J. Great Lakes Res. 31:343–359
Internat. Assoc. Great Lakes Res., 2005

Influence of Lake Michigan Holocene Lake Levels on Fluvial Sediments of the Lower Pigeon River, Wisconsin

Diann S. Kiesel[1,*] and David M. Mickelson[2]

[1]*Department of Geography and Geology*
University of Wisconsin-Baraboo/Sauk County
1006 Connie Road
Baraboo, Wisconsin 53913

[2]*Department of Geology and Geophysics*
University of Wisconsin-Madison
1215 W. Dayton St.
Madison, Wisconsin 53706

ABSTRACT. *Thirty-seven vibracores, extending up to 4.86 m, collected along the lower Pigeon River north of Sheboygan, Wisconsin, were used to interpret Holocene lake-level fluctuations of Lake Michigan. The sediments reflect numerous cycles of degradation and aggradation as well as marshland and submergence. The basal unit is till and glaciolacustrine silt and clay. The river cut through these sediments prior to 6,500 ^{14}C years BP, probably during the Chippewa Low stand, and deposited the lowest gravel unit in the cores. Between 6,500 and 5,500 ^{14}C years BP, yellowish-red fluvial sand and silt were deposited in the northern half of the valley. Aggrading point bar gravel and overbank silt and fine sand throughout the entire valley record the lake rise to the Nipissing level from about 5,500 to 5,000 ^{14}C years BP. Subsequent deposition of organic-rich, muddy palustrine sediment indicates that Nipissing water flooded the valley. A lack of sediments with ages between 5,000–2,000 ^{14}C years BP suggests lack of aggradation, indicating a lowering lake level until about 2,000 ^{14}C years BP. Sandy-silt overbank sediment deposited over the palustrine sediment since 2,000 ^{14}C years BP marks the return of floodplain aggradation as lake level stabilized or rose slightly to the modern level. While the river-mouth sediments are not useful for refining Holocene lake-level curves, they do corroborate major events such as the Chippewa Low, the rise to the Nipissing level followed by a period of declining lake level and fluvial erosion, and the small rise to the modern lake level.*

INDEX WORDS: *Holocene, stratigraphy, fluvial, Lake Michigan, lake level.*

INTRODUCTION

The most common method of studying lake-level fluctuations of the Great Lakes is to relate the elevation of beach ridges and their associated water levels to their time of formation, which is normally determined by ^{14}C-dating of organic sediment in interdune swales (*e.g.,* Thompson 1992, Dott and Mickelson 1995, Johnson *et al.* 1990, Petty *et al.* 1996, Baedke and Thompson 2000, Johnston and Thompson 2000). These studies are not without complications, however (Hansel *et al.* 1990, Lichter 1995 and 1997, Thompson 1992, Baedke and Thompson 2000). Other lake-level interpretations have come from deep-lake cores taken from Lake Michigan (Folger *et al.* 1994, Colman *et al.* 1994). Stratigraphy in cores of dune-dammed lakes near Lake Michigan has also provided information for interpretation of lake-level fluctuations (Fisher *et al.* 2001, Wilkinson and Fuks, 2001). Pollen and macrofossils have been included in lake-level studies as well (Booth *et al.* 2002).

Many studies show that estuarine and fluvial deposits record a variety of late and postglacial events. Coakley and Lewis (1985), Fraser *et al.* (1990), and Larsen (1985) suggest that river-mouth geomorphology and stratigraphy hold important information about the interpretation of Great Lakes levels. A few studies of fluvial terraces have been

*Corresponding author. E-mail: dkiesel@uwc.edu

conducted, but these concentrate mostly on glacial lakes of the Erie (Klotz 1981, Bay 1937) and Huron basins (Karrow 1986, Anderson 1979). Richardson (1991) and Knox and Leigh (1987) studied fluvial terraces along the Sheboygan and Onion rivers, just south of the Pigeon River, and found that most of the terrace levels correspond to glacial, not Holocene, lake levels in the Lake Michigan basin. Similar interpretations were made by Chapel (2000) for the Pigeon River. Stratigraphic and palynological analyses of an abandoned oxbow-channel fill exposed along the Erie lakeshore revealed lake-level fluctuations during the Holocene (Barnett *et al*. 1985). Cowen (1978) interpreted exposures along meandering rivers entering Lake Superior and Lake Huron to indicate alluviation and burial as Nipissing waters rose. Flint *et al.* (1988) and Dalrymple and Carey (1990) studied the sediment record in stream lagoons at river mouths to interpret the past 4,000 years of lake-level and climate-change history for Lake Ontario. Along the Wisconsin-Illinois line, Larsen (1985) studied a beach-ridge complex crossed by four streams whose banks expose the record of stream-mouth deposition associated with changes in base level for the past 2,000 years.

In this paper, the authors use sediments near the mouth of the Pigeon River valley to interpret Lake Michigan lake-level fluctuations throughout the Holocene. Because Lake Michigan is the base level for rivers entering the lake, lower reaches of these streams should have responded to lake-level change. A rise of lake level, due to wetter and cooler climatic conditions or uplift of the outlet, should have caused flooding in river mouths, resulting in fluvial aggradation and increasingly marshy conditions, and, if lake level rose enough, possibly lacustrine deposition in the lower reaches of the valleys (Fraser *et al.* 1990). As water level declined due to drier and warmer climatic conditions, outlet downcutting, or uplift of the shoreline relative to the outlet, fluvial down-cutting through the lacustrine, wetland, and stream sediment should have occurred. Thus, the sediments preserved in the Pigeon River mouth should record Holocene lake-level fluctuations.

BRIEF LATE QUATERNARY HISTORY OF LAKE MICHIGAN

The present Great Lakes came into existence during retreat of glacier ice between about 16,000 and 11,000 radiocarbon years ago. As glacier ice retreated into the Lake Michigan basin, proglacial lakes occupied the depression in front of the ice. Over the next 5,000 years, lake level fluctuated through several lake phases. Glacial Lake Milwaukee was the first phase. This was followed by three high phases of Lake Michigan, sometimes called Lake Chicago (Glenwood I and II and Calumet), which occurred when glacier advances blocked the Straits of Mackinac and the lake drained through the Cal-Sag Channel and the Des Plaines River valley near Chicago. These were separated by low phases (Intra-Glenwood, Two Creeks) that occurred when ice retreated north of the Straits of Mackinac, exposing lower northern outlets. The Algonquin phase was the last proglacial lake to occupy the Lake Michigan and Lake Huron basins (Hansel *et al*. 1985, Hansel and Mickelson 1988, Schneider and Need 1985, Larson and Schaetzl 2001).

