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REVIEW

Origin and Evolution of the Great Lakes

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ABSTRACT. *This paper presents a synthesis of traditional and recently published work regarding the origin and evolution of the Great Lakes. It differs from previously published reviews by focusing on three topics critical to the development of the Great Lakes: the glaciation of the Great Lakes watershed during the late Cenozoic, the evolution of the Great Lakes since the last glacial maximum, and the record of lake levels and coastal erosion in modern times.*

The Great Lakes are a product of glacial scour and were partially or totally covered by glacier ice at least six times since 0.78 Ma. During retreat of the last ice sheet large proglacial lakes developed in the Great Lakes watershed. Their levels and areas varied considerably as the oscillating ice margin opened and closed outlets at differing elevations and locations; they were also significantly affected by channel downcutting, crustal rebound, and catastrophic inflows from other large glacial lakes.

Today, lake level changes of about a 1/3 m annually, and up to 2 m over 10 to 20 year time periods, are mainly climatically-driven. Various engineering works provide small control on lake levels for some but not all the Great Lakes. Although not as pronounced as former changes, these subtle variations in lake level have had a significant effect on shoreline erosion, which is often a major concern of coastal residents.

INDEX WORDS: *Glacial geology, geologic history, proglacial lakes, Great Lakes basin, shoreline erosion, lake levels.*

INTRODUCTION

The North American Great Lakes watershed (Fig. 1) covers about 765,990 km^2 and is home to one-tenth of the population of the United States and one-quarter the population of Canada. It includes part or all of eight U.S. states and a Canadian province, and contains the five Great Lakes which collectively represent the largest unfrozen freshwater body on Earth.

The origins of the watershed are a product of multiple glaciations during the late Cenozoic as well as redirected drainage, particularly during retreat of the last ice sheet. As a result, its history and evolution have long attracted the attention of glacial and Quaternary geologists and have resulted in a number of excellent reviews during the last 85 years (Leverett and Taylor 1915; Hough 1958, 1963, 1966; Fullerton 1980; Mickelson *et al.*1982; Dyke and Prest 1987; Karrow 1989). In addition, there have been several syntheses that have focused on parts of the Great Lakes watershed (Chapman and Putnam 1984, Dreimanis and Karrow 1972, Farrand and Eschman 1974, Barnett 1992, Dreimanis 1977) and some of the best reviews specific to development of glacial and post glacial lakes can be found in Farrand 1969, Karrow *et al.* 1975, Karrow and Calkin 1985, Larsen 1987, Teller 1987, Hansel and Mickelson 1988, Schneider and Fraser 1990, Colman *et al*. 1994a, and Lewis *et al.* 1994.

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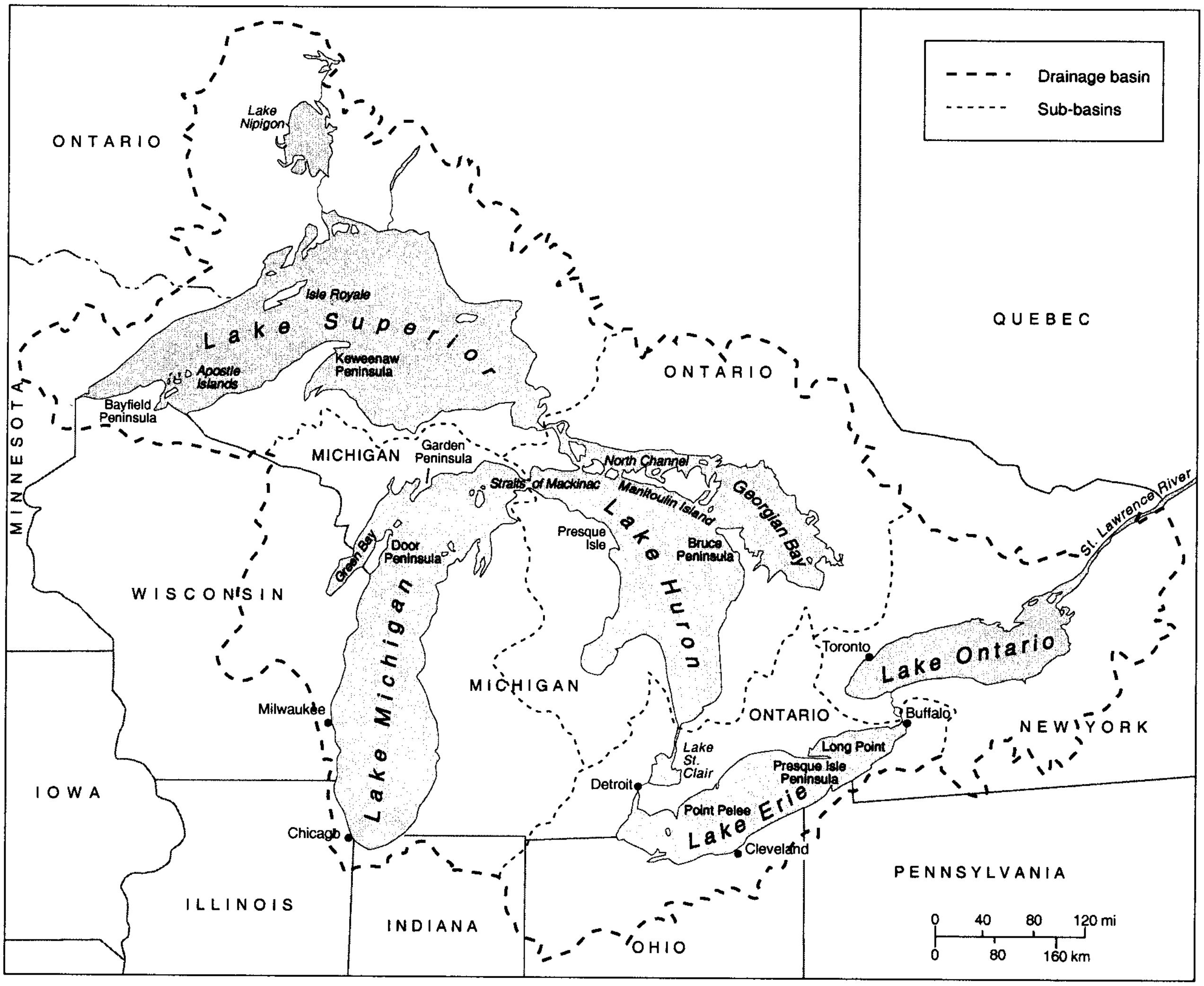

FIG. 1. The Great Lakes watershed. The watersheds of each particular lake are indicated by thin dashed lines. Modified from Botts and Krushelnicki (1988).

The following review differs from those previously published by focusing on several topics critical to the history of the Great Lakes watershed, mainly the glaciation of the watershed during the late Cenozoic, the evolution of the Great Lakes since the last glacial maximum, and the record of lake levels and shore erosion over the last several decades. It also includes considerable new information not available to earlier reviewers. Collectively, these topics provide a unique insight into the watershed's history; the potential impact of future events such as climate change or drainage diversion can only be assessed in light of that history.

PHYSICAL SETTING

The Great Lakes watershed (Fig. 1) can be divided into a southern, lowland region underlain by relatively gently-dipping sedimentary rocks of Paleozoic age, and a northern, upland region (Canadian shield) underlain by granite, gneiss, and metavolcanic and metasedimentary rocks of Precambrian age (Fenneman 1938, Hough 1958, Dickas 1986). In general, the lowland region includes the Erie and Michigan basins[1] and most of the Huron and On-

[1]Throughout this paper, the term "Michigan basin" is used to refer to the basin in which the waters of Lake Michigan reside and should not be confused with the geologic structural feature also known as the Michigan Basin.

tario basins. Except for broad, low morainal ridges and a few bedrock escarpments, it tends to be an area of low relief, and is generally blanketed by a continuous mantle of glacial sediments, often greater than 50 m in thickness and in places over 350 m thick (Rieck and Winters 1993, Soller 1998). The upland region includes most of the Superior and Georgian Bay basins and parts of the Ontario basin. It can be distinguished topographically by a distinct bedrock-dominated topography formed as a result of bedrock structure and differential erosion by glaciers. Thin, discontinuous glacial sediments blanket this region (Fenneman 1938, LaBerge 1994).

Preglacial Landscape

Prior to Quaternary glaciations, the Great Lakes watershed was subjected to long-term subaerial erosion. Evidence for this, however, is sparse and includes fragments of former bedrock valley systems developed on the preglacial bedrock landscape (Fig. 2). Of these, the best known is the Laurentian drainage system, which can be traced in the subsurface from the western end of the Ontario basin to the southern end of the Georgian Bay basin (Spencer 1891, White and Karrow 1971, Eyles *et al.* 1985). It formerly extended eastward through the Ontario basin to the head of the St. Lawrence valley and included tributaries that reached as far west as the Erie, Huron, Superior, and Michigan basins (Spencer 1907, Horberg and Anderson 1956). The Teays-Mahomet valley system occurs south of the watershed and was tributary to the ancient Mississippi drainage system (Tight 1903, Horberg 1950, Gray 1991, Teller and Goldthwait 1991). Both the Laurentian and Teays-Mahomet systems, evolved throughout the late Tertiary and Quaternary, were undoubtedly modified by repeated glaciations. It is therefore doubtful that the integrated drainage networks depicted in Figure 2 existed at any one time.

Evidence for preglacial weathering processes and soil formation occurs in the form of saprolite (highly weathered rock, in place) that is found in Quebec (LaSalle and De Kimpe 1989, Bouchard and Pavich 1989) and in the Adirondack Mountains of New York (Muller 1965), though some of these may also have formed during an interglacial period. Neither saprolite nor residuum have been reported in the Great Lakes watershed proper, probably because of removal by Quaternary glaciers or because it now lies deeply buried beneath glacial sediments.

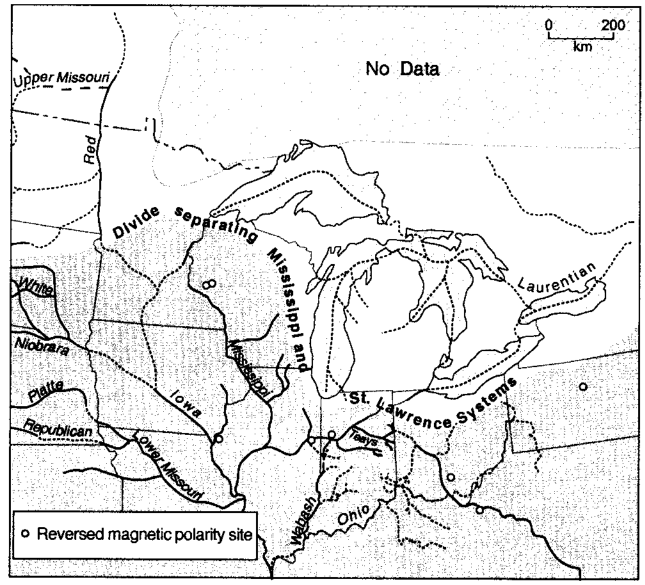

FIG. 2. Preglacial drainage systems inferred for the Great Lakes region. Sites where magnetically-reversed sediments (> 0.78 Ma) have been documented are shown by an open circle. In part, after Spencer (1907), Horberg and Anderson (1956), and Gray (1991).

