III (237)
			Arkona (216–212)	
		Mackinaw (170?)	post-Arkona low (?)	Ypsilanti (<166)
16000		Glenwood II (195)	Saginaw (212)	Whittlesey (225)
15000			Warren I & II -Wayne (210-199)	
			Warren III (206-203)	
			Grassmere-Lundy (195-189)	
		Calumet? (189)	Early Algonquin (184)	
14000				Early Erie (<159)
		Two Creeks (170?)	Kirkfield (173?)	
		Calumet (189)	Huron Algonquin (184)	
		Toleston? (184)		
13000		Main Algonquin (184)		
	Duluth (331)			
	Duluth (331)	Algonquin (184)		
		Algonquin (descending)		
12000				
	Duluth (331)/Minong (220)	Algonquin (descending)		
		Chippewa (75)	Stanley (66)	
11000				
	Minong (220)			
	Mattawa flood			
10000				
	Mattawa flood			
		-closed basins?-		
9000		Olson forest bed (153)		
	-Superior basin confluent with Huron basin-			
8000				
			Sanilac forest bed (165)	
7000				
6000		Nipissing (184)		
		-coastal dune building begins-		
5000				
4000				
		Algoma (181)		
3000				
	-Superior basin outlet rebounds above Huron Basin, present configuration of lakes achieved-			
2000	*-lake levels continue to fluctuate due to climate and rebound-*			

1. Compiled from Fullerton (1980), Calkin and Feenstra (1985), Eschman and Karrow (1985), Farrand and Drexler (1985), Hansel et al. (1985), Larsen (1987), Lewis and Anderson (1989), Schneider and Hansel (1990), Colman et al. (1994), Larson et al. (1994), Baedke and Thompson (2000), and Lewis at al. (2007).

FIGURE 13.8 Map of the ice margin about 16,500 years, ago showing Lake Arkona (A), the Glenwood I phase of Lake Chicago (G1) and Glacial Lake Oshkosh (O) in Wisconsin.

Lake Arkona, at an initial elevation of 216 m (Fig. 13.8). Based on shoreline data, Lake Arkona subsequently declined to 213 m and then to 212 m. The reason for the decline in Lake Arkona's level has been variously attributed to decreases in meltwater production during a glacial readvance, outlet downcutting, climate, and the ability of a single outlet to drain a lake at several elevations (Calkin and Feenstra 1985). Well-developed Arkona beaches exist on the lake plain east of Maple Rapids in southeastern Gratiot County, northeastern Clinton County, and along the Shiawassee-Saginaw County line.

Lake Arkona expanded to the north and east as the ice margin retreated, during a period known as the Mackinaw interstade (Eschman and Karrow 1985). Eventually, the ice margin retreated far enough into southern Ontario that it opened isostatically-depressed outlets (see FOCUS BOX below) in southeastern Georgian Bay, thereby draining the Saginaw lowlands. This event left the Lake Erie basin with no meltwater source (the St. Clair River was dry at the time), as well as an isostatically-depressed outlet in the vicinity of the Niagara River. As a result, a very low level lake named Lake Ypsilanti existed for a short time in the Erie basin, which only had drainage from local rivers (Calkin and Feenstra 1985) and probably occupied about half the area that Lake Erie does today.

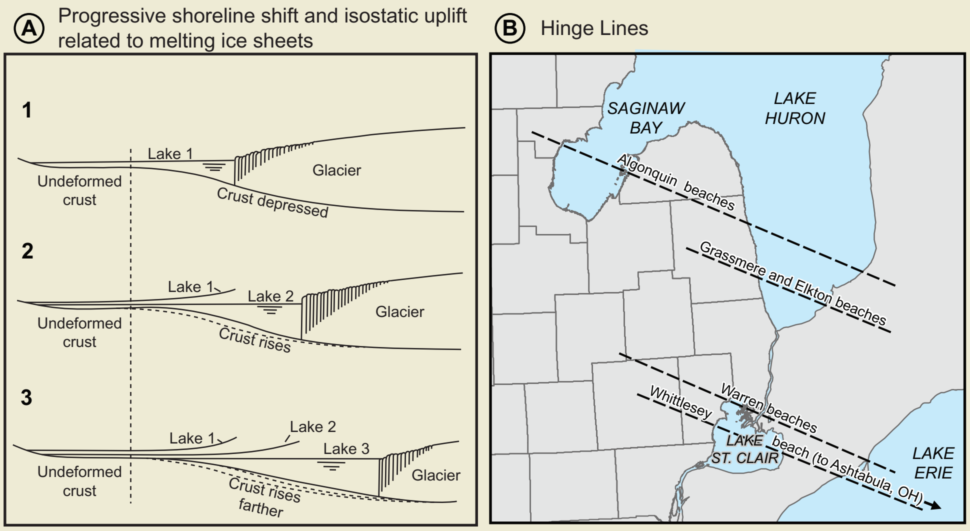
FOCUS BOX: Isostatic rebound

Obviously, the water level of a lake is the same elevation all around its shoreline, making a perfect horizontal "trace" of the water line. Today, however, ancient shorelines of glacial and post-glacial Great Lake phases are often not at the same elevation as which they formed. Instead, they commonly "rise" as they are traced to the north and northeast (Stanley 1936, Deane 1950, Schaetzl et al. 2002). What would cause these shorelines to rise, when they were originally horizontal?

The answer is a concept called *isostasy*. The rigid lithosphere (crust) of the earth "floats" on top of a viscous asthenosphere. Any change in the load placed upon the crust causes the asthenosphere to flow—either toward an area where the load has decreased or away from an area with an increased load. In the case of glacial isostasy, adding *glacial* mass depresses the lithosphere into the softer asthenosphere. Conversely, melting the glacier removes the load and the land bounces back, or rebounds—a process called isostatic rebound (see Figure). Greater amounts of rebound in northern Michigan illustrate that the glacier was thicker, and had more mass there, than in the south (Martini et al. 2001). Glacial Lake Algonquin exhibits the best preserved paleolake shoreline in Michigan (Fig. 13.5) and shows remarkable amounts of rebound. At Manistee, its shoreline elevation is 184 m, while at Sault St. Marie it has been uplifted to nearly 305 m. Other paleo-lakes show similar patterns of uplift (Calkin and Feenstra 1985, Fraser et al. 1990).