As the glacier continued to retreat, lower outlets to the north of the Straits opened and allowed even lower lake levels. The North Bay outlet was utilized for about 5,000 years, creating very low levels in the Lake Michigan basin (Chippewa low phase). Isostatic rebound gradually raised the elevation of the North Bay outlet, and by about 6,000 years BP, lake level was at its present level and continuing to rise to the higher Toleston level during the Nipissing phase. Southern outlets were again occupied. Shortly thereafter, the North Bay outlet was abandoned due to rebound, and continued erosion of the Port Huron outlet caused the abandonment of the Chicago outlet when erosion of that outlet was slowed by bedrock in the channel floor. Lake-level fluctuations after the Nipissing phase are attributed to climatic fluctuations superimposed on continued isostatic adjustment of the region (Fraser *et al.* 1990, Larsen 1985, Hansel and Mickelson 1988, Hansel *et al.* 1985, Larson and Schaetzl 2001).

GEOLOGIC SETTING OF THE PIGEON RIVER

The Pigeon River lies just north of Sheboygan, Wisconsin, and flows into western Lake Michigan (Fig. 1). The entire drainage basin of Pigeon River is approximately 200 km^2 and drains rural land, except for the small community of Howards Grove and the edge of the northern part of the city of Sheboygan. It was chosen for this study because of its river-mouth geomorphology as well as its location close to the same isobase for glacial rebound as the outlet (Clark *et al.* 1990, Larson 1994, Coordinating Committee on Great Lakes Basic Hydraulic and

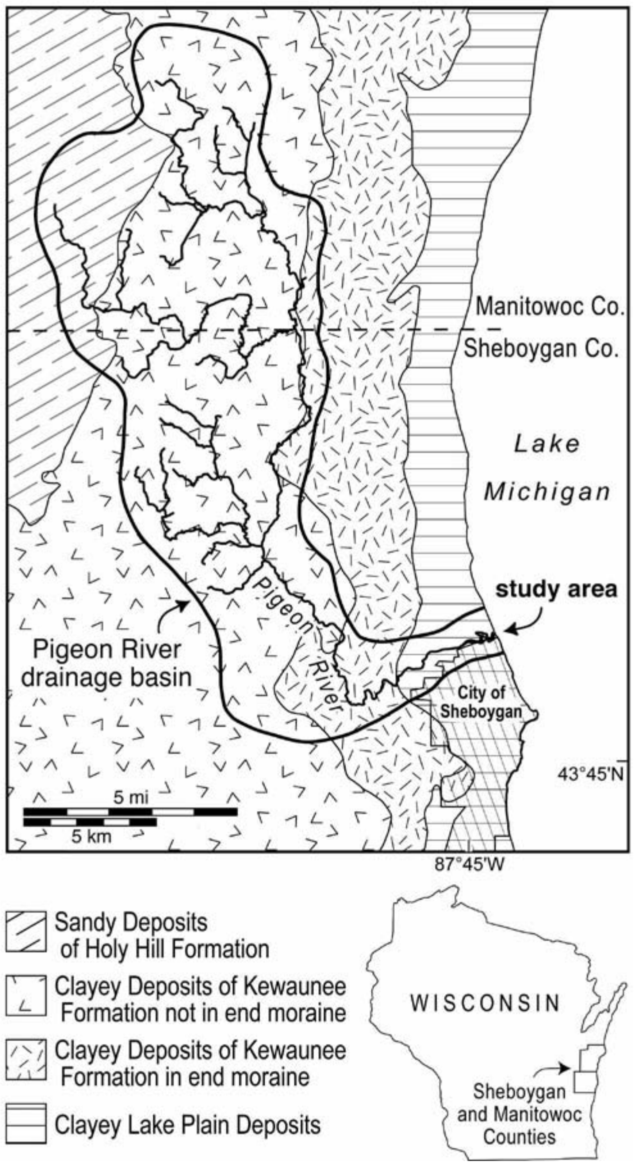

FIG. 1. Location of the Pigeon River drainage basin and the study site (geology after Principato 1999, Chapel 2000).

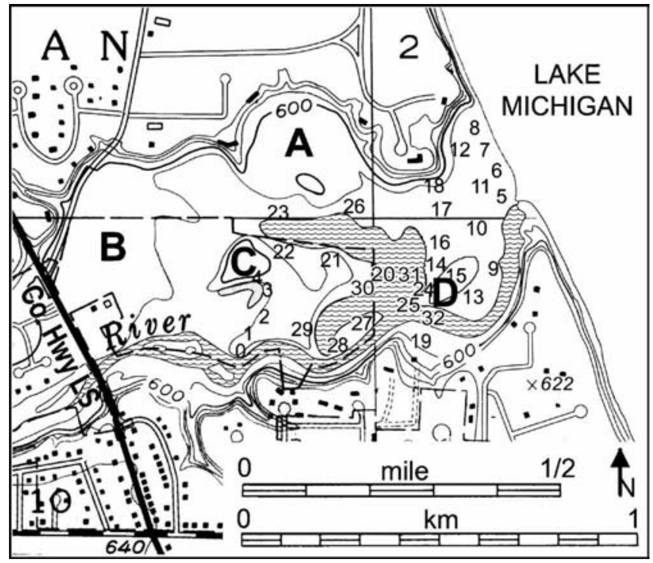

FIG. 2. Topographic map of the mouth-area of the Pigeon River. Numbers indicate locations of core sites. See text for explanations of A, B, C, and D. (From USGS Sheboygan North, WI, 7.5′ topographic series, revised 1994; contours in feet.)

Hydrologic Data 1977, and Tushingham and Peltier 1991).

The geomorphology of the lower reach of the Pigeon River valley (Fig. 2) suggests that this area may have been a freshwater estuary during high lake-level stands. The mouth area is wide, ranging from 0.35 km to 0.8 km, and narrows rapidly upstream at a bedrock sill near County Hwy LS (Fig. 2). The elevation of the floodplain is low and changes little for more than 1.5 km upstream from Lake Michigan. Steep bluffs rising to an elevation of 192 m (630 ft) bound both sides of the valley. Thus, any Holocene rise in lake level was confined to the valley, creating either an aggrading river or, if high enough, embayed conditions in the lower reach of the valley. Likewise, during times of lower lake level, the river remained confined within the valley walls.