Finally, deep weathering of limestones within the watershed led to the development of karst—sinkholes, underground streams, and caves. These features, common on the Devonian limestone (Fig. 3), were undoubtedly partially eroded by Quaternary glaciations, and then covered by (filled with) glacial sediments (Farrand 1995). A particularly interesting solution feature developed on Silurian limestone is the Pipe Creek Jr. sinkhole located just south of the Great Lakes watershed in Grant County, Indiana (Farlow *et al.* 1997, 1998, 2001; Holman 1998). It is buried by drift and is filled with locally derived rubble containing a diverse assemblage of late Miocene to early Pliocene (Late Hemphillian) fossil vertebrates and plants.

Origin of the Lake Basins

The Ontario, Erie, Huron, Superior, and Michigan basins owe their origin mainly to channeling of ice flow along major bedrock valley systems that existed prior to glaciation, and to increased glacial scouring and erosion in areas of relatively weak bedrock (Fenneman 1938, Hough 1958, Cvancara and Melik 1961, Wold *et al.* 1981). This is particu-

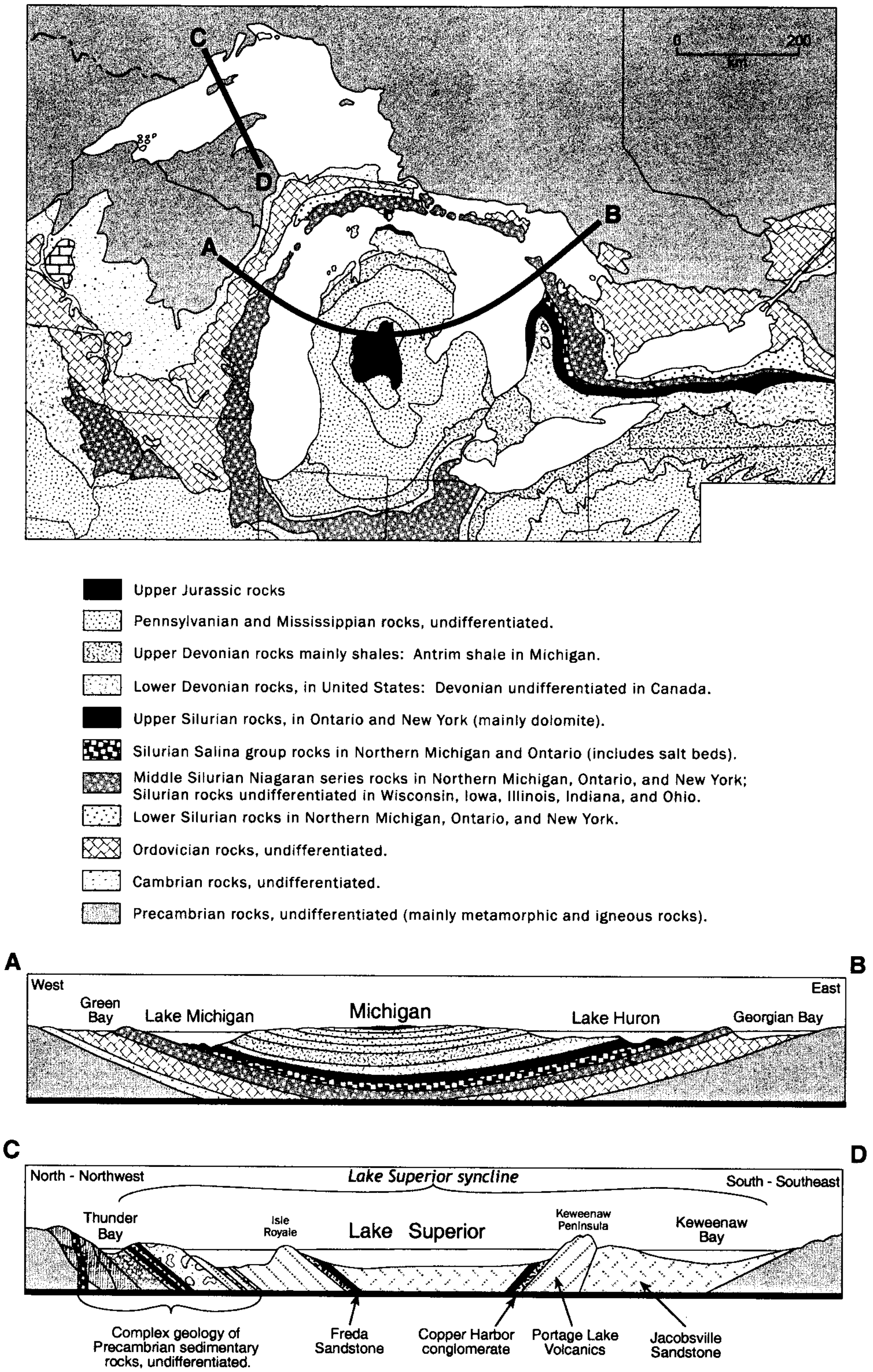

FIG. 3. Bedrock geology of the Great Lakes watershed. All bedrock shown in the C-D transect is Precambrian in age; different stippling and shading patterns are shown to differentiate one rock unit from another, and do not refer back to the key shown above. Modified from Hough (1958) and Westjohn and Weaver (1998).

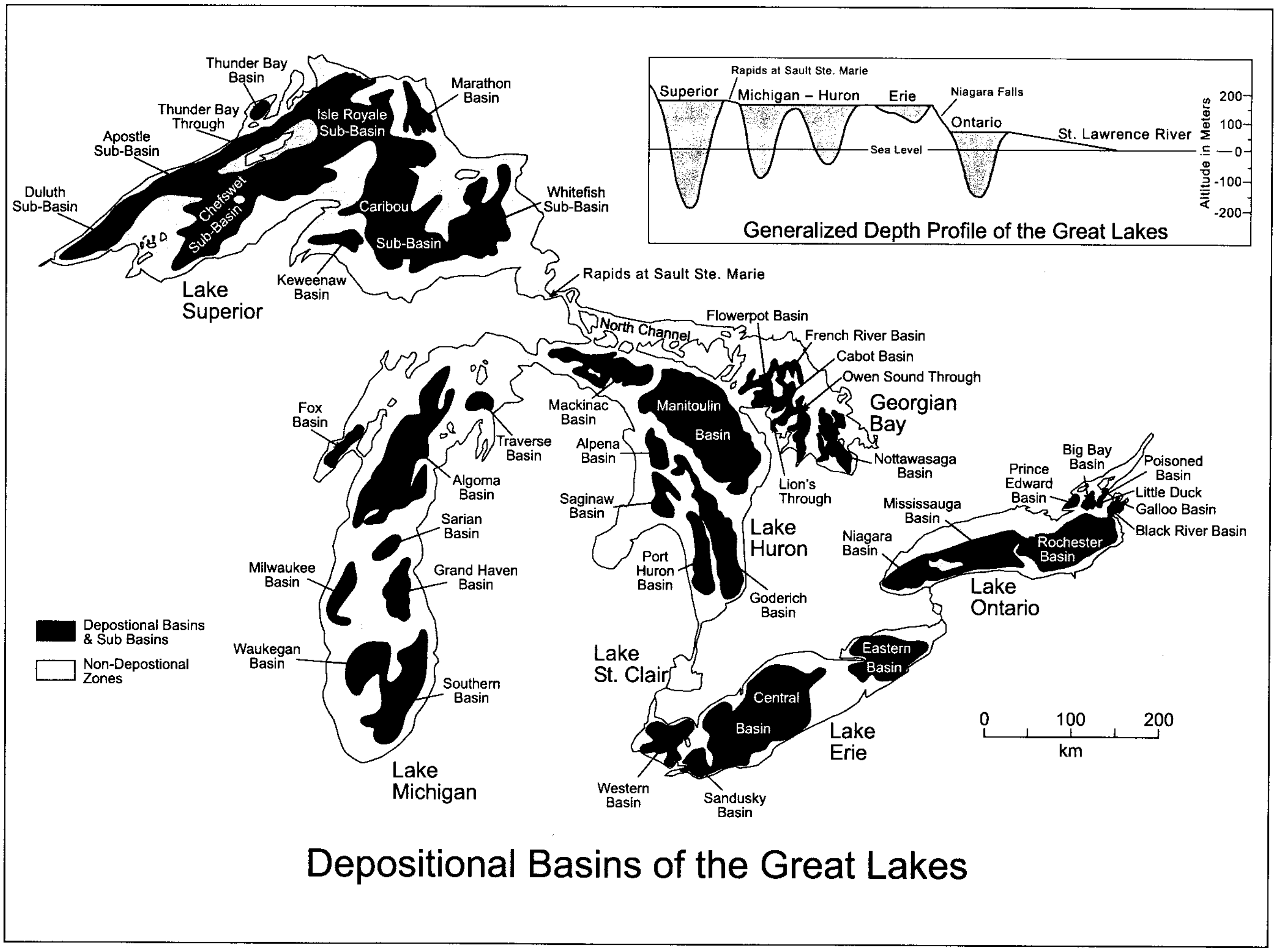

FIG. 4. Depositional basins of the Great Lakes, modified from Cahill (1981). Depth profile of the Great Lakes, modified from Hough (1958).

larly evident from Figure 3, which shows parts of the Huron, Erie, and Michigan basins conforming to the outcrop pattern of Devonian and Upper Silurian rocks that are, in large part, erodible shales and limestones. Likewise, a belt of weak Ordovician shales underlies Green Bay (on the west side of the Michigan basin), North Channel and Georgian Bay (on the northern and northeastern part of the Huron basin), and much of the southern half of the Ontario basin. Even the Superior basin, which lies almost wholly within the Canadian Shield, is largely developed along the length of a structural basin that includes sandstones of Precambrian age and Upper Keweenawan (late Precambrian) sedimentary rocks that are slightly metamorphosed and considerably less resistant to glacial erosion than underlying, older volcanic rocks (Hough 1958, Dickas 1986).

It is likely that the ancestral Great Lakes were shallower water bodies than the current Great Lakes, and have been considerably deepened by glacial scour (Rieck and Winters 1982). The amount of scouring and overdeepening in each basin varied considerably. For example, the Superior basin, the deepest of the five, has a floor that lies approximately 213 m below sea level or over 397 m below the basin's rocky outlet at Sault Ste. Marie (Fig. 4). In contrast, the Erie basin is the shallowest and has a floor that lies approximately 110 m above sea level or 64 m below its outlet near Niagara Falls. Lake St. Clair, which lies between Lakes Huron and Erie, is no deeper than 6.4 m, and in most places is shallower than 3 m. The floors of the Superior, Huron, and northern part of the Michigan basins also tend to be irregular and complicated by the presence of resistant bedrock layers. In contrast, the floors of the Erie, Ontario, and southern part of the Michigan basins tend to be relatively smooth due to the "softness" and relative homogeneity of the underlying bedrock.

In many places, Quaternary sediments underlie the floor of the Great Lakes. In the Michigan, Huron, Erie, and Ontario basins their thickness can exceed 100 m and in the Superior basin they can exceed 250 m (Soller 1993, 1998). These sediments, particularly within the Erie basin, indicate that lake floor topography is not just the product of glacial erosion but also of glacial and post glacial deposition. Each basin can also be subdivided in depositional basins and subbasins (Fig. 4) that are currently receiving the bulk of fine-grained sediment washing into the Great Lakes (Cahill 1981). Their number, depth, size, and shape, however, vary from basin to basin. For example, the Erie basin includes four depositional basins, most of which are shallow, broad, and semicircular. In contrast, the Huron basin contains 10, most of which are deep, elongate, or irregularly shaped.