The asthenosphere does not react quickly to glacial loading and unloading; it lags behind due to its high viscosity. In fact, Michigan is still very slowly "rebounding" from the ice load that has been gone for almost 11,000 years. Southwestern Lake Michigan and southwestern Lake Superior appear to actually be sinking, i.e., shorezones here are very slowly being drowned, because their outlets (at Port Huron and Sault Ste. Marie, respectively) are rising!

When glacial rebound was first being studied, it was thought that there were structurally-controlled "hinge lines," south of which there was no uplift, on the assumption that the underlying bedrock moves like a door on a hinge. Leverett and Taylor (1915), Hough (1958), and Dorr and Eschman (1970) all developed maps showing hinge lines in different places for each major glacial lake (see Figure below). However, the mapped hinge lines do not align with either the underlying bedrock (Plate 6) or any known fault systems. Rebound measurements of shorelines from different times, e.g., Algonquin and Nipissing, show that the rate of rebound is initially fast and decreases over time. Isostasy is a better explanation of these observed responses of the earth's surface rather than the old idea of hinge lines.



Illustrations of concepts of isostatic rebound. A. Figure showing how the mass of a glacier causes isostatic depression, whereas the removal of the glacial mass allows for isostatic rebound. Also shown are preglacial shorelines, initially formed horizontally, but now rising as the glacial load decreases and rebound proceeds. After Martini et al. (2001). B. Map showing four separate hinge lines, once proposed as a mechanism for rebound in southeastern Michigan and southwestern Ontario. After Leverett and Taylor (1915, 505).

The Port Huron glacial advance about 16,000 years ago (Blewett and Winters 1995; Chapters 6, 17) ended the Mackinaw interstade and not only closed the outlet in Ontario but also separated the Saginaw lowlands from southeastern Michigan (Fig. 13.9). The Port Huron moraine cuts off Lake Arkona beaches in southeastern Michigan from Lake Arkona beaches in the Saginaw lowlands, illustrating that they preceded the Port Huron glacial advance. Thus separated from the Saginaw Valley, a large lake with well-developed beach ridges at 225 m called Lake Whittlesey formed in the Lake Erie basin (Leverett and Taylor 1915). Lake Whittlesey drained across the Thumb through a channel near Ubly, in Huron County. The Ubly channel ends at a delta (now cut by the Cass River, 4 km upstream of Caro, in Tuscola County), where it emptied into Lake Saginaw at 212 m, which itself overflowed into the Glacial Grand River and eventually into the Glenwood II phase of Lake Chicago (contributing more sediment to the Allendale delta). Lake Whittlesey shorelines in Michigan appear along the eastern edge of the Defiance moraine from Lenawee County through to Sanilac County.

Retreat from the Port Huron maximum exposed lower outlets and reconnected the proglacial lakes of southeastern Michigan with the Saginaw lowland, ending Lake Whittlesey and forming Lakes Warren I and II at 210–205 m, respectively. These Warren lakes drained through the Glacial Grand River to Lake Chicago; there is debate as to whether Glenwood II or Calumet existed at the time. The Warren lakes were followed by Lake Wayne at 201–199 m, when ice retreat exposed a lower outlet to the east through New York state (Fullerton 1980). Closure of the eastern outlet returned drainage to the Glacial Grand River for Lake Warren III at 206–203 m, based on evidence in north-central New York and incision of the Allendale delta (Muller 1977, Fullerton 1980).

The eastern outlet apparently reopened during the ensuing ice retreat, allowing the lower Lake Grassmere (195 m) to follow Warren III (Calkin and Feenstra 1985). Two other drops in lake level stabilized briefly, forming stages known as Lakes Lundy (189 m) and Elkton (187 m). After this event, around 14,400 years ago, the Lake Erie basin was separated from direct meltwater input and dropped to the low, Early Lake Erie level. Isostatic uplift

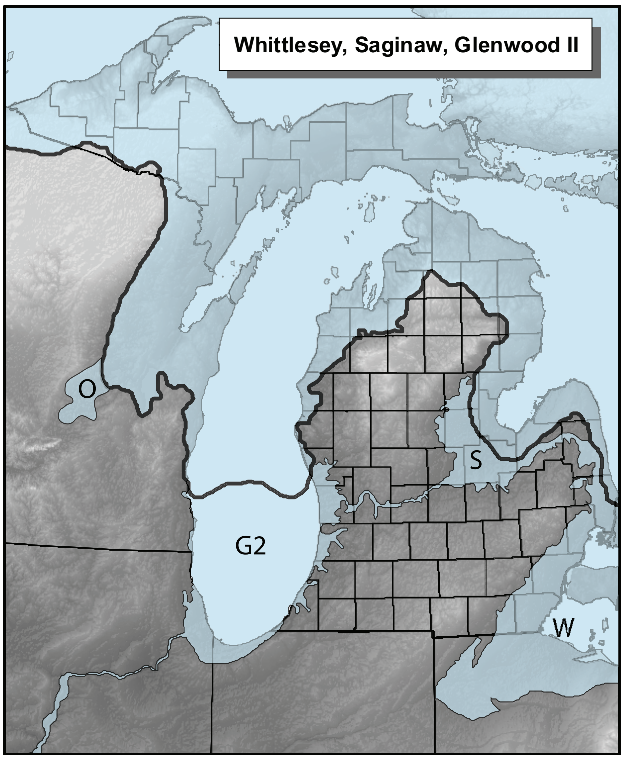


FIGURE 13.9 Map of the ice margin about 15,900 years ago, showing Lake Whittlesey (W), Lake Saginaw (S), the Glenwood II phase of Lake Chicago (G2) and Glacial Lake Oshkosh in Wisconsin.

FIGURE 13.10 Map of the ice margin about 14,000 years ago, showing the low lake levels associated with the Twocreekan interstadial (Two Rivers Phase).

caused the outlet at Niagara River to slowly rise, and as a result the level of Lake Erie also rose over the ensuing 14,000 years, to its present level.