The lower valley contains numerous oxbow scars as well as two oxbow lakes, the largest one of which is still connected to the river. Most of the valley floor in the study area lies at elevations between 177 m (580 ft) and 178.5 m (586 ft). Today, the river is only slightly sinuous and flows mainly along the south bluff.

A sand and cobble bay-mouth bar is present across the mouth of the river. The crest of the bar is usually about 1 to 1.5 m above the lake level, and the width of the bar varies from approximately 6 to 15 m. The bar blocks the direct entrance of the river into the lake. The river flows through a narrow (usually 0.6 to 2 m) and sometimes rather deep (approximately 1 to 1.3 m) channel through the bar. Channel width, depth, texture of exposed sediments, and even location vary with wave and current activity, sometimes within a matter of days, but generally the river drops about 0.3 m across the bay-mouth bar.

Some areas that rise above the low-lying floodplain deserve special attention. Area A (Fig. 2) is a south-sloping terrace located along the north bluff of the valley, north of the large oxbow. Area B (Fig. 2) is a higher terrace of broader extent, mostly composed of till over shallow (~1.5 m) bedrock. The highest feature in the valley, Area C, is a hill just north of the small oxbow. The northwest side of this hill was oversteepened by river-cutting at some time in the past. This hill also has on its south side what appears to be a subtle bench that is close to the Toleston level (184 m, 604 ft) and which may be an erosional feature of the Nipissing phase. Between the large oxbow and Lake Michigan is a northeast-southwest trending ridge that does not show on the topographic map (letter D and an oval indicate its location and trend on Fig. 2). The core taken from this ridge indicates that it is another erosional remnant of till and glaciolacustrine sediment.

The entire drainage basin is underlain by Silurian dolomite (Alden 1918; well records of Wisconsin Geological and Natural History Survey), and the depth to bedrock varies considerably throughout the small study area. Alden (1918) and field observations indicate bedrock at or near the surface in the western quarter of the study area (Fig. 3). It crops out near the bridge of Hwy LS and is very near the surface just downstream near the large island in the river. Soil probing on the east side of the island and on low terraces to the south of the river failed to penetrate more than 0.6 m anywhere because of bedrock. The bedrock surface slopes steeply to greater depths just to the east of the river island. Cores 1, 2, 3, and 4, located just south of hill C (Fig. 2), are up to 3.5 m deep and did not hit bedrock. None of the other cores penetrated to bedrock. Examination of well records from north and south of the present valley also indicate deep bedrock east of Hwy LS.

The surficial sediment of the drainage basin is related to glaciation. The west side of the basin is underlain by sandy, gravelly Holy Hill Formation deposits (Fig. 1). The remainder of the basin is underlain by more clayey till of the Kewaunee Formation and silty and clayey lake sediment (Acomb *et al.* 1982, Principato 1999, Chapel 2000). From the lakeshore inland to about 195 m (640 ft) in elevation, the till is capped by thin, patchy, glaciolacustrine sand, silt, and clay (Bay-Lakes Regional Planning Commission 1997, Principato 1999, Chapel 2000).

The closest lake-level gages are at Milwaukee and Sturgeon Bay. Averaging those data from the National Oceanic and Atmospheric Administration (NOAA) and converting to International Great Lakes Datum (IGLD) 1955, the average lake level during this study was 176.4 m (579 ft), 0.1 m higher than the historical lake-level average in this area.

METHODS

Various physical barriers (steep slopes, soft and marshy areas, etc.) prohibited the use of truck-mounted drill rigs, so vibracoring was the only viable alternative for obtaining sediment cores. Normally, procedures dictate that only one 7.6-cm diameter vibracore tube per site be used (Smith 1984). In this area, however, single-use insertions failed to penetrate to sufficient depths in most places due to the variety of sediment types, so a technique of multiple-insertions was developed on site. Several tubes of varying lengths were prepared at the start of each hole. The shortest tube was inserted first and removed, the second longest was then immediately inserted, and so on. In most holes, three to four lengths were inserted. Because sloughing of material often occurred with each change of tube length, marbles were dropped into the hole immediately after a core segment had been extracted in an attempt to mark a rough boundary between the sloughed material and the in-place sediment. Detailed descriptions of the cores are recorded in Kiesel (1998). The orientation of each core segment was marked prior to extracting it from the ground so that all cores could all be cut open on their east-west axis, revealing a north-south (parallel to the valley) cross-section. The elevation of each core site was surveyed by either transit or total station using known elevations of a bench mark on the Hwy LS bridge over the Pigeon River and from known elevations of fire hydrants on the south bluff and on Hwy LS.

Cores were sealed with aluminum foil and duct tape in the field and transported to the lab. The cores were opened by cutting (using a circular saw with a carbide blade) through the metal on opposite sides of the tube so as not to disturb the sediment. A copper wire was then pulled through the sediment to split the core. Wood and organic material exposed in the open core were immediately removed, wrapped in aluminum foil, put in Whirl-Pak bags, and refrigerated until processed for ^{14}C dating.

Sieving was done on selected samples to establish standards for visually inspecting the sediments.