Islands and Peninsulas

Most of the Great Lakes islands and major peninsulas are underlain by resistant bedrock that has withstood erosion from multiple glaciations. For example, Silurian dolomite forms the Door and Garden peninsulas, and islands that separate Green Bay from Lake Michigan (Figs. 1, 3). The same bedrock (Niagaran series) also occurs along the northern shore of Lake Michigan and forms the islands that separate North Channel and Georgian Bay from Lake Huron. Additionally, it forms Lake Huron's Bruce Peninsula (Fig. 1) and can be traced southeastward along a broad arch to where it forms a prominent escarpment over which the Niagara River flows at Niagara Falls. Most of the islands in Lake Michigan just west of the Straits of Mackinac are also associated with resistant dolomite and limestone of the Bois Blanc Formation (Devonian). The archipelago known as the Less Cheneaux Islands, east of the Straits, are knobs of resistant Engadine dolomite (Silurian). The small islands at the western end of Lake Erie are likewise associated with resistant layers of the Columbus Limestone and Upper Bass Island Dolomite (Devonian). In the Superior basin, resistant bedrock associated with the multiple strata of Portage Lake volcanics (Keweenawan) form the backbone of the Keweenaw Peninsula as well as much of Isle Royale (Figs. 1, 3). In the western part of the basin, resistant sandstone of the Bayfield Group (Precambrian) forms the Apostle Islands and underlies the Bayfield Peninsula.

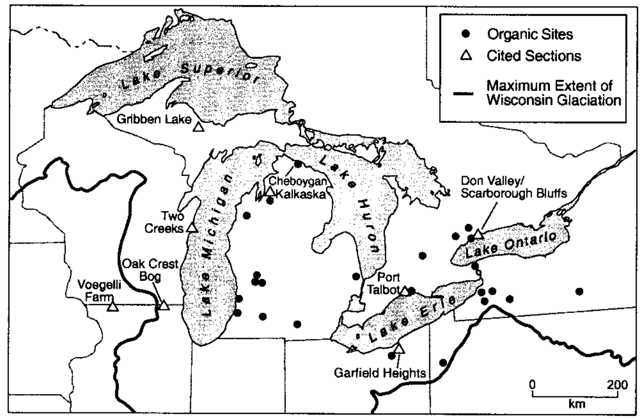

FIG. 5. Locations of organic deposits in glacial sediments within the Great Lakes watershed, dated between 64.5 and 25 ka. Also shown are the locations of major stratigraphic sections, and the maximum southern extent of the Wisconsin glaciation. Note that Voegelli Farm and Oak Crest Bog sections lie outside the Great Lakes watershed.

GLACIAL HISTORY

The record of the most recent glaciation is well preserved in the Great Lakes watershed and includes a number of stratigraphic sections that have been studied in great detail; the most commonly cited sections are shown in Figure 5. The stratigraphic record is incomplete, however, for glaciations that preceded the last, mainly because much of the sedimentary record has either been completely eroded away by subsequent glaciations, or is buried too deeply to be easily studied (Blewett 1991, Rieck and Winters 1993). South and west of the Great Lakes watershed, however, where glacial erosion was less severe, the record of earlier glaciations is better preserved and provides some insight to the watershed's early glacial history. Oceanic records, particularly oxygen isotope ratios (^{18}O) from deep sea sediments, also provide clues as to when global ice volumes may have been large enough to allow glacier ice to invade the watershed (Ruddiman and Raymo 1988).

Evidence of Early Glaciations

Glaciolacustrine silts having reversed magnetic polarity have been reported in Pennsylvania (Gardner *et al.* 1994), West Virginia (Bonnett *et al.* 1991), Ohio (Hoyer 1976, 1983), Indiana (Bleuer 1976), Illinois (Johnson 1986) and Wisconsin (Baker *et al.* 1983) (Fig. 2). Their presence has been used to argue that glacial ice had penetrated into at least part of the Great Lakes watershed before the Matuyama-Brunhes geomagnetic reversal[2] of approximately 0.78 Ma[3] (Fullerton 1986, Johnson 1986). The magnetically reversed silts in Wisconsin, however, are particularly noteworthy because they may be the same age as magnetically reversed glacial sediments in Iowa and Nebraska dated at > 2.01 Ma (Richmond and Fullerton 1986). If so, it would suggest that glacier ice extended across much of the northern Great Plains and possibly into the western end of the Great Lakes watershed during the Matuyama-Reunion reversal which occurred around 2.14 Ma (Richmond and Fullerton 1986). On the other hand, it appears that most, if not all, of the existing loess-stratigraphic (windblown silt) units in the Mississippi Valley are younger than 0.79 Ma (Pry and Johnson 1988, Clark *et al.* 1989, Forman *et al.* 1992, Leigh and Knox 1993) and the lack of older loess-stratigraphic units there suggests that glacier ice rarely advanced into the upper Mississippi watershed (Iowa and Minnesota) prior to 0.79 Ma (Leigh and Knox 1993).

The oxygen isotope marine record likewise shows that initiation of moderate-sized ice sheets in the Northern Hemisphere began no earlier than 2.4 Ma and that large-scale ice sheets, comparable in size to the ones that covered much of North America and northern Europe during the last glaciation, did not develop until about 0.8–0.7 Ma (Ruddiman and Raymo 1988). The marine record also indicates that large ice sheets in the northern hemisphere have waxed and waned on a ~100,000 year cycle since about 0.74 Ma (Ruddiman and Raymo 1988, Shackleton *et al.* 1988).

Evidence for Multiple Glaciations

Many sites in Ohio, Indiana, and Illinois show stratigraphic evidence for multiple glaciations. When viewed collectively they indicate that the Great Lakes watershed must have been glaciated partially or totally at least six times since 0.78 Ma (Fullerton 1986, Johnson 1986) and that some glaciations must have extended as far south as northern Kentucky (Leighton and Ray 1965, Ray 1974, Swadley 1980).

The last two glaciations extended to the Ohio River near Cincinnati and are referred to as the Illinoian and Wisconsin glaciations, which occurred between 0.302 and 0.132 Ma, and between 0.79 Ma and 10 ka, respectively (Richmond and Fullerton 1986). Separating the two is the Sangamon interglaciation which, from Ohio westward to Iowa and Kansas, is often represented by an extensive and well developed soil (paleosol, geosol) believed to have formed in a climate warmer and perhaps drier than present (Follmer 1978, Schaetzl 1986, Curry and Pavich 1996). Glacial deposits older than those deposited by the Illinoian ice (referred to as pre-Illinoian) also occur in these states but, because their correlation with respect to the Pearlette ash was thrown into doubt about 30 years ago, their absolute age has yet to be worked out (Hallberg 1986).

The best evidence for multiple glaciations in the Great Lakes watershed occurs in the Don Valley Brickyard near Toronto (Fig. 5) where a fossiliferous sand (Don Formation) rests on till and is overlain by a thick sequence of sediments associated with the Wisconsin glaciation. The fossils found in the sand include pelecypods and gastropods (Coleman 1933, Baker 1931, Kerr-Lawson *et al.* 1992), pollen and plant remains (Terasmae 1960, Richard *et al.* 1999), diatoms (Duthie and Mannada Rani 1967), caddisflies (Williams and Morgan 1977), ostracods (Poplawski and Karrow 1981), and vertebrates (Karrow 1969, Harington 1990). Studies of the diatoms, caddisflies, and ostracods all indicate that the sand was deposited in a freshwater estuary or river mouth environment and in a climate typical of temperate North America today (Duthie and Mannada Rani 1967, Williams and Morgan 1977, Poplawski and Karrow 1981). Studies of the pollen and plant remains indicate the presence of a hardwood forest which was yielding to spruce/pine (Terasmae 1960, Richard *et al.* 1999). On the basis of the pollen record it has been suggested that the annual mean temperature at the time the sand was deposited was probably 3°C warmer than present (Terasmae 1960). The general consensus is that sand at the brickyard was probably deposited during the Sangamon interglaciation, whereas the underlying till was most likely deposited during the

[2]A geomagnetic reversal is when the earth's magnetic fields periodically reverses polarity.

[3]Throughout this paper "Ma" refers to million years and "ka" refers to thousand of years. Ages < 65 ka are in radiocarbon years.

Illinoian glaciation (Terasmae 1960, Karrow 1984, Karrow 1990).

Evidence for multiple glaciations also occurs at Garfield Heights, near Cleveland (Fig. 5), where a well developed paleosol formed on outwash gravel is overlain by a series of tills. The paleosol has been associated with the Sangamon interglaciation (White 1953, 1968), but it may have just as likely formed before the Illinoian glaciation (Fullerton 1986).

Other sites in and adjacent to the Great Lakes watershed that contain paleosols or organic beds that probably represent interglaciations (Fullerton 1986) occur east of Lake Ontario on the Ontario-Quebec border (Anderson *et al.* 1990), in northwestern Pennsylvania (White 1969), in western New York (Calkin *et al.* 1982), and in Illinois, Indiana, and Iowa (Ruhe 1956, Follmer 1982, Olson 1989, Johnson and Balek 1991). Of particular interest is also a pollen record from south-central Illinois. It shows that that region was characterized by deciduous forest with bald cypress during the Sangamon interglaciation (Grüger 1972a, Grüger 1972b, Teed 2000).

Advance of the Last Ice Sheet

During the Wisconsin glaciation the Laurentide Ice Sheet, centered in northern and eastern Canada, expanded southward and eventually covered the entire Great Lakes watershed. While at its maximum, it probably had two centers from which ice flow radiated, one located near Labrador and the other immediately west of Hudson Bay; the thickness of ice at the centers was probably between 2,500 and 3,000 m (Boulton *et al.* 1985). Over the Great Lakes watershed, however, it was considerably thinner. Recent estimates based on glaciological theory place it between 750 and 2,500 m (Hughes *et al.* 1981, Boulton *et al.* 1985). The reason for such a wide range in estimated thickness is that the calculation of the lower value takes into account basal shear-stress conditions that occur when impermeable and deformable sediments, such as shale, clayey till, or fine lacustrine sediments, underlie that glacier bed (Boulton *et al.* 1985), whereas calculation of the higher value takes into account basal shear-stress conditions that occur when only permeable and nondeformable bedrock (or sediments) underlies the glacier bed (Hughes *et al.* 1981). The presence of highly impermeable and/or deformable beds beneath the ice would also mean ice moved more rapidly and had a gentler surface slope than ice that flowed across mainly permeable and/or nondeformable beds (Boulton *et al.* 1985, Clayton *et al.* 1985, Clark 1992).

Stratigraphic evidence from Scarborough Bluffs near Toronto (Fig. 5) indicates that during the early part of the Wisconsin glaciation, between 65 to 79 ka, ice advanced from the northeast and dammed a lake in the Ontario basin (Karrow 1984). How far west the ice advanced, however, is debatable (Fullerton 1986, Karrow 1989, Eyles and Westgate 1987, Dreimanis 1992, Hicock and Dreimanis 1992, Miller *et al.* 1992, Szabo 1992), but it probably did not extend as far as western Indiana. In the Michigan and Superior basins no reliable stratigraphic evidence exists to suggest the presence of glacier ice there during the early part of the Wisconsin glaciation. However, it is likely that large lakes existed in those basins at this time.

At Scarborough Bluffs (Fig. 5), stratigraphic evidence also shows that during the middle part of the Wisconsin glaciation, between 35 and 65 ka, the margin of the ice sheet oscillated within the Ontario basin (Karrow 1969, 1984, 1989). Exactly where it oscillated has been debated (Karrow 1984, 1989; Eyles and Westgate 1987; Eyles and Williams 1992), but it is clear that it terminated in a proglacial lake. Fine-grained sediments in a high-level, ice-dammed lake in the Finger Lakes region of New York contain wood that has yielded a radiocarbon age of 41.9 ka. This finding also supports the presence of glacier ice in the Ontario basin during the middle part of the Wisconsin glaciation (Bloom and McAndrew 1972).