The lake in Lake Huron basin, still in contact with the retreating Port Huron ice margin, probably dropped to the 184 m level of Early Lake Algonquin, most likely draining south via an outlet at Port Huron (Deane 1950). Direct evidence for this lake has not been found, because subsequent lake stages at the same elevation obscured its beaches (Eschman and Karrow 1985). Continued ice retreat past the Straits of Mackinac however, resulted in extension of Early Lake Algonquin into the Lake Michigan basin, eventually leading to the Two Creeks low phase (Fig. 13.10; Hansel et al. 1985). When the retreating ice margin opened a lower outlet near Fenelon Falls, SE of Georgian Bay, Early Lake Algonquin dropped an even lower phase—the Kirkfield Phase. The Kirkfield Phase ended and returned to the Algonquin level when its outlet either rebounded or was covered by a glacial advance. The latter seems more likely, since the timing appears to coincide with the Two Rivers Phase advance (Chapter 6). This glacial readvance also separated the Lakes Michigan and Huron basins again.

Early lakes of the Lake Michigan basin

Whenever the Straits of Mackinac were covered by ice, lake-level fluctuations in the Lake Huron basin did not affect events occurring in the Lake Michigan basin, except for some drainage down the Glacial Grand River. Recall that the Saginaw lobe retreated into south-central Michigan long before the Lake Michigan lobe retreated into the Lake Michigan basin. Meltwater from south-central Michigan, therefore, initially overflowed down the Wabash River, along with overflow from Lake Maumee. Ice retreat to the Valparaiso moraine allowed much of the drainage from south-central Michigan to be redirected to the SW, via the Kankakee River, toward Illinois, and also formed several ice-marginal lakes between the ice margin and the Kalamazoo moraine (Figs. 13.4, 13.6). Only after ice retreated from the Valparaiso position, toward the Lake Border moraine, probably after 17,500 years ago (Farrand and Eschman 1974), did meltwater finally form a lake within the Lake Michigan basin. Known as the Glenwood I phase of Lake Chicago (Fig. 13.7), it stood at 195 m, based on an extensive spit in the Chicago area—about 18 m above current lake level. Lake Chicago drained to the south into the Des Plaines and Illinois Rivers (known as the Chicago outlet), to the Mississippi River.

Prolonged ice recession from the Lake Border moraine during the Mackinaw Interstade eventually formed a lower level Mackinaw phase lake (also referred to as the intra-Glenwood low phase) in the Lake Michigan basin. This lake drained eastward into the Huron basin via an outlet at or near the Straits of Mackinac (Monaghan and Hansel 1990). The elevation of this lake is not known, but Hansel et al. (1985) cited a lack of shore-feature evidence as reason to believe that the Mackinaw Phase water level was lower than present. Monaghan and Hansel (1990) reported a 16,400 year age from wood at the base of a spit near present lake elevation in Berrien County. The rising lake level, which formed the spit, was caused by closure of the Mackinac Straits by the Port Huron ice advance; this marks the end of the Mackinaw Phase. Isolation of the Lake Michigan basin returned drainage to the Chicago outlet, beginning the Glenwood II phase lake level at 195 m (Fig. 13.9). The Glenwood II phase lasted until at least 15,150 years ago (Karrow et al. 1975). Hough (1963) believed that erosion of the Chicago outlet initiated two later lake phases at lower levels. The first was the Calumet phase at 189.0 m; the second was the Toleston at 184.5 m. Both of these phases were initially based on beaches, spits and wave-cuts near the original Glenwood phase spit. Calumet beaches have been traced into southwestern Michigan. The St. Joseph River also built a large delta in Berrien County during the Calumet phase (Kincare 2007). Tracing Toleston phase beaches is difficult, because two subsequent lake phases were at the same elevation. There exists debate as to when the Calumet phase was initiated, with some researchers arguing it was prior to development of the Two Creeks Forest bed found in eastern Wisconsin that has been dated at 13,760 years ago (Bretz 1951, 1959, Eschman and Farrand 1970, Kaiser 1994). Others have argued that it was later (Hough 1958) and possibly related to the amount of discharge entering the lake (Hansel and Mickelson 1988). Hansel et al. (1985) doubted the existence of a Toleston phase entirely.

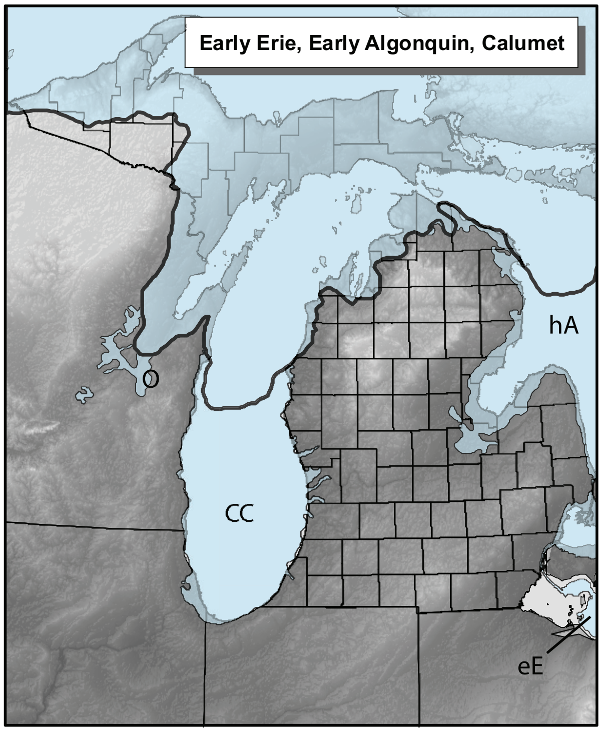


FIGURE 13.11 Map of the ice margin about 13,700 years ago, showing Early Lake Erie (eE), Huron Lake Algonquin (hA), the Calumet phase of Lake Chicago (CC) and Glacial Lake Oshkosh in Wisconsin.

Retreat of the Lake Michigan lobe ice margin from the Port Huron moraine, and the opening of the Straits of Mackinac, eventually allowed Lake Chicago to drop to the level of Early Lake Algonquin in the Huron basin, and then to the level of the Kirkfield phase. During the Kirkfield phase, lake level in the Lake Michigan basin must have been lower than present to allow for the growth of the Two Creeks forest, near present lake level.