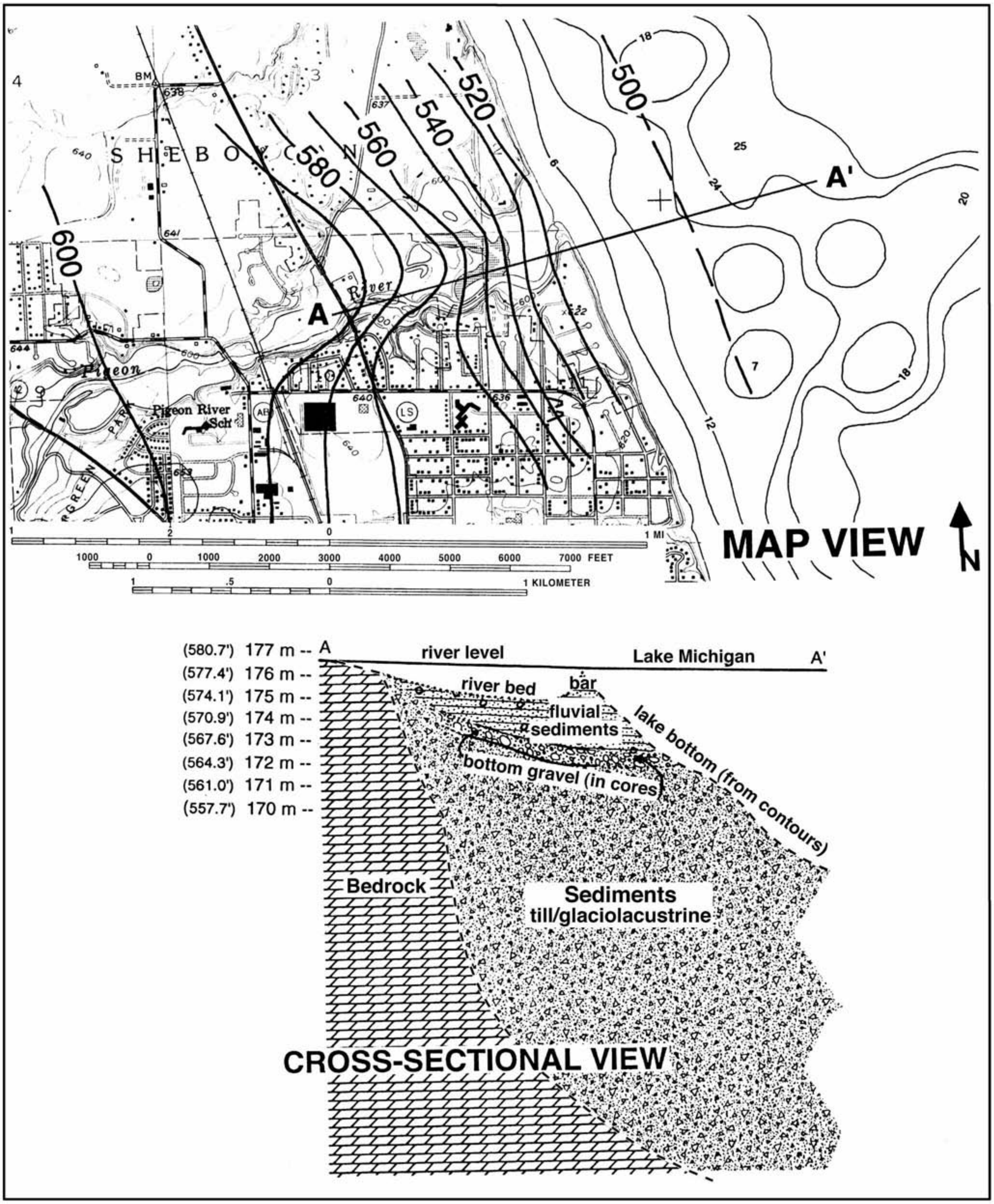

FIG. 3. Bedrock contour map (top) and diagrammatic cross-section (bottom). On map, contours are in feet (CI = 10 ft); SI units are shown in cross-section. (Constructed from WG&NHS water-well records, cores, and from Alden, 1918; topographic base map is USGS Sheboygan North, WI, 7.5′ series, revised 1994.)

Counts of the lithology, roundness, and sizes of grains were made on nearly all the gravel layers in the cores. These data indicate that all gravel layers are of the same origin, presumably winnowed from the glacial sediments of the adjacent bluffs (Kiesel 1998).

Carbonate content for representative samples was determined by Chittick analysis (Dreimanis 1962). Carbonate ranges from approximately 42% to 53%. Most of the carbonate is dolomite, presumably from the local Paleozoic bedrock (Kiesel 1998).

Selected samples of diatoms, snails, bivalves, and pollen were examined with the assistance of Dr. M. Winkler, D. Lewis, L. Kitchel, and Dr. L. Maher, Jr., respectively. They provided little conclusive paleoenvironmental information and are discussed in Kiesel (1998).

STRATIGRAPHY AND INTERPRETATION OF THE SEDIMENTS

In order to interpret geologic history, sediments were grouped into lithostratigraphic units, each interpreted to represent a specific paleoenvironment. Interpreted units are summarized in Table 1. The unit definitions are based mostly on definitions and interpretations of selected sediments in other studies (McDowell 1980 and 1983, Bettis *et al.* 1992, Platt and Wright 1992, Cant 1982, Barnett *et al.* 1985) and somewhat on modern sediments in the Pigeon River (Cores 31 and 32).

Interpretation of lithostratigraphy allowed units in the cores to be correlated (Figure 4). Fine and coarse channel-bed deposits are grouped together as are the laminated and massive fluvial sand and silt

TABLE 1. Summary of diagnostic criteria used to define Pigeon River lithostratographic units.

UNITS	CRITERIA for defining units
Modern Overbank	silt, sandy silt; very dark brown; gradual lower boundary; some soil development
Organic Layer	organics; dark; usually within fine floodplain sequence
Beach sand (modern)	generally medium sand, can contain well-rounded pebbles
Palustrine	silty clay, light gray (mainly 10YR 6/2 to 7/2), usually highly calcareous
Pre-modern floodplain	(layers relatively thin compared to channel bed deposits)
Vertical Accretion overbank deposits (fine sequence)	silt, fine sand; massive, or more commonly laminated/bedded; often contains layered organic material (leaf litter)
Lateral Accretion point bar deposits (coarse sequence)	fine to medium sand, some fine gravel, interbedded with silt layers; often contains coarse detrital organics either scattered or in layers; possible cross-bedding
Channel bed	(relatively thick compared to individual floodplain layers)
fine sequence	silt, fine sand; usually dark brown; massive; can contain organic fragments
coarse sequence	medium sand to coarse gravel in coarse, sandy matrix; poorly sorted; can grade upward into fine sequence; can contain layers of concentrated coarse detrital organics
Pre-modern Fluvial sand and silt	fine-to-medium sand and silt, sand imbedded in matrix of colordistinctive yellowish-red (5YR 5/6 to 5/8) silt w/ some clay
massive	mostly non-stratified
laminated or bedded	sand and silt; stratified; can contain some organics
Till and glaciolacustrine sediments	clay-rich diamicton or clay, silt; reddish-brown (10R 6/2 to 6/3, 5 YR 6/3); massive to laminated; some rhythmic layers