In contrast to the eastern end of the Great Lakes watershed, there is no firm stratigraphic evidence elsewhere in the watershed to suggest the presence of glacier ice during the middle part of the Wisconsin glaciation (Winters *et al.* 1986). In fact, radiocarbon ages from buried organic deposits found in the Erie, Huron, and Michigan basins (Fig. 5) show that much of the southern part of the watershed was generally free of glacial ice during the interval from 64.5 to 25 ka (Winters *et al.* 1986; Curry and Follmer 1992; Karrow 1984, 1989). The association of fine-grained sediments with some of the organic deposits also suggests that lakes were present and that their levels fluctuated—probably as the result of glacier ice blocking drainage to the east (Dreimanis *et al.* 1966, Winters *et al.* 1986). At times, some of the lake drainage may also have been directed southward into the Illinois River, carrying with it sediment that may have been the source for the Roxana loess along the lower Illinois

River valley (Winters *et al.* 1988). An alternative explanation for the loess, however, is that it was derived from flood plains of proglacial rivers associated with an ice advance into the upper Mississippi River valley (Minnesota and western Wisconsin) sometime between 55 ka and 27 ka (Johnson and Follmer 1989; Leigh and Knox 1993, 1994; Leigh 1994). If true, it would suggest that glacier ice probably also extended into Superior basin and perhaps even into the northern end of the Michigan basin during the middle part of the Wisconsin glaciation (Grimley 2000).

Some buried organic deposits have yielded pollen that sheds light on the vegetation and climate that existed in unglaciated parts of the Great Lakes watershed during the middle part of the Wisconsin glaciation. For example, an extensive pollen record obtained from Port Talbot along the north shore of Lake Erie (Figure 5) shows initial warm and dry climatic conditions (but cooler than an interglacial) followed by climatic cooling and possibly a forest-tundra environment (Berti 1975). Another pollen record from near Kalkaska in northern lower Michigan (Fig. 5) shows that about 35 ka the vegetation there evolved from a cold, open forest into a closed boreal forest not unlike that of the northern Great Lakes today (Winters *et al.* 1986). In contrast, pollen from Voegelli Farm (Whittecar and Davis 1982) and Oak Crest Bog (Meyers and King 1985), just outside of the watershed in northern Illinois (Fig. 5), shows the existence there of a forest or open woodland dominated by pine and spruce during the period 47 to 24 ka (Heusser and King 1988). Lastly, pollen from south-central Illinois shows that during the same time interval that region was characterized by prairie with oak/hickory stands (Grüger 1972a, 1972b; Teed 2000).

During the late part of the Wisconsin glaciation, between 35 and 10 ka, the margin of the Laurentide ice sheet advanced in a series of sublobes that eventually covered the entire Great Lakes watershed (Grimley 2000). This expansion and subsequent withdrawal was characterized by a series of major and minor advances and retreats, evidence for which occurs in the stratigraphic record and in the cross cutting relationship of end moraines. Although no compelling evidence exists to suggest that any major advance or retreat of a sublobe was out of phase with the rest, minor advances or retreats may not have been synchronous (Mickelson *et al.* 1982).

Ice initially advanced into the Michigan basin about 26 ka (Winters and Rieck 1991), and by about 20 to 19 ka the ice margin reached its most southerly position in east-central Illinois (Fig. 5) (Hansel and Johnson 1992). There, it fluctuated for about 3,000 years, at times retreating as far north as the southern end of the Michigan basin before readvancing to a slightly less southerly position than that of the previous advance (Hansel and Johnson 1992, Johnson *et al.* 1997).

In the Erie basin a minor ice advance may have occurred as early as about 27.5 ka, but it probably did not extend much beyond the southern shore of Lake Erie (Fullerton 1986). This was followed by an ice advance into the Erie and Huron basins at about 22 ka (Karrow 1984), and by about 21 ka ice extended well into southwestern Ohio and south-central Indiana where it coalesced with ice flowing southward out of the Michigan basin (Fig. 5) (Fullerton 1986). There, the ice margin fluctuated for nearly 4,000 years, at times overriding forests and burying logs beneath glacial sediments (Lowell *et al.* 1990, 1999a). Farther east, however, in northeastern Ohio, northern Pennsylvania, and western New York, the advance of ice was constrained by the Appalachian Plateau and did not extend as far south as it did further west (Muller and Calkin 1993). By about 20.5 ka the ice margin in the Superior basin also advanced well into northern Wisconsin and east-central Minnesota (Match and Schneider 1986).

Retreat of the Last Ice Sheet

After reaching its maximum extent and oscillating near that position for several thousand years, the southern margin of the Laurentide ice sheet began a general retreat northward into the Great Lakes watershed, interrupted by several major readvances that culminated at about 15.5, 13.0, 11.8, and 10.0 ka (Fig. 6). The readvance of about 15.5 ka extended into central and western New York, northeastern Pennsylvania and northeastern Ohio, central Ohio and Indiana, southern lower Michigan, northwestern Indiana, northeastern Illinois and eastern Wisconsin (Mickelson *et al.* 1982). In places, the limit of this advance is marked by a well developed end moraine, but in some areas it is poorly defined due to collapse topography or obscured by a later readvance of the ice margin. The readvance of about 15.5 also deposited till which today locally forms steep bluffs along the Lake Erie (Barnett 1987, Szabo and Bruno 1996) and southern Lake Michigan shore (Hansel 1983, Mickelson *et al.* 1984, Monaghan *et al.* 1986b).

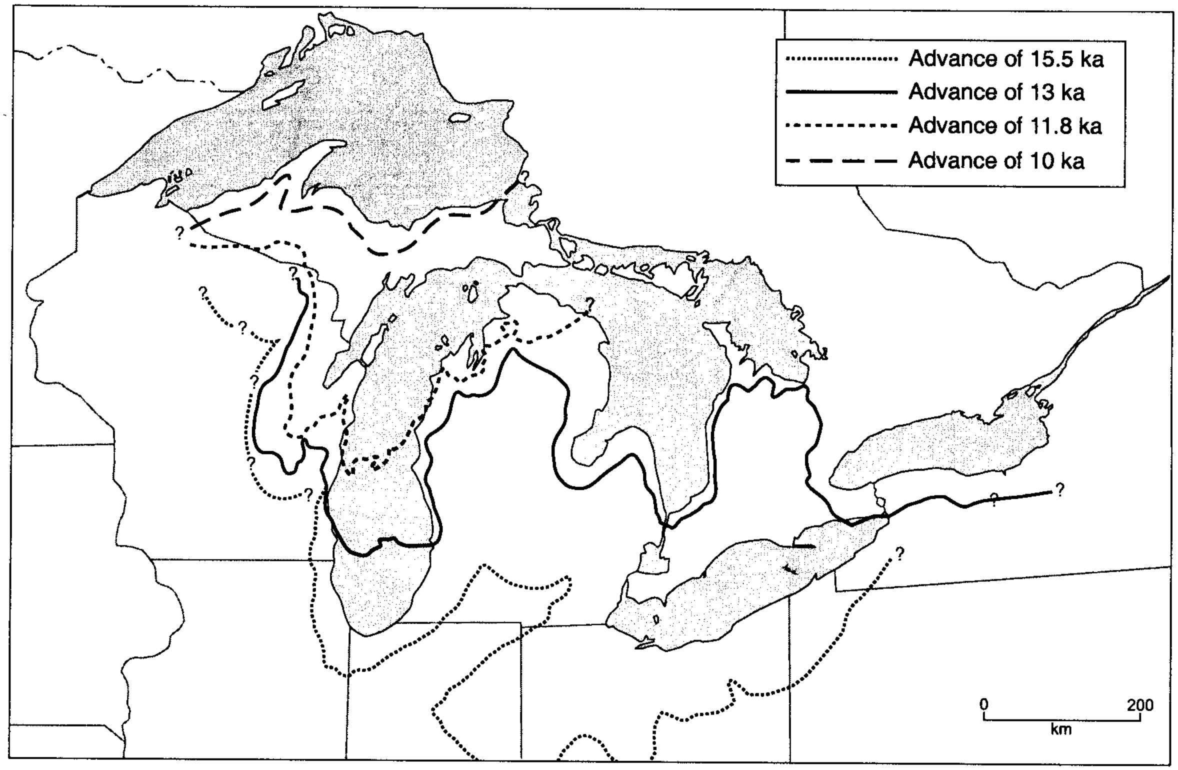

FIG. 6. Limit of the ice readvances of 15.5, 13.0, 11.8 and 10.0 ka. Adapted from Larson* et al. *(1994).

The readvance of about 13.0 ka (Fig. 6) covered two thirds of the Michigan basin and most of the Huron basin and the eastern end of the Erie basin. Its maximum extent is marked by the Port Huron end moraine system which extends across southern Michigan (Leverett and Taylor 1915, Fullerton 1980, Blewett 1991) and can be traced almost continuously eastward into the Ontario basin (Cowan *et al.* 1975, Barnett 1992). A till sheet associated with this advance has been identified at a number of locations in the Great Lakes watershed, but is particularly well exposed in bluffs along the shore of Lake Ontario (Dreimanis and Goldthwait 1973, Calkin and Muller 1992), the western shore of Lake Michigan (Acomb *et al.* 1982, Hansel and Johnson 1992), and the southeastern shore of Lake Huron (Cooper and Clue 1974).

An especially rapid and vigorous readvance at about 11.8 ka covered only the northern half of the Michigan basin and northwestern end of the Huron basin (Fig. 6) (Mickelson *et al.* 1982, Schaetzl 2001). Nowhere did it produce a prominent end moraine, but it did leave behind a till sheet that is well exposed in bluffs near Two Creeks, Wisconsin (Fig. 5), where it overlies a spruce and pine forest bed radiocarbon dated at about 11.8 ka (Broecker and Farrand 1963, Leavitt and Kalin 1992, Kaiser 1994). A till sheet associated with the same readvance also occurs in northern lower Michigan and overlies a bryophyte bed near Cheboygan, Michigan (Fig. 5), believed to be about the same age as the forest bed near Two Creeks, Wisconsin (Larson *et al.* 1994).

The readvance of about 10.0 ka (Fig. 6) extended to the southern rim of the Superior basin where it built a prominent end moraine across much of northern Michigan (Drexler *et al.* 1983, Farrand and Drexler 1985). When at its maximum it buried a pine and spruce forest near Lake Gribben, Michigan (Fig. 5) which has been radiocarbon dated at around 10 ka (Drexler *et al.* 1983, Lowell *et al.* 1999b, Pregitzer *et al.* 2000). Following this readvance the ice margin retreated northward for the last time and by about 9 ka completely withdrew from the Great Lakes watershed (Barnett 1992, Karrow *et al.* 2000).