A glacial readvance then covered the Straits of Mackinac outlet again, during the Two Rivers Phase, burying the Two Creeks forest bed beneath glacial till (Chapter 6). The advance again isolated the Lake Michigan basin from the Huron basin, causing lake level in the Lake Michigan basin to rise to 189 m, the level of the Calumet phase of Lake Chicago, and returning drainage to the Chicago outlet (Fig. 13.11). The ice advance, during the Two Rivers Phase in Michigan, is assumed to have been at about 13,700 years ago by radiocarbon dates on the Cheboygan bryophyte bed in Cheboygan County (Larson et al. 1994; Chapter 6). This was the last oscillation of ice into the Lower Peninsula of Michigan. Its departure across the Straits of Mackinac ended Lake Chicago, allowing free drainage through the Straits of Mackinac and forming Main Lake Algonquin at about 13,000 years ago (Hansel et al. 1985). However, glaciers would enter the Upper Peninsula at least two more times and continue to influence the levels of the Great Lakes.

Main Lake Algonquin to Lakes Chippewa and Stanley

Main Lake Algonquin, the most extensive of the proglacial Great Lakes, probably drained through both the Chicago and Port Huron (St. Clair River) outlets (Leverett and Taylor 1915, Hough 1958). Hansel et al. (1985) and Larsen (1987), however, contended that outlets uncovered in Georgian Bay, Ontario by glacial retreat, also may have been drainage paths. Regardless, this lake left an indelible mark upon Michigan (Fig. 13.12); its shorelines are marked by well-developed wave-cut bluffs (Schaetzl et al. 2002), beaches (Futyma 1981), and extensive spits (Krist and Schaetzl 2001). As the ice retreated, Lake Algonquin expanded northward, covering much of the eastern half of the isostatically depressed Upper Peninsula, with many islands protruding above the water surface (Schaetzl et al. 2002). The Lake Algonquin shoreline eventually reached at least as far as 65 km north of Sault Ste. Marie (Farrand and Drexler 1985).

Continued ice retreat eventually began uncovering a series of lower outlets east of Georgian Bay—similar to outlets used during the earlier Kirkfield Phase. These outlets were isostatically depressed by the weight of the ice, allowing for drainage across the divide through Ontario. Thus, several post-Main Lake Algonquin phases temporarily stabilized at successively lower elevations as each new outlet was opened. Some of these lower beaches are easily visible on Mackinac Island (Stanley 1945) and many other places near the coast along the northern sections of Lakes Michigan and Huron and in the eastern Upper Peninsula (Schaetzl et al. 2002).

The lowest (by far!) lake level began when the North Bay, Ontario outlet opened around 11,200 years ago, allowing most of the volume of Lake Algonquin to suddenly drain eastward, through Canada (Fig. 13.13). As a result, two very low, small lakes formed in the Lake Michigan and Huron basins. Lake Chippewa (in the Lake Michigan basin) may have been as low as 70 m (Larsen 1987) while Lake Stanley (in the Huron Basin) was even lower—about 45 m (Eschman and Karrow 1985). The level of Lake Chippewa was controlled by a river channel eroded into soft bedrock at the bottom of the Straits of Mackinac, where it flowed eastward into Lake Stanley (Hough 1958). Today, we refer to this valley as the Mackinac Gorge. Only about 800 years separated the Main Algonquin high level from the Chippewa low level. Recently, Lewis et al. (2007) suggested Lake Stanley may have only dropped to 127 m, basing their argument on the elevation of seismic reflections and an erosion surface seen in offshore cores. They also have suggested that, by 9000 years ago, Lakes Chippewa and Stanley were temporarily closed basins, i.e. had no outward drainage, and that an early Holocene dry climate played a part in the inability of lake levels to keep pace with isostatically rising outlets.

FIGURE 13.12 Map of the ice margin about 12,900 years ago, showing Main Lake Algonquin (mA), Early Lake Erie (eE), and Lake Ontonagon (O).

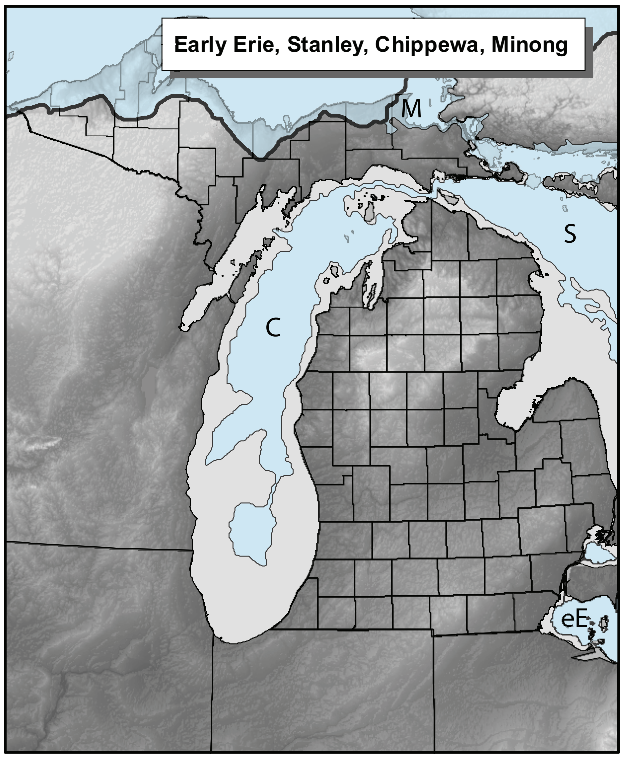


FIGURE 13.13 Map of the ice margin about 11,600 years ago, during the Marquette glacial readvance (Gribben Phase), showing Early Lake Erie (eE), Lake Stanley (S), Lake Chippewa (C), and Lake Minong (M).

Opening the Lake Superior basin

In some ways, the development of Lake Superior is simpler than the other Great Lakes, because the Superior basin was the last to be deglaciated. However, complicating its development are the facts that the basin twice had separate eastern and western lakes, and was subjected to inflows from Lake Agassiz—the largest North American glacial lake, covering a large area in Manitoba and western Ontario, Canada, at the time of the inflows. The Lake Superior basin also experienced the largest amount of isostatic depression (and hence, rebound) of the Great Lake basins, and its outlet at Sault Ste. Marie, being on the eastern edge of the basin, is rebounding more than its western margins.