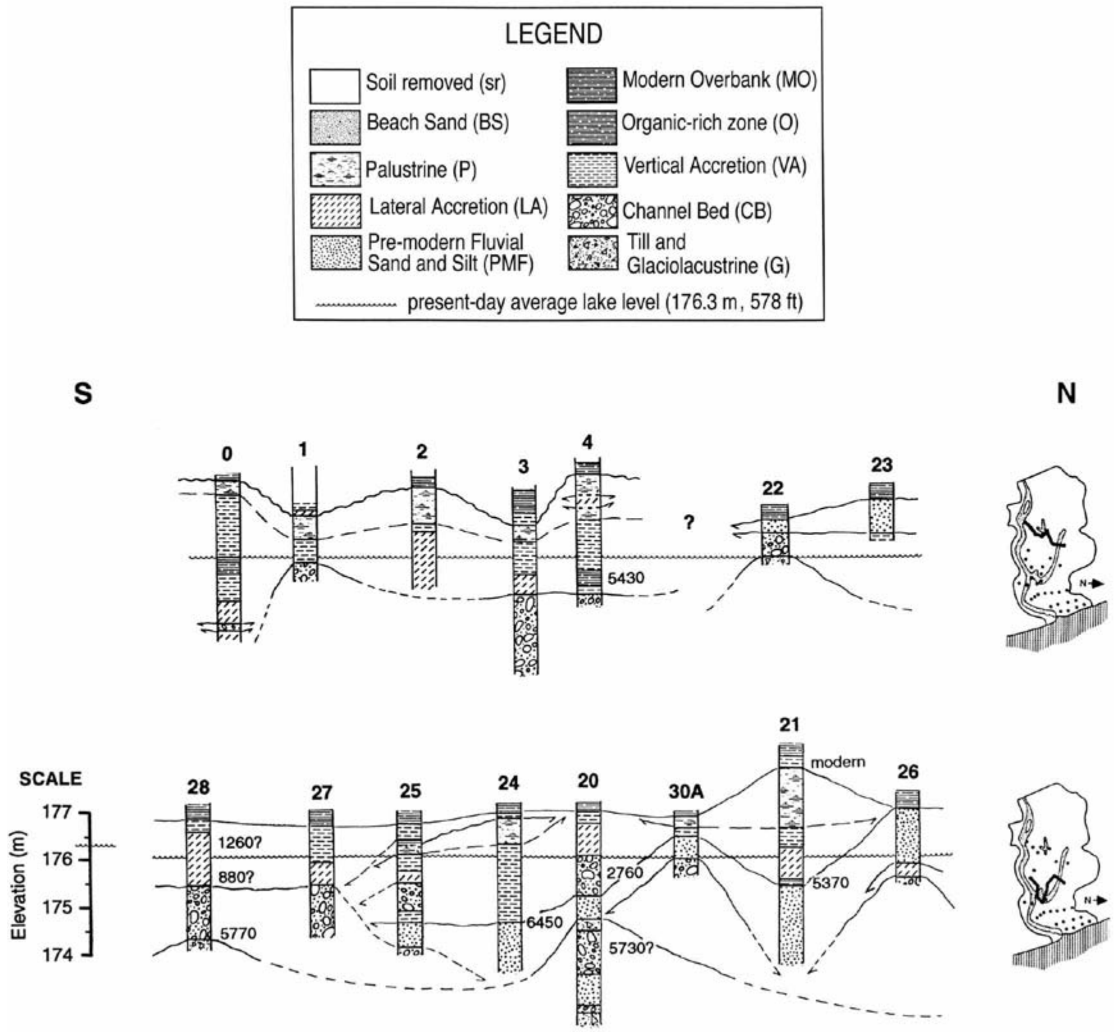

FIG. 4A. Cross-sectional correlations of stratigraphy in across-valley (N-S) transects, western part of valley. Inset maps show locations of transects.

units of the older sequence. Although lateral and vertical accretion deposits are represented with different patterns (because they are the most prominent units in the valley), they are correlated as one unit on these diagrams. Correlations are based primarily on texture of sediments, elevations, and stratigraphic relationships of like units in adjacent cores and secondarily on ages of units.

Till and Glaciolacustrine Sediments (G in Fig. 4)

The lowest stratigraphic unit encountered in the cores is present mostly in the southern half of the valley and is composed of Kewaunee Formation diamicton and glaciolacustrine silt and clay. Cores 9, 13, 15, 20, 22, 22A, and 28 (Fig. 4) contain the reddish-brown diamicton that appears to be till or clay. Many cores did not extend deep enough to encounter this sediment, indicating that the top of the pre-Holocene sediment has a very irregular surface, probably due to erosion by the river. In fact, Core 15 (Fig. 4), taken from ridge D in Figure 2, is nearly all glacial sediments, indicating that the ridge is an erosional remnant. Core 13, just south of Core 15, and Cores 9, 20, 22, 22A, and 28 all have Kewaunee Formation diamicton or glaciolacustrine silt and clay at elevations ranging from 172.7 m (567 ft) to 177.8 m (583 ft) (Fig. 4). Some glaciolacustrine sediments, such as those in Cores 9 and 15, show rhythmic bedding, and Core 9 contains clay drop-stones.