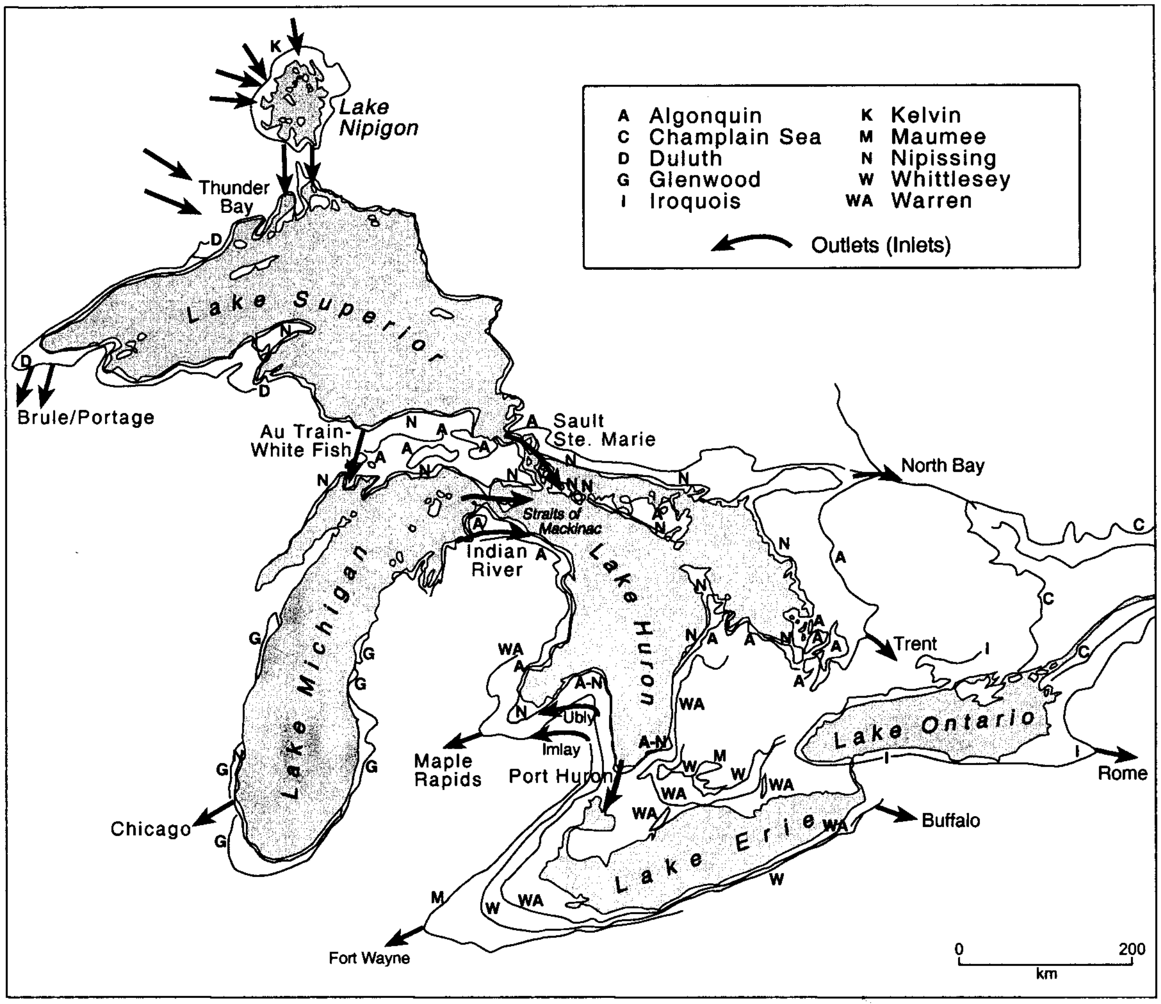

FIG. 7. ***Locations of shorelines of prominent proglacial lakes in the Great Lakes watershed, and their spillways and outlets. After Karrow (1984).***

EVOLUTION OF THE GREAT LAKES

The record of glacial and postglacial lakes in the Great Lakes watershed consists of bars, lake floor sediments, and abandoned spillways and channels, as well as wave-cut cliffs, beach ridges, and deltas that indicate shorelines now abandoned (Fig. 7). Not surprisingly, the spillways are of particular importance because they controlled the level of lakes and were subject to periodic ice blockage, isostatic uplift, and downcutting. In general, they can be divided into those that channeled water from one lake basin to another and those that discharged into river systems that drained out of the watershed (Fig. 7). A set of spillways, located along the north shore of the Superior basin, also periodically channeled water into the watershed from glacial Lake Agassiz, which occupied an ice dammed basin in parts of Manitoba, Ontario, Saskatchewan, Minnesota, and the Dakotas (Teller 1985).

Besides contributing information about lake levels, former shorelines shed light on the history of glacial retreat and isostatic uplift. For example, the shoreline associated with glacial Lake Algonquin (Fig. 7), which formed about 11 ka and was one of the last of the major glacial lakes to develop in the watershed, has been traced almost continuously from the Huron and northern Michigan basins to the eastern end of the Superior basin (Eschman and Karrow 1985, Farrand and Drexler 1985, Larsen 1987). (It cannot be traced further west into the Superior basin because ice associated with the readvance of about 10.0 ka destroyed the shoreline.) An increase of the Algonquin shoreline elevation at the present time of over 150 m, northeastward across the Huron basin, is due to differential isostatic uplift since abandonment of the lake just after about 10.5 ka (Fullerton 1980, Karrow *et al.* 1975, Kaszycki 1985).

The Glacial Great Lakes

As the southern margin of the Laurentide ice sheet receded, large proglacial lakes formed in the lake basins between high topography to the south and the ice margin to the north. These lakes generally widened and expanded northward with the retreat of the ice margin, but during readvances they were displaced by glacier ice and made smaller (or totally displaced).

The first of the proglacial lakes formed about 16 ka when the ice margin retreated northward some distance into the watershed (Fig. 8A). The lakes formed included glacial Lake Milwaukee which occupied the southern part of the Michigan basin and presumably drained south into the Mississippi drainage system (Schneider and Need 1985), and glacial Lake Leverett which occupied the Erie basin and for a time drained east over the Niagara Escarpment to the Atlantic (Mörner and Dreimanis 1973, Karrow 1984, Fullerton 1980, Barnett 1992). A third unnamed glacial lake probably existed at about this time in the southern part of the Huron basin and drained south into the Erie basin (Karrow 1984). All three of the lakes, however, lasted less than 1,000 years and were completely destroyed by the ice readvance of about 15.5 ka.

Glacial Lakes from 15.5 ka to 13 ka

With subsequent ice-marginal retreat after the ice readvance of about 15.5 ka, large glacial lakes once again formed in the Great Lakes watershed (Fig. 8B). One of these was glacial Lake Chicago which developed in the Michigan basin and drained south into the Mississippi drainage system via an outlet near Chicago (Figs. 7, 8B) (Leverett and Taylor 1915). The lake persisted through the readvances of about 13.0 and 11.8 ka, except for two brief intervals when drainage appears to have been temporarily diverted northward across the northern tip of southern Michigan (Hough 1963, 1966; Bretz 1964; Hansel *et al.* 1985; Colman *et al.* 1994a). It included at least two phases or lake levels, the Glenwood phase being the earliest and highest, followed by the Calumet phase (Leverett and Taylor 1915; Hough 1963, 1966; Bretz 1964; Hansel *et al.* 1985; Colman *et al.* 1994a). Exactly when the Calumet phase began has been debated (Farrand and Eschman 1974, Fullerton 1980, Hansel *et al.* 1985) but one explanation proposed for the change in lake levels has been periodic downcutting of the outlet near Chicago (Bretz 1951, 1955; Kehew 1993). More recently, however, it has been proposed that major changes alone in the net input of glacial meltwater and precipitation entering the Michigan basin could explain the differences in lake levels (Hansel and Mickelson 1988).

Another important proglacial lake was glacial Lake Maumee. It developed in the Erie basin (Fig. 8B) and formed slightly earlier than did glacial Lake Chicago (Leverett and Taylor 1915). Initially, it discharged southwest into the Wabash Valley via an outlet near Fort Wayne, Indiana, and thence to the Ohio River (Figs. 7, 8B) (Leverett and Taylor 1915, Eschman and Karrow 1985). Later, however, as the ice margin retreated northward, the lake expanded into the southern part of the Huron basin and drained north and then west across the "thumb" of southern Michigan via an unknown buried outlet (Leverett and Taylor 1915, Eschman and Karrow 1985). Following a minor readvance of the ice margin it drained via the Imlay Channel that today cuts north and then west across the axis of the "thumb" (Leverett and Taylor 1915, Eschman and Karrow 1985). Once across the "thumb," drainage from glacial Lake Maumee was directed west down the glacial Grand River Valley and into glacial Lake Chicago (Leverett and Taylor 1915, Eschman and Karrow 1985). The change from a southwestern to a northern outlet led to several lake phases known as Maumee I, II, and III (Leverett and Taylor 1915). It has been suggested, however, that the last two phases may have continued to drain southwest via the outlet near Fort Wayne despite the opening of a northern outlet (Bleuer and Moore 1972).

Late in the history of glacial Lake Maumee drainage was also directed into Early Lake Saginaw which developed in the Saginaw lowlands as the ice margin retreated northward (Leverett and Taylor 1915, Eschman and Karrow 1985). Early Lake Saginaw drained west into the glacial Grand Valley via an outlet near Maple Rapids, Michigan (Leverett and Taylor 1915, Eschman and Karrow 1985) and it is possible that the outlet may have been intermittently incised due to an increase in the amount of meltwater from the retreating ice front (Bretz 1951), or from the addition of water by catastrophic

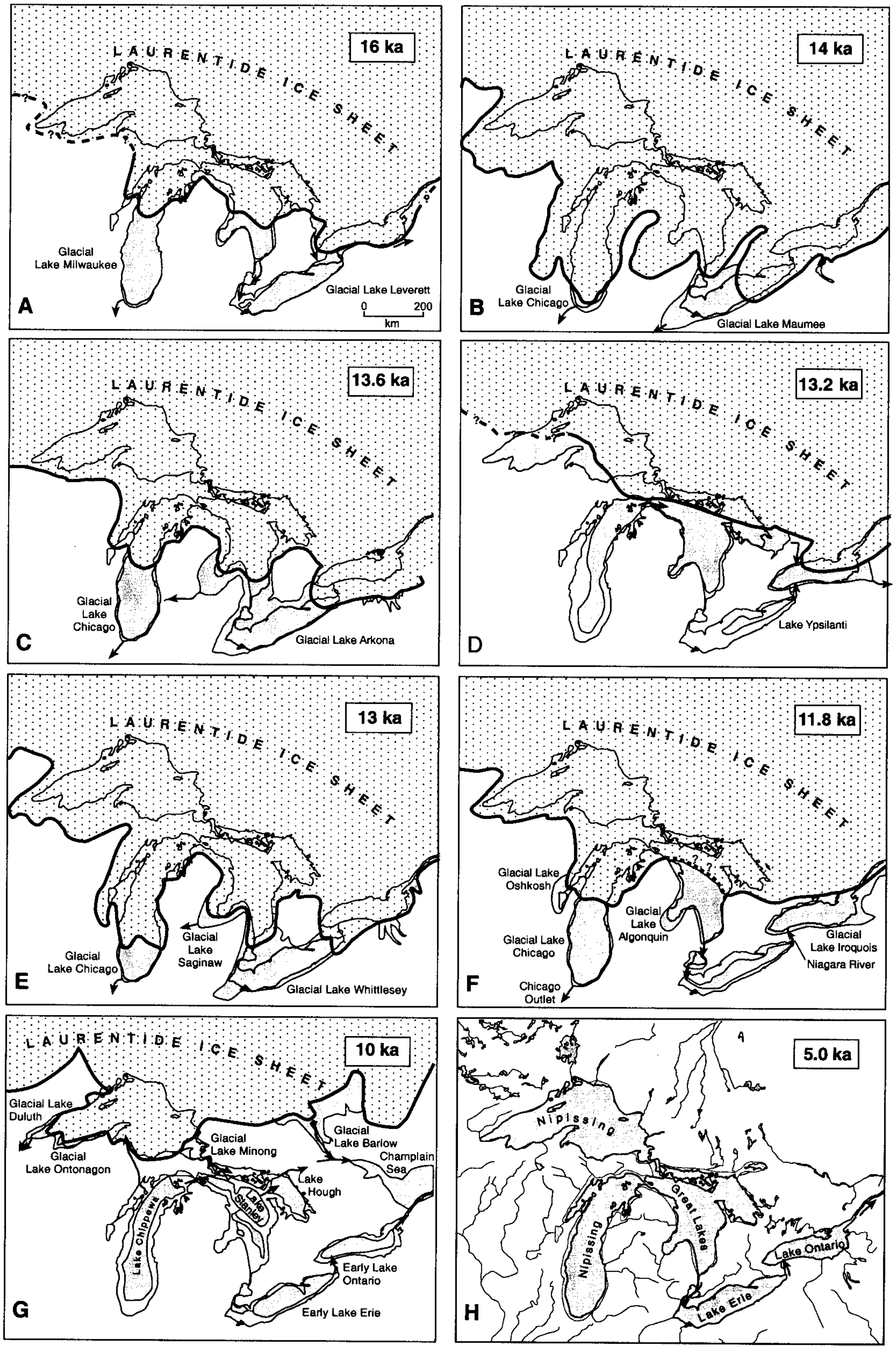

FIG. 8. Locations and general extent of the major proglacial lakes associated with the retreat of the Laurentide ice sheet.

draining of glacial Lake Maumee (Eschman and Karrow 1985).