The oldest existing shorelines in the Superior basin were formed during the retreat of the Two Rivers Phase ice, roughly 12,900 years ago (Fig. 6.3). At this time, Lake Duluth formed at the western end of the basin, as well as the much smaller Lake Ontonagon in Gogebic, Ontonagon, and Houghton Counties (Fig. 13.12). Lake Ontonagon's outlet was at 403 m, draining to the SW through Wisconsin (Leverett 1929). Lake Duluth overflowed to the south, through outlets in Minnesota and Wisconsin, at an elevation of ~331 m. As Lake Duluth was expanding northward along the western shore (and extinguishing Lake Ontonagon), Lake Algonquin was also expanding northward across the eastern Upper Peninsula. Once the retreating ice margin cleared the Keweenaw Peninsula and the Huron Mountains, Lake Duluth merged with Lake Algonquin—which was at a lower level. It is not known how far north Lake Algonquin eventually extended, because the subsequent Marquette phase ice readvance wiped away its northern shorelines as far south as Alona Bay, Ontario (Farrand and Drexler 1985). Following the merger of these two lakes, the level of Lake Algonquin did not remain stable, but slowly fell due to the progressive opening of lower outlets to the east (see above). A bedrock drainage divide at Sault St. Marie, however, prevented the water level in the Superior basin from falling to the level of Lake Stanley in the Huron basin, establishing a new lake in the Superior basin named Lake Minong. It was around this time that a pathway also may have opened for water from Lake Agassiz (Moorhead phase) to escape east into the Superior basin and drain into the North Atlantic (Fisher 2003). Farrand and Drexler (1985) pointed out that Minong shorelines are 40 m above the present outlet at Sault Ste. Marie. A barrier to drainage, perhaps a moraine across Whitefish Bay (Saarnisto 1974), must therefore have existed during Lake Minong, to hold this water up. During the peak of the Marquette Phase ice readvance (Chapter 6), Lake Minong was quite small, pinned into the SE corner of the Superior basin. But as the ice retreated, Lake Minong expanded to the north and west. When the ice cleared the Keweenaw Peninsula, Lake Duluth in the west merged with Lake Minong in the east. Shorelines from Lake Minong are found on Isle Royale, and are the highest shorelines along the north shore of Lake Superior.

During the Marquette Phase ice readvance around 11,580 years ago (Fig. 13.13), the advancing ice margin once again separated the eastern and western portions of the Lake Superior basin, leaving a much diminished Lake Minong in the east and reestablishing Lake Duluth in the west. The advance even briefly squeezed Lake Duluth out of Michigan, leaving a renewed Lake Ontonagon in the western corner of the Upper Peninsula. As the ice margin subsequently retreated, Lake Duluth began expanding northward again, subsuming Lake Ontonagon (Fig. 13.14). When the ice margin pulled back from the Huron Mountains in Marquette County, eastern outlets opened up and again allowed the level of Lake Duluth to fall (Farrand and Drexler 1985). Discharge from Lake Duluth initially went south from Munising via the AuTrain-Whitefish channel, and then found a lower outlet on the north side of the Marquette phase moraines in Marquette and Alger Counties toward Lake Minong (Leverett 1929, Blewett 1994).

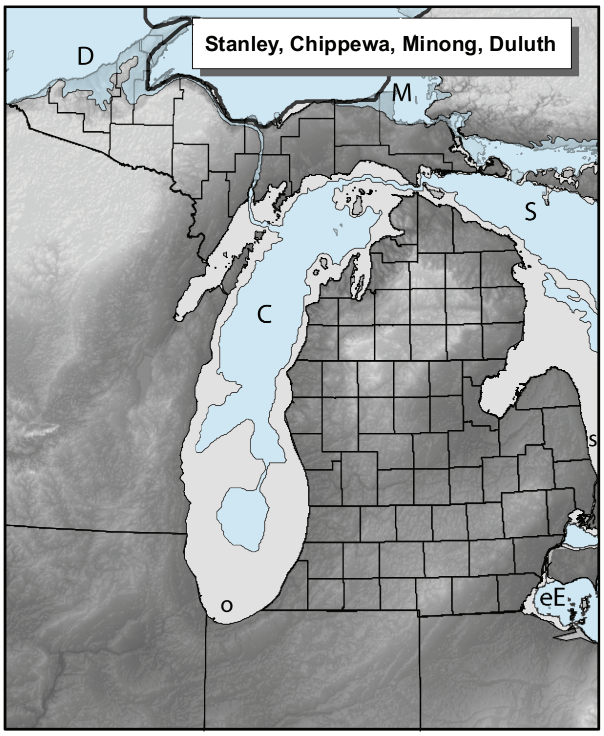


FIGURE 13.14 Map of the ice margin about 11,200 years ago, during retreat of the Marquette ice, showing the rising Lakes Stanley (S) and Chippewa (C), and Lakes Duluth (D) and Minong (M). The locations of the Olson (o) and Sanilac (s) drowned forest beds are also shown.

The Nipissing transgression

As soon as the North Bay, Ontario outlet was uncovered by glacial retreat (about 11,200 years ago), leading to the rapid draining of Lake Algonquin and forming the low-level Lakes Chippewa and Stanley, the outlet started to rebound. The slow but steady rise of the outlet caused the level of these low lakes to rise as well. Without any other influences, the rising lake level should have been a long, uninterrupted asymptotic curve—rapid lake level rises at first, and slowing down over time. However, brief spikes of high lake level interrupted this pattern, because the Marquette Phase ice retreat had opened outlets from Lake Agassiz into the Lake Superior basin, allowing Lake Agassiz floodwaters to discharge into the Superior basin, raising lake levels briefly but markedly each time. The high levels quickly overwhelmed Lake Minong, which was at the time overflowing into the Lake Michigan and Huron basins. At least two floods from Lake Agassiz tore through the Lake Superior basin, the first raising the lake level by as much as 35 m in a very short time period, and in so doing downcutting the Lake Minong outlet and lowering the Minong lake level (Safarudin and Moore 1999). These floods caused a series of temporary rises in levels of Lakes Chippewa and Stanley called the Mattawa highstands (Lewis and Anderson 1989, Lewis et al. 2007), which peaked between 10,600 and 9,300 years ago, before dropping back down to the rising post-Chippewa/Stanley level (Fig. 13.14). Recall that the Chippewa-Stanley lake level was already gradually rising due to the slow isostatic rebound of the North Bay outlet, before flood inflows from the Mattawa highstands occurred. It is likely that the Agassiz floods also eroded the St. Marys River down to bedrock, allowing the rising Chippewa-Stanley Lakes to merge with the a post-Minong lake in the Superior basin.