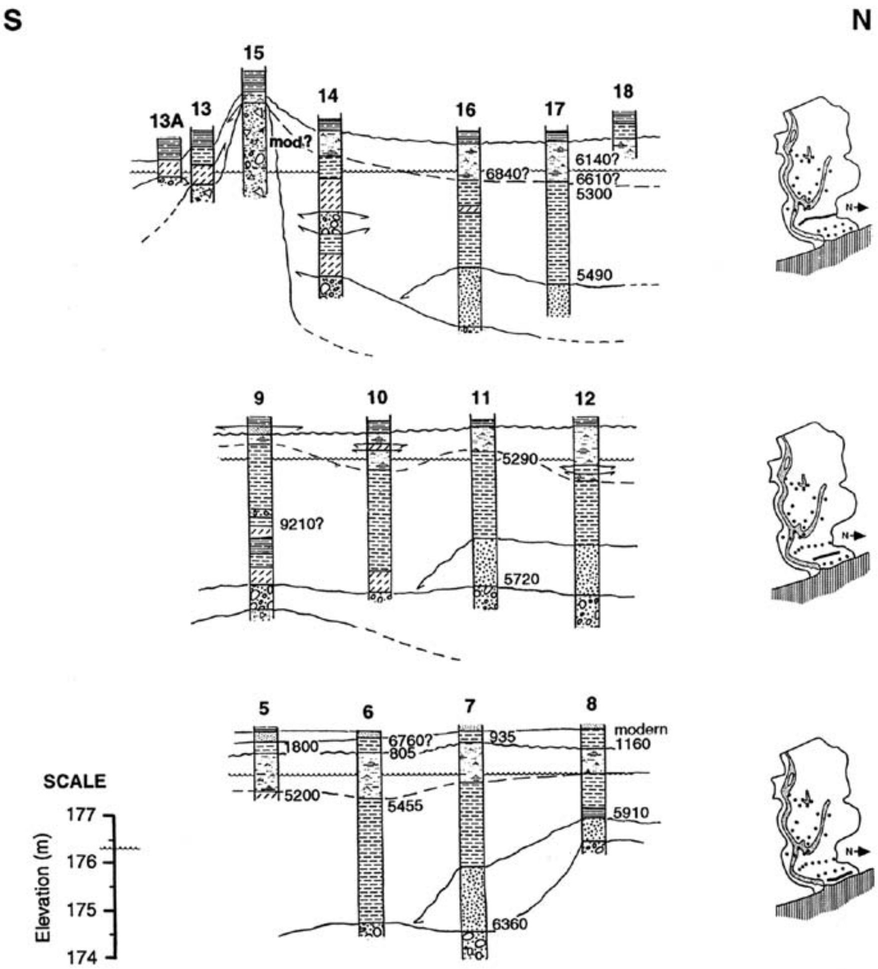

FIG. 4B. Cross-sectional correlations of stratigraphy in across-valley (N-S) transects, eastern part of area. Inset maps show locations of transects.

Channel-Bed Deposits (CB in Fig. 4)

Sediments interpreted to be channel-bed deposits occur as both a coarse- and a fine-grained unit. Almost all cores contain one or both of these units. Cores 0, 4, 7, 8, 9, 10, 11, 12, 13A, 14, 15, 16, 19, 20, 22, 22A, 25, 26, 27, 28, 29, and 30 all contain the coarse gravel unit; Core 3 contains both fine- and coarse-grained sediments. The coarse-grained unit usually is composed of gravel in a sandy matrix. All particles are subrounded to subangular and are poorly sorted. In some cores, thin silt layers are interbedded as are thin layers of detrital organic fragments. The coarse-grained units generally fine upward to sand and can grade rapidly upward into the fine-grained unit, if it exists. Where present within a core, the basal contact is always abrupt. However, because many cores end in gravel, the basal contact is not seen, and the thickness of gravel is unknown. The gravel is assumed to be unconformable with underlying glacial (G) sediment.

The coarse-grained channel-bed unit overlies glacial sediment in every core that contains glacial material (Figs. 3 and 4). Knox and Leigh (1987) found similar sediments and stratigraphy in the Onion and Sheboygan river valleys about 5 miles southwest of the Pigeon River location. Several cores contain more than one coarse gravel unit, usually separated by what are interpreted to be lateral and vertical accretion sediments.

The fine-grained unit of channel-bed deposits consists mostly of silt and clay with a small amount of fine sand. The fine-grained unit is not nearly as common or as thick as the coarse-grained material.

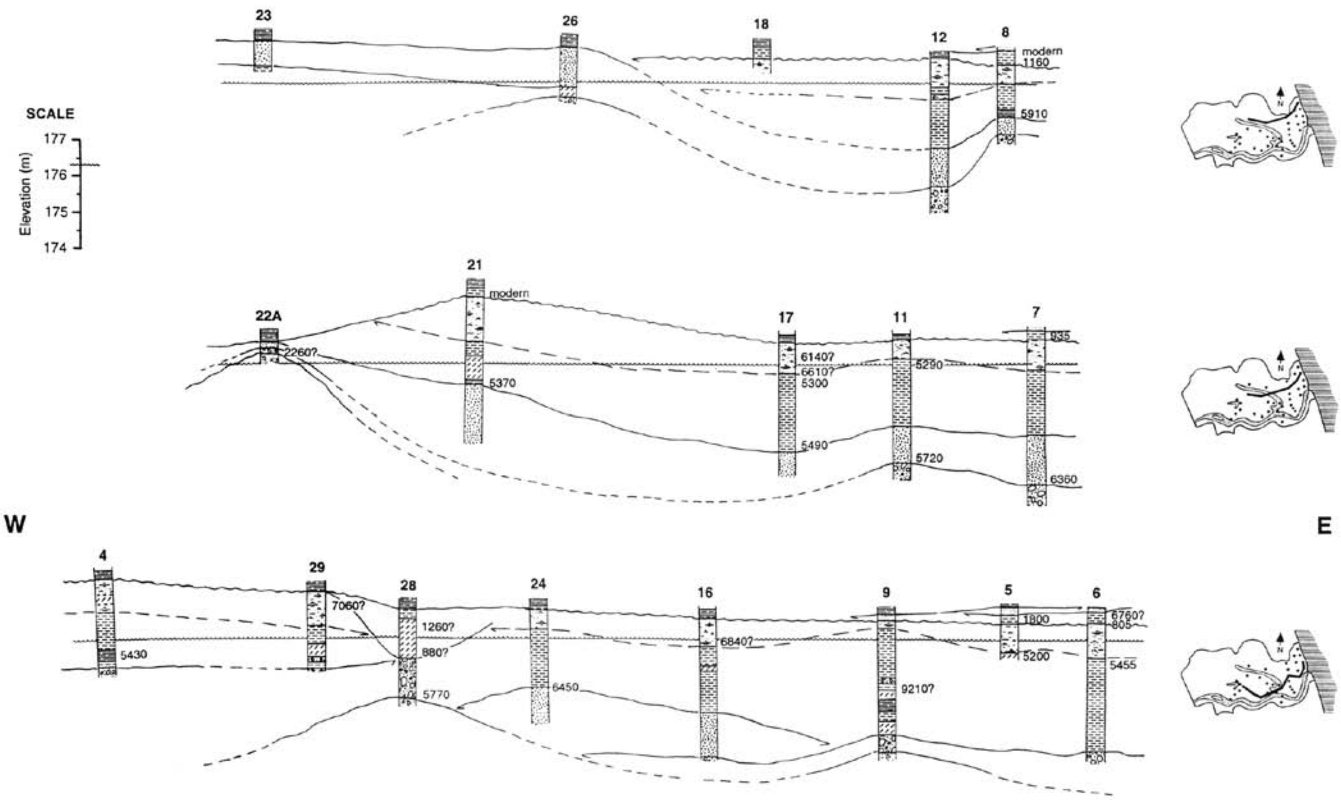

FIG. 4C. Cross-sectional correlations of stratigraphy in down-valley (E-W) transects, northern part of valley. Inset maps show locations of transects.