Both Early Lake Saginaw and glacial Lake Maumee were replaced by glacial Lake Arkona just prior to about 13.5 ka when the ice margin finally retreated far enough north across the "thumb" to allow for the merging of water in the Huron and Erie basins with that in the Saginaw lowlands (Fig. 8C) (Leverett and Taylor 1915, Fullerton 1980). Glacial Lake Arkona, which drained into the glacial Grand Valley via the outlet near Maple Rapids, then expanded northward and eastward against the retreating ice margin until a still lower outlet became ice free near the Trent lowlands in Ontario (Eschman and Karrow 1985; Barnett 1985, 1992). As a result, glacial Lake Arkona drained and was replaced by two eastward-draining low-level lakes, glacial Lake Ypsilanti in the Erie basin (Kunkle 1963) and an unnamed lake in the Huron basin (Fig. 8D) (Dreimanis and Karrow 1972, Fullerton 1980, Eschman and Karrow 1985). At about the same time the Indian River lowland and Straits of Mackinac may have become ice free, resulting in a drop in the level of glacial Lake Chicago and development of a low-level lake (intra-Glenwood low phase) in the Michigan basin that drained east into the Huron basin via the Straits (Fig. 8D) (Hough 1958, 1963, 1966; Hansel *et al.* 1985; Monaghan and Hansel 1990).

Glacial Lakes from 13 ka to 11.8 ka

The readvance of about 13 ka to the Port Huron moraine closed off the Trent lowland outlet and produced glacial Lake Saginaw in the Huron basin and glacial Lake Whittlesey in the Erie basin. It also closed off the Indian River lowlands which lead to restoration of glacial Lake Chicago in the Michigan basin. Glacial Lake Whittlesey drained north and then west across the "thumb" of Michigan into glacial Lake Saginaw via the Ubly channel, whereas glacial Lake Saginaw drained west into the glacial Grand Valley via Maple Rapids and thence into glacial Lake Chicago (Figs. 7, 8E) (Leverett and Taylor 1915, Fullerton 1980, Eschman and Karrow 1985, Calkin and Feenstra 1985). Subsequent retreat of the ice margin, however, resulted in the northward and eastward expansion of both glacial Lake Saginaw and glacial Lake Whittlesey, and eventually they coalesced to the level of glacial Lake Warren which formed after down-cutting of the spillway at Maple Rapids, possibly by catastrophic drainage from glacial Lake Whittlesey (Kehew 1993). As the ice margin continued to retreat north and eastward, a new lower outlet for glacial Lake Warren became exposed just south of Buffalo, New York which drained east into the Mohawk River valley (Eschman and Karrow 1985, Calkin and Feenstra 1985).

The new outlet near Buffalo led to drainage of glacial Lake Warren and the establishment of glacial Lake Grassmere and later glacial Lake Lundy in the Huron and Erie basins. These new lakes also appear to have drained eastward into the Mohawk River Valley (Eschman and Karrow 1985, Calkin and Feenstra 1985) but their actual outlets remain uncertain (Hough 1958, Eschman and Karrow 1985, Calkin and Feenstra 1985). In addition, it is generally believed that a short-lived low-level lake, glacial Lake Wayne, developed in the Huron and Erie basins before the level of glacial Lake Warren fell to the level of glacial Grassmere, and that this lake drained east into the Mohawk River Valley (Fullerton 1980, Muller and Prest 1985).

When a still lower outlet just north of Buffalo became ice free soon after 12.5 ka, water levels in the Erie basin fell resulting in Early Lake Erie which drained north via the Niagara River and then into the Ontario basin where glacial Lake Iroquois was expanding northeastward against the retreating ice margin and draining via an outlet near Rome, New York (Figs. 7, 8F). Early Lake Erie probably occupied only the eastern end of the Erie basin, but in time isostatic uplift of the outlet caused the waters to deepen resulting in the lake's westward expansion (Fig. 8G) (Lewis *et al.* 1966, Calkin and Feenstra 1985, Coakley and Lewis 1985).

At about 12 ka Early Lake Algonquin was also forming in the Huron basin. It initially drained south into the Erie basin via an outlet at Port Huron (Hough 1958, 1963), but soon its level fell to a low (Kirkfield) phase when an outlet (Fenelon Falls) near the head of the Trent River Valley again became ice free and channeled water east into glacial Lake Iroquois and thence into the Mohawk Valley via Rome, New York (Eschman and Karrow 1985). Possible deglaciation of the Straits of Mackinac at about the same time may also have caused glacial Lake Chicago to drain and allow the low-water phase of glacial Lake Algonquin to extend into the Michigan basin (Broecker and Farrand 1963, Fullerton 1980, Larson *et al.* 1994). This low-water phase in the Michigan basin persisted until ice associated with the readvance of about 11.8 ka once more blocked the Straits, thus reestablishing glacial Lake Chicago (Hansel *et al.* 1985, Colman *et al.*

1994a). Thereafter, the Straits once more became ice free, which allowed glacial Lake Chicago to drain for the last time and be replaced once again by the low-water phase of glacial Lake Algonquin (Lewis *et al.* 1994, Hansel *et al.* 1985).

Glacial Lakes after 11.8 ka

Continued retreat of the ice margin northward from the north slope of the Adirondack Mountains just after 11.8 ka ultimately caused glacial Lake Iroquois to drain in a series of steps that led to Early Lake Ontario (Sutton *et al.* 1972, Muller and Prest 1985). This low-level lake (Fig. 8G) drained to the Champlain Sea via the Upper St. Lawrence Valley, and eventually was replaced by present Lake Ontario when its outlet was raised by isostatic uplift (Anderson and Lewis 1985).

In the Huron basin, continued northward retreat of the ice margin was followed by uplift of the outlet (Fenelon Falls) near the head of the Trent Valley which caused glacial Lake Algonquin to slowly transgress southward (Eschman and Karrow 1985). An environment dominated by strong easterly winds and high waves at this time is suggested by the presence of immense spits that trail off to the northwest of islands in glacial Lake Algonquin, in northern lower Michigan (Krist and Schaetzl 2001). Whether glacial Lake Algonquin again reached the level of the Port Huron outlet is debated (Eschman and Karrow 1985, Finamore 1985, Kaszycki 1985, Larsen 1987, Lewis and Anderson 1992) but, due to the uncovering of a succession of lower outlets at the head of the Ottawa River near North Bay, Ontario shortly after 11 ka (Fig. 7), the level of glacial Lake Algonquin rapidly fell to form a series of short-lived post Algonquin lakes. The last in the series of lower lake stages occurred around 10 ka (Fig. 8G) and include Lake Hough in the Georgian Bay basin (Lewis 1969), Lake Stanley in the Huron basin (Hough 1955, Eschman and Karrow 1985, Karrow 1989), and Lake Chippewa in the Michigan basin (Hough 1955, 1958; Hansel *et al.* 1985).

In the Superior basin the ice margin retreated from the south rim about 11 ka, resulting in the formation of glacial Lake Ontonogan which drained west into glacial Lake Duluth located in the western end of the basin (Clayton 1983). From there, drainage was south into the St. Croix River Valley via the Brule and Portage outlets, and thence down the Mississippi River. Continued retreat of the ice margin, however, led to drainage of these lakes and allowed glacial Lake Algonquin to flood northward into the Superior basin until it was displaced by the ice associated with the readvance of about 10.0 ka (Farrand and Drexler 1985). Glacial lakes Ontonogan and Duluth were then reestablished along the southwestern rim of the basin and a new glacial lake, Minong, was formed in the eastern end of the basin (Fig. 8G) (Farrand and Drexler 1985). This new lake drained east via an outlet just west of Sault Ste. Marie and expanded to fill the Superior basin as the ice margin once again retreated northward. Subsequent downcutting of the lake's outlet however, caused the level of the lake to drop in several steps and eventually stabilize on a bedrock sill near Sault Ste. Marie to form Lake Houghton (Farrand and Drexler 1985).

The Postglacial Lakes

Deglaciation of the north rim of the Superior basin by about 9 ka (Barnett 1992) marked the end of the glacial history of the Great Lakes watershed. Isostatic uplift continued, however, and its effect has had a major role in the postglacial evolution of the Great Lakes (Leverett and Taylor 1915, Hough 1958, Larsen 1985b). For example, uplift of the outlet (North Bay) for Lake Hough in the Georgian Bay basin forced waters to again transgress southward and westward across the Huron and Michigan basins as well as northwestward across the Superior basin until they once more spilled over outlets at Port Huron and Chicago, creating the Nipissing Great Lakes about 5.5 to 5.0 ka (Figs. 7, 8H) (Eschman and Karrow 1985, Hansel *et al.* 1985, Farrand and Drexler 1985). With downcutting of the Port Huron outlet, however, the Chicago outlet was abandoned at about 4 ka and the level of Lake Nipissing eventually fell to that of modern Lake Huron and Lake Michigan (Eschman and Karrow 1985, Hansel *et al.* 1985, Thompson and Baedke 1997, Baedke and Thompson 2000). Commensurate with, as well as shortly after, the high Nipissing stillstand, sediment accumulation rates in the Lake Michigan basin slowed considerably, presumably due to a decrease in coastal erosion (Rea *et al.* 1980, Colman *et al.* 2000). An intermediate level common to all three basins known as Lake Algoma also may have developed about 3 ka, but there is some debate as to whether strands related to this level were the result of temporary stabilization of the outlet at Port Huron or whether this high stand was climatically induced (Larsen 1985b). High lake stands during the Holocene were times of rapid bluff erosion and concomitant building of large perched dune fields at various sites in the Great Lakes watershed (Arbogast 2000, Loope and Arbogast 2000).

Recently, study of the internal architecture and timing of development of beach ridges at five sites around Lake Michigan has produced four relative curves of late Holocene lake level for Lake Michigan that show similar lake level variations but record different rates of vertical movement due to glacial isostasy and/or tectonism (Thompson and Baedke 1997). Subtraction of best-fit rates of rebound from these curves has resulted in residual lake level curves that, when combined and smoothed, show the timing and magnitude of prominent lake level events in the Michigan basin since the end of Lake Nipissing (Baedke and Thompson 2000). These events include, among others, a rapid fall in lake level associated with the end of Lake Nipissing, a significant rise in lake levels probably associated with Lake Algoma, and a second significant rise in lake levels centered about 1,700 years ago (Baedke and Thompson 2000).