The results of these floods are observable in the bathymetry of Lake Michigan—between the Garden Peninsula of Delta County and the Door Peninsula of Wisconsin (Fig. 13.15). A distinct channel ending in a delta can be seen

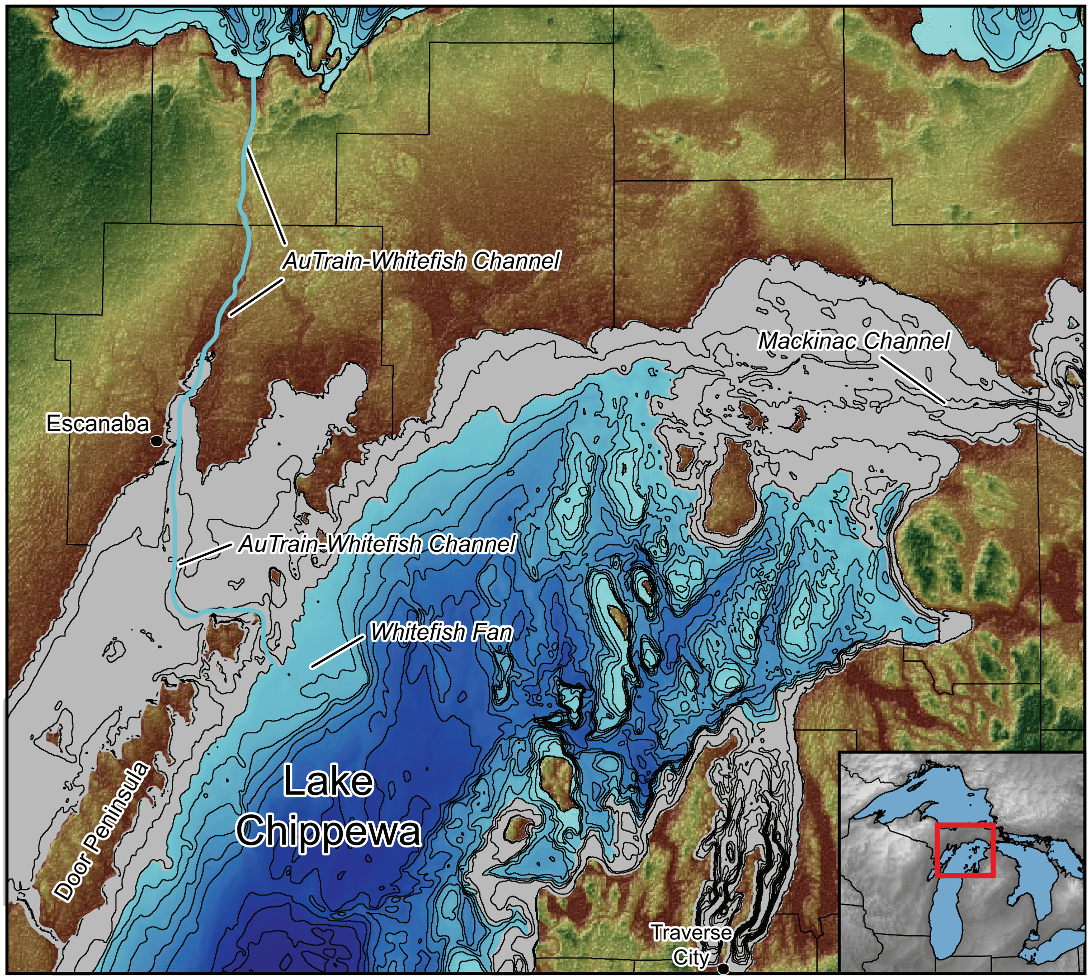


FIGURE 13.15 Digital elevation model (DEM) of the northwest edge of Lake Michigan, showing the Au Train-Whitefish channel descending from onshore to offshore and connecting with the submerged Whitefish fan. DEM data courtesy of the Michigan Center for Geographic Information (www.michigan.gov/cgi).

along the lake bottom. The channel and delta clearly extend directly from the onshore AuTrain-Whitefish Channel, which is a major topographic feature of the Upper Peninsula today. The delta (called the Whitefish fan) is at an elevation of about 126 m and represents the water surface at the time of its formation.

Several drowned forests have been found under the present day upper Great Lakes, attesting to the once low Chippewa-Stanley levels in the Michigan and Huron basins (Fig. 13.14). Wood samples from the Olson drowned forest site, 25 m under Lake Michigan near Chicago, have yielded an average age of 9,155 years (Schneider and Popadic 1994), and samples from the Sanilac drowned forest dated to about 7,350 years ago in 12.5 m water depth (Hunter et al. 2006).

By 6,300 years ago, the North Bay outlet had rebounded to 184 m, roughly the same elevation as the Chicago and Port Huron outlets, allowing all three outlets to be active simultaneously. This point marks the beginning of the Nipissing phase of the upper Great Lake basins (Fig. 13.16). The North Bay outlet continued to rebound, however, and therefore soon all drainage was via the two southern outlets (Chicago and Port Huron) only. Reoccupation of the Port Huron outlet also reestablished drainage from the upper Great Lakes into the lower Great Lakes (Erie and Ontario) after thousands of years of hydrologic separation. Studies by Fraser et al. (1990) indicate that the Nipissing phase high level (as well as the subsequent fall in lake level) was also strongly influenced by climate as well as by rebound.

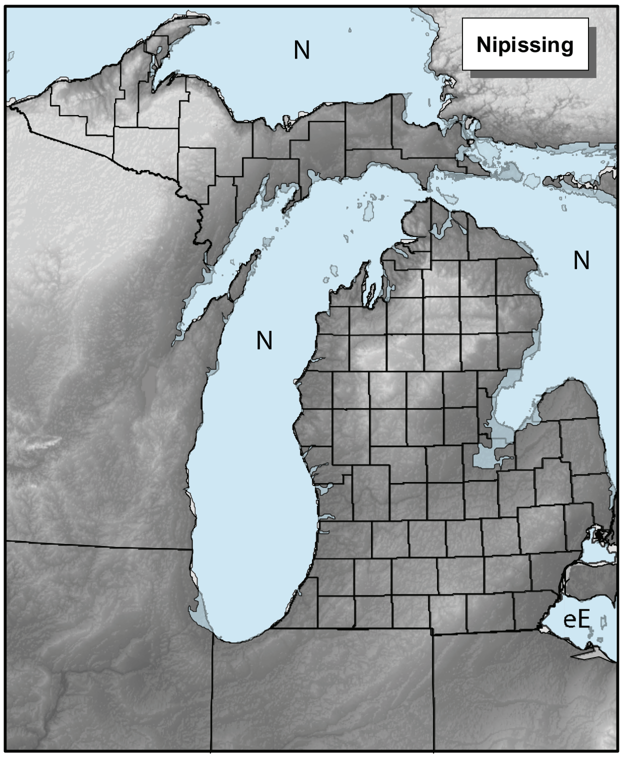


FIGURE 13.16 Map of the Nipissing phase (N) highstand about 6,000 years ago.