It appears in only four of the cores (1, 3, 6, and 21), representing only about 10% of what is interpreted to be channel-bottom sediments. The present-day bed of Pigeon River has a similar fine-grained layer 50–60 cm thick, in both the modern channel and the large oxbow. Fine-grained channel-bed deposits can be confused with vertical accretion (overbank) deposits because both have similar texture and color, but usually channel-bed deposits appear more massive.

Pre-modern Fluvial Sand and Silt (PMF in Fig. 4)

Yellowish-red sand and silt are found in the lower segments of most cores in the northern half of the valley and in only Cores 24 and 25 in the southern half of the valley. Because it occurs as high as 177.5 m (582 ft) in elevation in Core 23 and at an even higher elevation (~178 m, 584 ft) farther upvalley, about 250 m west-northwest of Core 23, it probably was more extensive at one time. Except in Cores 23 and 26, it directly overlies the lowest coarse-gravel channel-bed sediment in the cores, so this unit is interpreted to be a pre-modern sequence of fluvial sediments.

The sand is fine-to-medium quartz sand with some non-quartz grains. The entire unit appears yellowish-red (5YR 5/6 to 5/8) because of the color of the silt and clay making up the matrix of the unit. In some cores, the sediment is mostly massive (Cores 22, 22A, 23, 25, and 26), but in other cores, it is laminated or bedded (Cores 11, 12, 20, 30, and 30A). In some cores, both forms exist, but not in any given order: Core 17 has layered sediments over massive sediment; Cores 7, 8, 16, and 24 have massive sediment overlying layered sediments; and Core 21 has massive sediment between layered sediments.

Pre-modern Floodplain Deposits (VA and LA in Fig. 4)

Sediments interpreted to be floodplain deposits consist of lateral (point bar) (LA) and vertical (overbank) (VA) accretion deposits. They are the most common as well as the thickest (2 to 3 m) stratigraphic unit. Almost all cores contain one or

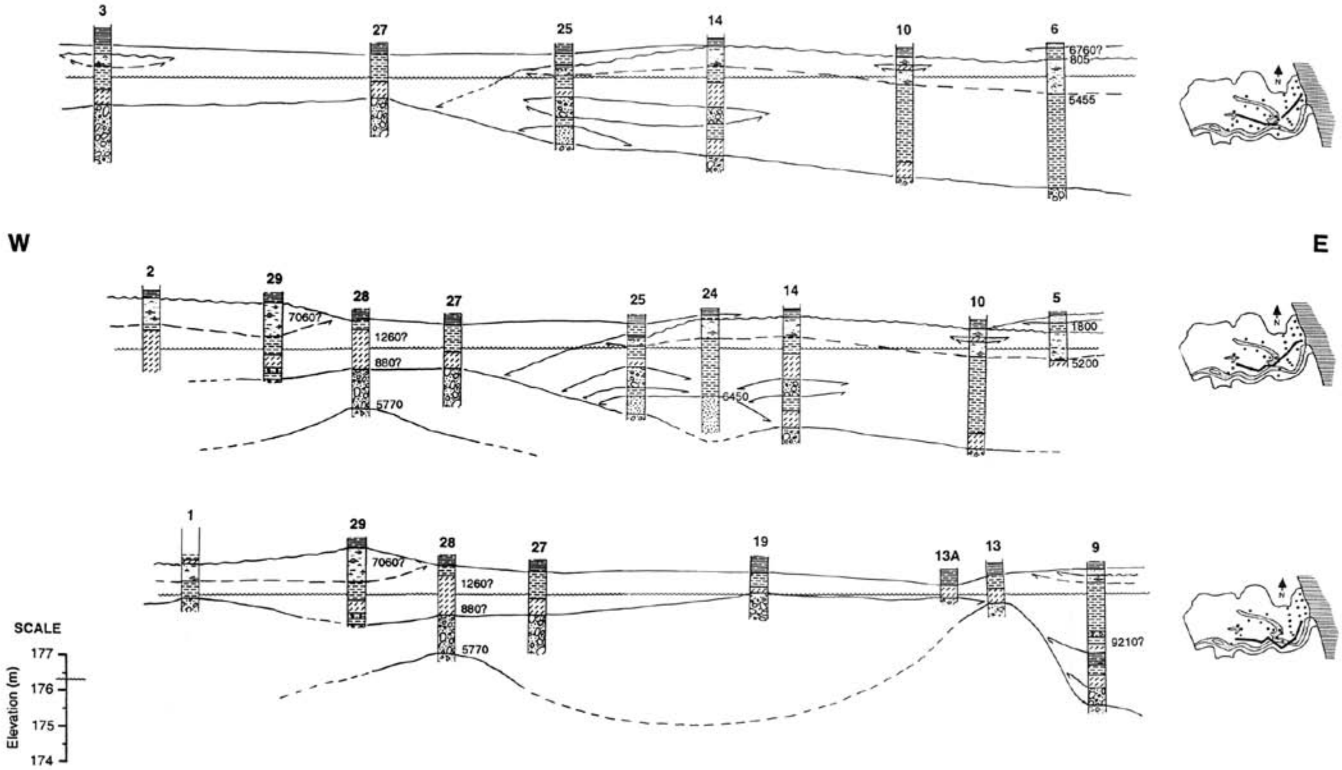

FIG. 4D. Cross-sectional correlations of stratigraphy in down-valley (E-W) transects, southern part of valley. Inset maps show locations of transects.

both units below the present-day soil, which is considered to be composed of modern overbank sediments; only Cores 22 and 22A lack older floodplain deposits. In Cores 13 and 28, floodplain sediments directly overlie glacial sediments. In Cores 1, 3, 4, 6, 9, 10, 13A, 14, and 27, they lie directly on coarse gravel that presumably overlies glacial sediments. In Cores 7, 8, 11, 12, 16, 17, 21, 24, 30, and 30A, they directly overlie the yellowish-red sand and silt unit.

The lateral accretion (LA) unit is relatively coarse-grained, composed mostly of fine-to-medium sand (and some fine pebbles), and exhibits some cross-bedding. It also has interbedded silt layers as well as coarse detrital organic fragments and mollusk shells. The lateral accretion deposits are distinguished from the coarse channel-bed sediments by having finer grains, fewer and smaller pebbles, and usually more scattered detrital organic fragments. This unit is interpreted to have been deposited in point bars.

The vertical accretion (VA) sediments are generally massive silt with some fine sand scattered throughout or, more commonly, laminated and bedded silt and fine sand. They commonly contain thin but distinct organic laminae of partly decomposed leaf-litter. Impressions of leaves are preserved in some laminae. Mollusk shells are fairly common, either scattered or concentrated in layers. These relatively fine floodplain sediments are interpreted to be overbank deposits.