In the upper Great Lakes, isostatic uplift of the bedrock sill near Sault Ste. Marie, coupled with drainage of the Nipissing Great Lakes, led to the separation of the Superior and Huron basins about 2.2 ka and formation/isolation of modern Lake Superior (Farrand 1969, Farrand and Drexler 1985).

Impact of Glacial Lake Agassiz

The Great Lakes played an important role in the dispersal of waters from glacial Lake Agassiz which developed about 11.7 ka as the margin of the Laurentide Ice sheet retreated northward into the Hudson Bay watershed (Fig. 9) (Fenton *et al.* 1983, Teller 1985). The maximum extent covered at one time by glacial Lake Agassiz was about 350,000 km^2 (Teller 1994). During the interval from 11 to 10.5 ka (Moorhead Phase, Fig. 9) and from 9.5 to 8.5 ka (Nipigon Phase) the Superior basin appears to have received large continuous volumes of outflow from the lake (Teller and Thorleifson 1983; Teller 1985, 1987). Some, however, have questioned whether glacial Lake Agassiz actually drained east into the Superior basin during the interval from 11 to 10.5 ka and, instead, propose that it drained west at that time (Minning *et al.* 1994).

The outflows into the Superior basin from glacial Lake Agissiz were the result of retreat of the ice margin in the vicinity of Thunder Bay and the Lake Nipigon basin (Fig. 7) which exposed several low outlets to the Superior basin (Teller and Thorleifson 1983; Teller 1985, 1987). Glacial Lake Agassiz outflow was interrupted, however, during the interval from 10.5 to 9.5 ka by the ice readvance of about 10.0 ka, which temporarily blocked outlets near the Nipigon basin (Teller and Thorleifson 1983, 1987).

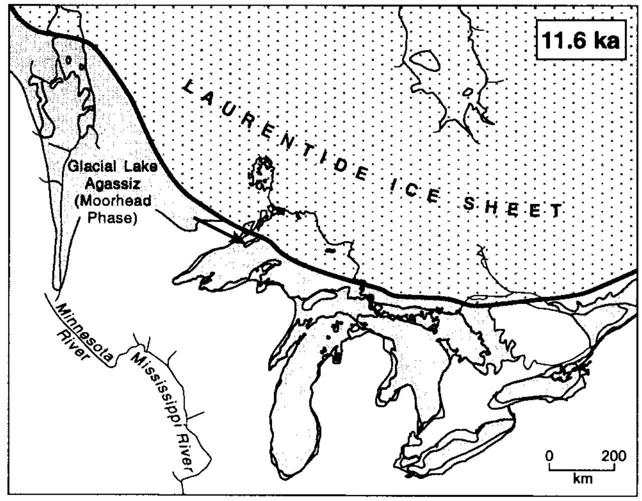

FIG. 9. Map showing glacial Lake Agassiz and Great Lakes ~10.8 ka, during the first episode of eastward discharge of glacial Lake Agassiz into the Great Lakes basin. Modified from Lewis and Anderson (1989), and Lewis **et al.** *(1994).*

At times, outflows from glacial Lake Agassiz into the Superior basin must have been punctuated by periodic outbursts of catastrophic discharge (Teller and Thorleifson 1983, Teller 1985, Leverington *et al.* 2000). The first of these outbursts probably occurred shortly after 11 ka and spilled eastward into the Superior basin, flooding glacial Lake Duluth. From there, drainage would have been south into the Michigan basin via the Whitefish-Au Train channel in northern Michigan (Fig. 7), resulting in flooding of glacial Lake Algonquin (Clayton 1983). Shortly thereafter, however, as the ice margin retreated further north, subsequent catastrophic outbursts passed eastward through glacial Lake Minong in the Superior basin and directly into glacial Lake Algonquin in the Lake Huron basin via the St. Marys River Valley (Drexler *et al.* 1983, Clayton 1983). Outbursts occurring during the interval from 9.5 to 8.5 ka were likewise directed eastward by Sault Ste. Marie to the Huron basin (Clayton 1983, Teller and Thorleifson 1983, Teller 1987), which at that time had already fallen to its Stanley low level (Prest 1970, Eschman and Karrow 1985, Barnett 1992, Lewis *et al.* 1994). These outbursts are

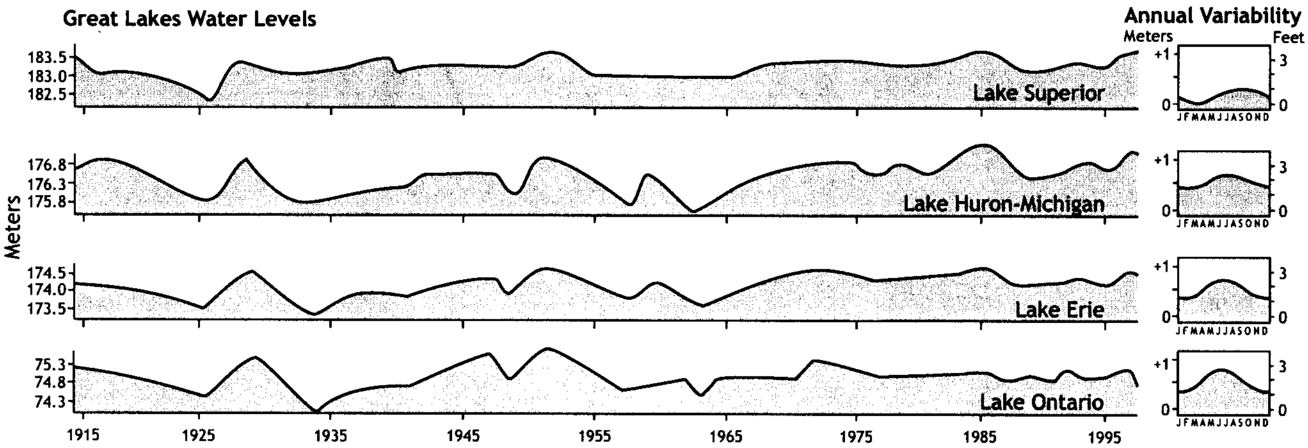

FIG. 10. Lake level fluctuations on the Great Lakes, since 1915 and annually. After Botts and Krushelnicki (1988).

thought to have flooded Lake Stanley, producing temporary higher lake levels known as the Early and Main Lake Mattawa highstands (Lewis *et al.* 1994).

The catastrophic discharges from glacial Lake Agassiz must have had a profound effect on the water level in the basins of the upper Great Lakes (Teller 1985) and some of these diversions might have resulted in brief surges in lake levels on the order of 20 m (Farrand and Drexler 1985). Evidence for such surges, however, may exist only in sediments at the bottoms of the lakes. For example, the Wilmette bed which occurs in lake bottom sediments within the Michigan basin has been attributed to an episode of catastrophic discharge from glacial Lake Agassiz after 11 ka (Teller 1987; Colman *et al.* 1994a, b, c). In the Lake Huron basin, however, no equivalent to the Wilmette bed has been observed. However, negative excursions in the ^{18}O isotopic composition of ostracodes and bivalves in southwestern Lake Huron and eastern Lake Erie have been associated with this discharge (Lewis and Anderson 1992, Lewis *et al.* 1994, Rea *et al.* 1994). A negative isotope excursion in biogenic carbonate and changes in ostracode assemblages within younger lake bottom sediments in the Michigan basin also appear to mark a second episode of catastrophic discharge around 8.9 ka, which may have had at least two pulses (Colman *et al.* 1994b, c). Within the Huron basin this episode appears to be marked by a positive isotopic excursion in ostracodes and bivalves (Rea *et al.* 1994). "Varved" red and gray clays within lake bottom sediments of the Superior basin also may coincide with the second episode of catastrophic discharge (Teller 1985).

HISTORICAL LAKE LEVEL CHANGES

Historically, the Great Lakes have been in a constant state of flux (Karrow and Calkin 1985, Thompson 1992). As lake levels change and as shorezone materials erode or accrete, the physical location of the water-land interface, i.e., the shoreline, changes. As noted above, an indeterminable number of lake level changes have occurred since 16 ka, as outlets changed subtly in location, were downcut or uplifted, or as new, lower outlets opened. Recently, however, human activity, which is focused and most dense along the shorelines, has added a new dimension to the changes taking place on the Great Lakes.

Changes in lake levels over the past 6 ka are fairly well documented, although not all data are in complete agreement (Thompson 1992, Anderton and Loope 1995, Lichter 1995, Delcourt *et al.* 1996, Arbogast and Loope 1999, Baedke and Thompson 2000). Many lake level changes in the last 8 ka have had dramatic effects on ancient cultures (Larsen 1985a, Butterfield 1986). Many late Holocene lake level fluctuations were, at least in part, climatically-driven (Fraser *et al.* 1975, 1990; Hamblin 1987), still others were due to changes occurring at the lake outlets (Calkin and Feenstra 1985, Larsen 1985b, Monaghan *et al.* 1986a). Mid-range cycles, at least those observed by European settlers and their descendants since the early 1800s, are well documented for the Great Lakes (Fig. 10)

(Bishop 1990). These cycles occur despite the modulation of flows from several of the lake outlets and the placement of outlet control works at critical points within the basin (Brunk 1968, Derecki 1985, Bishop 1990). Human-induced factors that also affect lake levels include diversion of water from the basin, consumptive use of water by municipalities, construction of control structures, and land use alteration (e.g., urbanization of the watershed, deforestation) (Davis 1976, Bruce 1984, Bishop 1990, Changnon and Changnon 1996). Increased evaporation under a possible greenhouse-enhanced climate, coupled with even more consumptive use of the Great Lakes waters, could lead to lower lake levels in the near future (Bruce 1984, Hamblin 1987, Sousounis and Bisanz 2000).

Highest lake levels usually occur during prolonged periods of higher than normal precipitation and/or cooler than normal temperatures (Croley 1986, Bishop 1990). The latter acts to reduce evaporative losses from the lakes. Notable, high lake levels were observed on most of the Great Lakes from 1853 to 1862, 1882 to 1887, 1928 to 1931, 1943 to 1955, in the early and late 1970s, and again in the mid 1980s (Fig. 10) (Quinn and Sellinger 1990). At times such as these, shoreline erosion is often pronounced (Hadley 1976, Changnon 1993, Folger *et al.* 1994, Fraser *et al.* 1990). Notably low lake levels were recently observed in 1926, 1934, and 1964 (Fig. 10), and again from 1999 to 2001.

Imposed upon the longer-term fluctuations are annual and daily cycles (Fig. 10) (Platzman 1966). Annual or seasonal variations in water levels are due to subtle changes in the balance between (1) inputs incident upon the basin, as precipitation, surface runoff, and groundwater contributions, (2) outflows (runoff entering the St. Lawrence River), and (3) evaporation (Bruce 1984). Low lake levels are generally found in mid- to late-winter, when precipitation is minimal and much of what has fallen is tied up on the land as snow (Botts and Krushelnicki 1988). It follows, then, that lake levels are highest in July, after additions of summer precipitation, spring snowmelt and runoff (Fig. 10).