Nipissing phase shorelines appear in many places all around the coasts of Lakes Superior, Michigan and Huron, and are often the first prominent shoreline above present lake level (Hough 1958; Fig. 13.5). They are second only to Algonquin shorelines in strength of development and have a more widespread geographic occurrence. Nipissing shorelines appear within a few meters of 184 m as far north as Traverse City and then gradually increase to 197 m at Sault Ste. Marie.

The modern Great Lakes

Because the glacial sediments of the Port Huron outlet were more susceptible to erosion than the bedrock at the Chicago outlet, the Port Huron outlet was gradually lowered by discharge from the Huron basin, allowing it to eventually take all the waters from the upper Great Lakes. The slowly falling Nipissing lake level did, however, stabilize long enough for a poorly-developed beach system to form around 3,400 years ago. Known as Lake Algoma, its elevation was 181.4 m. This event probably represents a temporary shift to a cooler and/or wetter paleoclimate and a subsequent rise in lake level (Fraser et al. 1990). Lake Algoma ended around 2,300 years ago (Baedke and Thompson 2000).

Lastly, rebound of the St. Marys River outlet caused the Lake Superior basin to separate from Lake Huron around 2,280 years ago (Farrand and Drexler 1985). At this time, the modern Great Lakes achieved their present configuration. Studies of beach ridges in embayments around Michigan and Indiana suggest that lake levels have fluctuated on both a 160 year and a 30 year cycle since the end of Lake Algoma (Baedke and Thompson 2000; Chapter 14). Historical records show lake level has varied by about 1.2 m since 1880 A.D. (Fraser et al. 1990).

Literature Cited

Baedke, S.J. and T.A. Thompson. 2000. A 4,700-year record of lake level and isostasy for Lake Michigan. Journal of Great Lakes Research 26:416–426.

Blewett, W.L. 1994. Late Wisconsin history of Pictured Rocks National Lakeshore and vicinity. National Park Service, Pictured Rocks Resource Report 94–01.

Blewett, W.L. and H.A. Winters. 1995. The importance of glaciofluvial features within Michigan's Port Huron moraine. Annals of the Association of American Geographers 85:306–319.

Bretz, J.H. 1951. The stages of Lake Chicago: Their causes and correlations. American Journal of Science 249:401–429.

Bretz, J.H. 1953. Glacial Grand River, Michigan. Papers of the Michigan Academy of Sciences, Arts and Letters 38:359–382.

Bretz, J.H. 1959. The double Calumet stage of Lake Chicago. Journal of Geology 67:675–684.

Calkin, P.E. and B.H. Feenstra. 1985. Evolution of the Erie-basin Great Lakes. In: Karrow P.F. and P.E. Calkin (eds), Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30. pp. 149–170.

Clayton, L. 1997. Pleistocene geology of Dane County, Wisconsin. Wisconsin Geological and Natural History Survey Bulletin 95.

Deane, R.E. 1950. Pleistocene geology of the Lake Simcoe district, Ontario. Geological Survey of Canada Memoir 256.

Dorr, Jr. J.A. and D.F. Eschman. 1970. Geology of Michigan. Univ. of Michigan Press, Ann Arbor.

Eschman, D.F. and W.R. Farrand. 1970. Glacial history of the Glacial Grand valley. In: Guidebook to Field Trips, Geological Society of America, North-Central Section Annual Meeting, East Lansing, MI. pp. 131–157.

Eschman, D.F. and P.F. Karrow. 1985. Huron Basin glacial lakes. In: P.F. Karrow and P.E. Calkin (eds), Quaternary Evolution of the Great Lakes, Geological Association of Canada Special Paper 30. pp. 79–94.

Fairbanks, R.G., Mortlock, R.A., Chiu, T.-C., Cao, L., Kaplan, A., Guilderson, T.P., Fairbanks, T.W. and A.L. Bloom. 2005. Marine radiocarbon calibration curve spanning 0 to 50,000 years B.P. based on paired $^{230}Th/^{234}U$ and ^{14}C dates on pristine corals. Quaternary Science Reviews 24:1781–1796.

Farrand, W.R. and C.D. Drexler. 1985. Late Wisconsinan and Holocene history of the Lake Superior basin. In: P.F. Karrow and P.E. Calkin (eds), Quaternary Evolution of the Great Lakes, Geological Association of Canada Special Paper 30. pp. 17–32.

Farrand, W.R. and D.F. Eschman. 1974. Glaciation of the southern Peninsula of Michigan: A review. Michigan Academician 7:31–56.

Faure, G. 1986. Principles of Isotope Geology. John Wiley and Sons, New York.

Fisher, T.G. 2003. Chronology of glacial Lake Agassiz meltwater routed to the Gulf of Mexico. Quaternary Research 59:271–276.

Flint, R.F. 1971. Glacial and Quaternary Geology. John Wiley and Sons, New York.

Fraser, G.S., Larsen, C.E. and N.C. Hester. 1990. Climatic control of lake levels in the Lake Michigan and Lake Huron basins. In: Schneider A.F. and G.S. Fraser (eds), Late Quaternary History of the Lake Michigan Basin, Geological Society of America Special Paper 251. pp. 75–90.

Fullerton, D.S. 1980. Preliminary correlation of Post-Erie interstadial events (16,000–10,000 radiocarbon years before present), central and eastern Great Lakes region, and Hudson, Champlain and St. Lawrence lowlands, United States and Canada. US Geological Survey Professional Paper 1089.

Futyma, R.P. 1981. The northern limits of glacial Lake Algonquin in Upper Michigan. Quaternary Research 15:291–310.

Hansel, A.K. and D.M. Mickelson. 1988. A reevaluation of the timing and causes of high lake phases in the Lake Michigan Basin. Quaternary Research 29:113–128.

Hansel, A.K., Mickelson, D.M., Schneider, A.F. and C.E. Larsen. 1985. Late Wisconsinan and Holocene history of the Lake Michigan basin. In: P.F. Karrow and P.E. Calkin (eds), Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30. pp. 39–54.

Hough, J.L. 1958. Geology of the Great Lakes. Univ. of Illinois Press, Urbana.

Hough, J.L. 1963. The prehistoric Great Lakes of North America. American Scientist 54:84–109.

Hunter, R.D., Panyushkina, I.P., Leavitt, S.W., Wiedenhoeft, A.C. and J. Zawiskie. 2006. A multiproxy environmental investigation of Holocene wood from a submerged conifer forest in Lake Huron, USA. Quaternary Research 66:67–77.

Karrow, P.F., Anderson, T.W., Clarke, A.H., Delorme, L.D. and M.R. Sreenivasa. 1975. Stratigraphy, paleontology and the age of Lake Algonquin sediments in southwestern Ontario, Canada. Quaternary Research 5:49–87.

Kaiser, K.F. 1994. Two Creeks Interstade dated through dendrochronology and AMS. Quaternary Research 42:288–298.

Kehew, A.E., Nicks, L.P. and W.T. Straw. 1999. Palimpsest tunnel valleys: evidence for relative timing of advances in an interlobate area of the Laurentide Ice Sheet. Annals of Glaciology 28:47–52.

Kincare, K.A. 2007. Response of the St. Joseph River to lake-level changes during the last 12,000 years in the Lake Michigan basin. Journal of Paleolimnology 37:383–394.

Krist, F. and R.J. Schaetzl. 2001. Paleowind (11,000 BP) directions derived from lake spits in northern Michigan. Geomorphology 38:1–18.

Larsen, C.E. 1987. Geological history of Glacial Lake Algonquin and the Upper Great Lakes. US Geological Survey Bulletin 1801.

Larson, G.J., Lowell, T.V. and N.E. Ostrom. 1994. Evidence for the Two Creeks interstade in the Lake Huron basin. Canadian Journal of Earth Sciences 31:793–797.

Leverett, F. 1929. Moraines and shore lines of the Lake Superior region. US Geological Survey Professional Paper 154–A.

Leverett, F. and F.B. Taylor. 1915. Pleistocene of Michigan and Indiana and the History of the Great Lakes. United States Geological Survey Monograph 53.

Lewis, C.F.M. and T.W. Anderson. 1989. Oscillations of levels and cool phases of the Laurentian Great Lakes caused by inflows from glacial Lakes Agassiz and Barlow-Ojibway. Journal of Paleolimnology 33:445–461.

Lewis, C.F.M., Heil, C.W., Hubeny, J.B., King, J.W., Moore, T.C. and D.K. Rea. 2007. The Stanley unconformity in Lake Huron basin: evidence for a climate-driven closed lowstand about 7900 14C BP, with similar implications for the Chippewa lowstand in Lake Michigan basin. Journal of Paleolimnology 37:435–452.

Martini, I.P., Brookfield, M.E. and S. Sadura. 2001. Principles of Glacial Geomorphology and Geology. Prentice Hall, Upper Saddle River, NJ.

Mickelson, D.M., Clayton, L., Fullerton, D.S. and H.W. Borns Jr. 1983. Late glacial record of the Laurentide Ice Sheet in the United States, In: Late Quaternary Environments of the United States, Wright, H.E., Jr. (ed), Volume 1: The Late Pleistocene (S.C. Porter, editor), University of Minnesota Press, Minneapolis. pp. 3–37.

Monaghan, G.W. and A.K Hansel. 1990. Evidence for the intra-Glenwood (Mackinaw) low-water phase of glacial Lake Chicago. Canadian Journal of Earth Sciences 27:1236–1241.

Muller, E.H. 1977. Late glacial and early postglacial environments in western New York. Annals of the New York Academy of Sciences 288:223–233.

Rieck, R.L. and H.A. Winters. 1993. Drift volume in the southern peninsula of Michigan—a prodigious Pleistocene endowment. Physical Geography 14:478–493.

Saarnisto, M. 1974. The deglaciation history of the Lake Superior region and its climatic implications. Quaternary Research 4:316–339.

Safarudin and T.C. Moore. 1999. The history and architecture of lacustrine depositional systems in the northern Lake Michigan Basin. Journal of Paleolimnology 22:475–496.

Schaetzl, R.J., Drzyzga, S.A., Weisenborn, B.N., Kincare, K.A., Lepczyk, X.C., Shein, K.A., Dowd, C.M. and J. Linker. 2002. Measurement, correlation, and mapping of Glacial Lake Algonquin shorelines in northern Michigan. Annals of the Association of American Geographers 92:399–415.

Schneider, A.F. and T.H. Popadic. 1994. The Nipissing transgression in the Lake Michigan Basin: Summary and speculation. Geological Society of America, North Central Section 28th Annual Meeting Abstract 26(4):27.

Soller, D.R. 1998. Map showing the thickness and character of Quaternary sediments in the glaciated United States east of the Rocky Mountains; northern Great Lakes states and central Mississippi Valley states, the Great Lakes, and southern Ontario (80 degrees 31' to 93 degrees west longitude). US Geological Survey IMAP 1970–B.

Stanley, G.M. 1936. Lower Algonquin beaches of Penetanguishene Peninsula. Geological Society of America Bulletin 47:1933–1960.

Stanley, G.M. 1945. Prehistoric Mackinac Island. Michigan Geological Survey Publication 43.

Stuiver, M. 1978. Atmospheric carbon dioxide and carbon reservoir changes. Science 199:253–258.

Further Readings

Alley, R.B. 2000. The two-mile time machine: ice cores, abrupt climate change, and our future. Princeton Univ. Press, Princeton, NJ.

Schneider, A.F. and G.S. Fraser (eds). 1990. Late Quaternary history of the Lake Michigan Basin. Geological Society of America Special Paper 251.

Teller, J.T. 1987. Proglacial lakes and the southern margin of the Laurentide ice sheet. In: Ruddiman, W.F. and H.E. Wright Jr. North America and adjacent oceans during the last deglaciation. Geological Society of America Geology of North America K-3:39–69.