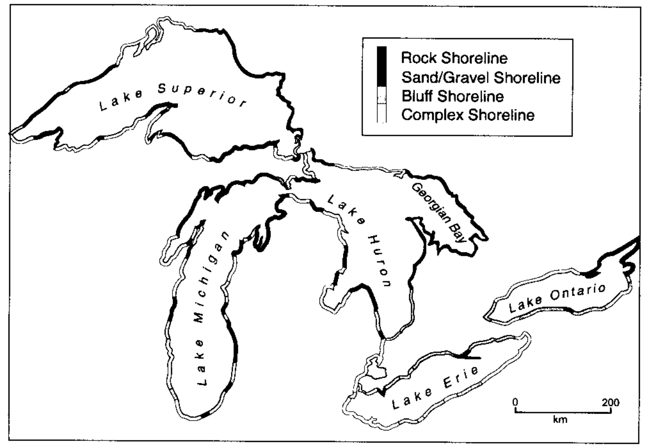

FIG. 11. Generalized shorezone types for the Great Lakes. Source: U.S. Army Corps of Engineers, Detroit.

Shore Types

Several distinct types of shorelines exist on the Great Lakes, each varying in form and composition (Fig. 11). Examples are cliffs of bedrock and drift, wide sandy strands, rocky and rubbly coasts, swampy and marshy flats, among many others. Bedrock cliffs are most common on Lakes Superior and Huron, where hard dolomite or crystalline bedrock has resisted the attack of ice and waves, and where continued wave erosion coupled with slow bluff recession have kept the bases of the bluffs free of sediment. Cliffs 40–80 m in height are common along the northern Lake Superior shoreline (Upchurch 1976, Johnson and Johnston 1995) whereas beaches are often sandy or gravelly along the southeastern margins of the lake (Adams and Kregear 1969). Limestone bedrock and gravel outcrop along much of the shoreline of Lake Huron immediately east of the Mackinac Bridge. High limestone and dolomite cliffs are common wherever the Niagara cuesta intersects Lakes Michigan and Huron, as on the eastern margin of Green Bay (Door peninsula), the Garden, Bruce, and Presque Isle peninsulas, and the western margin of Manitoulin Island (Powers 1958). Coasts in these areas consist of rocky headlands and small pocket beaches with rounded limestone gravels and sands. Bluffs cut into glacial sediments are especially prominent along southeastern Lake Huron, the central part of Lake Michigan (both sides), and the northern and southern shores of Lake Erie (Lee 1975, Pavey *et al.* 1994, Fig. 11).

Many Lake Erie shores are low and marshy, although in some sections bluffs of shale or clay-rich glacial drift are present (Rukavina and Zeman 1987, Carter and Guy 1988). Spits such as Point Pelee, Long Point, Presque Isle, and Cedar Point mark large accumulations of sandy sediments (Pincus 1959). Beaches are poorly-developed on Lake Ontario (Sutton *et al.* 1972). Rather, bluffs of varying heights are quite common. Numerous embayments cut into these cliffs have formed where drowned river valleys enter the lake. Many of these embayments have acted as sand "traps" (Upchurch 1976).

The Great Lakes boast some of the best, wide sand beaches in the world (Chrzastowski *et al.* 1994, Folger *et al.* 1994). The beaches on the eastern shore of Lake Michigan are especially noteworthy in this regard. These sands originated as glacial sediments derived from rocks ground up by the ice into particles 0.05 to 2 mm in diameter and washed out by meltwaters as the ice receded. As Holocene lake levels fluctuated (see above), many of these sands were alternately inundated by rising waters or left high and dry on abandoned beaches. In the latter instance, westerly winds lifted these sands into dunes on the eastern shores of Lakes Huron and Michigan (Olson 1958, Hazlett 1986, Arbogast and Loope 1999). It is the latter dunes that comprise the largest inland dune system, associated with lakes, in the world.

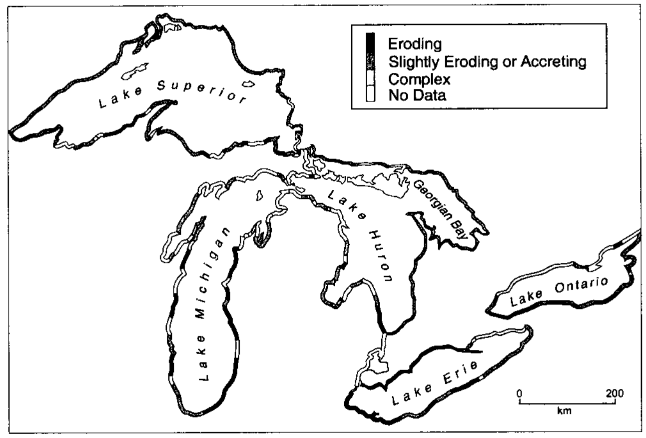

FIG. 12. Generalized areas of shorezone erosion and progradation on the Great Lakes. Source: U.S. Army Corps of Engineers, Detroit.

Recent Shoreline Erosion

Over the past century, humans have controlled lake outlets, regulated flow of water into and out of the lakes, and altered shorezone characteristics (Davidson-Arnott and Keizer 1982), leading to unprecedented rates of erosion along some stretches of shoreline, and to shorezone aggradation in others. Shoreline erosion is an ongoing environmental concern along much of the Great Lakes' coastal areas (Buckler and Winters 1975, 1983; Rasid and Hufferd 1989; Jibson and Staude 1991; Barnes *et al.* 1994). Much of the most serious erosion is occurring along sandy, highly erodible beaches on the eastern coast of Lake Michigan, southern Lake Erie, and Lake Superior, where long east-west fetches lead to large waves during storms (Fig. 12) (May *et al.* 1983, LaMoe and Winters 1989, Rasid *et al.* 1989, Dilley and Rasid 1990, Highman and Shakoor 1998). Sand and shoreline deposits are almost always in a state of flux, as waves and currents move the sediments landward in summer and out to deeper waters in winter. Evidence is mounting, however, that human interference with the hydrologic system has caused it to become imbalanced (Omohundro 1973).

Natural shoreline erosion and recession is a function of the following variables: (1) presence and height of uplands above lake level, (2) composition and erodibility of shorezone materials, (3) exposure to storms, waves, and surges, including storm duration and intensity, (4) lake levels (arguably the major factor in determining rates of bluff retreat (Jibson *et al.* 1994), (5) offshore water profile and beach width, (6) rates of longshore transport of sediment in the coastal zone, and (7) presence or absence of lake ice (Buckler and Winters 1983, Carter and Guy 1988, Lawrence 1994, Angel 1995, Johnson and Johnston 1995, Amin and Davidson-Arnott 1997). Sandy, stratified sediments in association with high lake levels and stormy conditions promote the greatest amount of shoreline recession (Omohundro 1973, Carter and Guy 1988, Jibson *et al.* 1994). Normally, storms and their associated waves pound the beach and undercut bluff slopes, especially if lake levels are high, leading to slumping and bluff retreat and in so doing providing the beach with its main source of sand (Lee 1975, Rukavina and Zeman 1987, Vallejo and Degroot 1988). This is the main way that Great Lakes coasts erode (Jibson *et al.* 1994). The sediment released to the surf zone at the base of eroding bluffs is acted on by waves and longshore currents, to be ultimately moved to deeper waters where much of it is stored in offshore bars. Shorefast ice can shield the bluff from waves and therefore prevent such erosion. Under natural conditions, movement of sand to offshore bars is the main agent by which sand is lost from the shorezone. The beaches are continually replenished by streams that bring sediment to the lakeshores and by erosion of bluffs and other shorezone sediments (Rukavina and Zeman 1987).

The sand in the shorezone is then moved along the shore by waves and offshore currents in longshore transport (Lawrence 1994).

However, two recent types of human intervention have seriously reduced the supply of sand to the shore zone and facilitated the loss of sand to deeper water: (1) dams on rivers that are tributary to the Great Lakes, and most importantly (2) jetties and other engineering structures at river mouths (Omohundro 1973, Shabica and Pranschke 1994). The effects of damming tributaries is obvious—sediment settles out in the relatively still waters of inland reservoirs and is not allowed to be transported to the Great lakes shore. Jetties function differently. They are engineering structures erected at river mouths, resembling two long walls that border both sides of the river and extend from the river banks and mouth, just inland of a harbor, to relatively deep waters in the lake proper. Jetties affect beach replenishment by diverting sand and other sediments that move along the shore by lake processes, into deep water (Bush *et al.* 1996). Once there, these sediments can no longer be transported to the beach by waves and are therefore permanently lost from the beach system. For this reason, coastal erosion is often most severe near harbor structures, rather than at more "open" coasts (Folger *et al.* 1994, Shabica and Pranschke 1994). Dredging of river mouths for shipping and boating purposes can facilitate further transport of sand and sediment into deep water.

Once the main source of sand to the beach system, river mouths and harbors have now become sites of beach impoverishment. Thus, shoreline erosion or retrogression, a natural process, has been much more dominant than has shoreline progradation (Powers 1958), and should be considered, when development along the always-variable Great Lakes shorelines is contemplated.

SUMMARY

The basins that contain the Great Lakes are the product of repeated scour and erosion of relatively weak bedrock by continental glaciers that advanced into the Great Lakes watershed beginning perhaps as early as 2.4 Ma. Most of the scouring, however, probably occurred after about 0.78 Ma when episodic glaciation of North America was much more extensive, with ice cover sometimes extending as far south as Kentucky. The number of times the watershed was completely glaciated is not known for certain because of an incomplete terrestrial record. However, stratigraphic evidence from outside the watershed indicates that glacier ice extended over all or part of the watershed at least six times since 0.78 Ma. Information about the environment present between the glaciations is also limited, but in one location within the watershed pollen and fossils have been found that suggest that during the last interglaciation, the climate was probably similar to or warmer than present.

In contrast to earlier glaciations, the last glaciation of the Great Lakes watershed is well documented by glacial sediments, recessional moraines, and buried organic deposits. It shows that the eastern part of the watershed was first glaciated between 65 to 79 ka and that the ice margin oscillated there until about 25 ka. At this time a boreal forest-tundra environment had also established itself in much of the unglaciated part of the watershed. After about 25 ka, the ice margin advanced from both the north and east to cover all of the watershed and by 18 to 21 ka it extended as far south as the Ohio River and as far west as northern Wisconsin and east-central Minnesota. After about 18 ka the ice margin began to retreat, but that retreat was interrupted by several major readvances at about 15.5, 13.0, 11.8, and 10.0 ka. After 10.0 ka the ice margin continued to retreated northward and by about 9.0 ka completely withdrew from the watershed for the last time.

During the retreat of the ice margin large glacial lakes developed within the basins of the Great Lakes. Their surface elevation and extent varied considerably over time as outlets were either blocked or uncovered by glacier ice. Outlets were also subject to isostatic rebound as well as by channel downcutting, which likewise affected the level of the glacial lakes and the lakes that followed. On several occasions lake levels were affected by catastrophic influx of meltwater from glacial Lake Agassiz which developed outside of the Great Lakes basin.

Issues associated with the modern Great Lakes center on lake levels and shoreline configurations. Lake levels are, and have always been, in a state of flux. Today, lake levels have a typical seasonal variation of less than a half meter. However, larger fluctuations of 1 to 2 m happen during extended dry or wet periods and create problems for lake users who expect lake levels to be essentially static.

Great Lakes beaches are widely variable, from bedrock promontories to wide expanses of fine, washed sand. Many of the sandy beaches are backed up by large dunes, and many of the high

bluffs are topped with perched dunes. High lake levels, coupled with human activities and engineering structures, have accelerated the naturally occurring processes of shoreline erosion and retreat. Low lake levels raise other issues, such as access to marinas and inadequate water depth in harbors and connecting channels for shipping, often necessitating dredging. This litany of issues illustrates the constant change that is the earmark of the Great Lakes, both recently and throughout the past 2+ Ma.

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