Palustrine Sediments (P in Fig. 4)

Most cores contain calcareous, light gray (10YR 6/2 to 7/2, 5YR 6/1 to 7/1, 5B 6/1) silty clay, usually within the upper 1–1.5 m (only Cores 13, 13A, 19, 20, 22, 22A, and 26 do not contain any of this sediment). It also has been detected with soil probes in an area about 250 m west-northwest of Core 23. It is interpreted to be palustrine (wetland) sediment. Depth of water probably varied, and the more clay-rich upper portions are interpreted to represent times when the water was consistently deeper. Because palustrine sediment is found in some of the highest cores (Cores 15 at 177.8 m (583 ft) and 21 at 178.2 m (585 ft)), it probably extended over much more of the area at one time but was eroded. Thickness of this sediment varies between about 0.5 to 1 m, and it overlies all floodplain deposits. The lower boundary of this unit in most cores is gradual, grading upward from overbank deposits, but the upper boundary is nearly everywhere abruptly unconformable with the sediments above it.

Particle-size analysis of selected samples shows silt to be the main component with nearly an equal amount of clay; the sand fraction is usually less than 10% (Kiesel 1998). The carbonate content ranges from 46 to 53%. This unit is sandier upvalley. Mollusk shells are very abundant in these sediments and occur either scattered throughout or in thin layers. The palustrine sediment in Cores 17 and 7 exhibit rhythmic layering with white laminae alternating with gray. Cores 5, 6, 10, 11, 12, 14, and 16 contain similar, but less well-developed white silt laminae. In many cores, the upper parts of this sediment are usually more clay-rich and are stained black, either from former soil development or from the modern soil.

Modern Overbank and Organic Layers (MO and O in Fig. 4)

Soil is usually present at the surface everywhere in the valley. Soil developed in modern overbank sediment overlies palustrine sediment or, in places where palustrine sediment does not exist, is developed in pre-modern floodplain sediment. Soil development in most places is weak or non-existent, indicating a surface that is actively eroding or aggrading. Buried soils are uncommon in the valley. In many cores, organic material darkens sediment at various depths which suggests a buried soil, but these areas are local and discontinuous and, thus, are interpreted to be organic-rich pockets in the premodern floodplain.

Beach Sand (BS in Fig. 4)

The cores nearest the lake (Cores 5, 6, 7, 8, and 9) contain beach sand as their highest stratigraphic unit. The sand is subrounded to subangular, mostly quartz, and generally medium-grained. Large, well-rounded pebbles are present in Cores 5 and 6. The sand and pebbles are similar to what exist on the present-day beach. Cores 5, 6, and 9 have a thin, weakly developed soil on the beach sand. The soil is virtually nonexistent in Cores 7 and 8 because beach sand too recently invaded those sites. These cores were taken behind the present-day beach where washovers encroach into the wetland. Here the beach sand overlies the soil developed in the floodplain sediments.

CHRONOLOGY AND PALEOENVIRONMENTAL INTERPRETATION

Radiocarbon (^{14}C) Dating

All ages in this study are given as ^{14}C years before present (BP). Of the 30 age determinations (Table 2) from this relatively small area, 18 are useful (*i.e.*, not modern, abnormally old, or inverted/out-of-place). Eleven of those fall within the 5,000–6,000 ^{14}C years BP range and the rest are younger. Reworking of sediment is the apparent cause of inverted dates in Cores 6, 17, and 28. With the influence of the riverine system, this is not unexpected. Sediment in other cores could also have been reworked, but they are not as obviously out of place with respect to surrounding dates.

There are no dates from about 5,000 to about 2,000 ^{14}C years BP. Stratigraphically, the palustrine sediments should have ages that fall within this gap, but the palustrine ages appear to be too old (6,000–7,000 ^{14}C years BP—Cores 16, 17, and 29), and they are consistently older than the overbank sediments above which they lie (5,000–6,000 ^{14}C years BP). Because both the palustrine and overbank sediments have high carbonate contents, the older ages of the palustrine unit are probably due to the existence of a different plant community, one that included many submergent plants (Keough *et al.* 1999). Unlike emergent plants that get carbon from the atmosphere, submergent plants obtain a higher proportion of their carbon from carbonate dissolved from the regional dolomitic bedrock. Because the rock contains no measurable ^{14}C when organisms use that carbon for photosynthesis and shell construction, the ratio of ^{14}C to ^{12}C is lower than if all the carbon came from the atmosphere, resulting in abnormally older apparent ages (Taylor 1987, DeNiro 1983). In some cases where the fraction of carbonate contamination from bedrock has been determined, age corrections have been made (*e.g.*, Farrand and Miller 1968, Deevey *et al.* 1954, Olson 1963). Because the amount of carbonate contamination was not determined in the palustrine sediments of the Pigeon River, no age corrections are possible.

TABLE 2. Radiocarbon dates and associated information. M = modern. Beta = Beta Analytic, Inc.; AA = NSF Arizona AMS Facility (with graphitization done at the UW-Madison Radiocarbon Lab); CAMS# = Lawrence Livermore (with graphitization done at the USGS Reston Lab).

Core	Depth m	Elev m	AGE (^{14}CBP)	Material	Lab No.	Unit Interpretation
4	2.50	175.80	5430 ± 70	wood	Beta-77074	vertical accretion
5	0.53	176.77	1800 ± 50	peaty matter	AA25949	vertical accretion
5	1.49	175.81	5200 ± 60	wood	CAMS#25713	lateral accretion
6	0.23	177.00	6760 ± 55	org. sed	AA22419	vertical accretion
6	0.50	176.73	805 ± 70	woody matter	AA21148	palustrine
6	1.89	175.34	5455 ± 60	twig	AA21546	vertical accretion
7	0.31	176.98	935 ± 50	peaty matter	AA22420	vertical accretion
7	4.28	173.01	6360 ± 60	wood	Beta-80196	older fluvial
8	0.10	177.17	1.205 ± .01%= M	twig	AA21150	vertical accretion
8	0.44	176.83	1160 ± 40	peaty matter	AA22417	vertical accretion
8	1.83	175.44	5910 ± 55	bulk sed.	AA25954	organics above older fluvial
9	2.55	174.65	9210 ± 90	wood	Beta-77075	vertical accretion
11	1.20	176.05	5290 ± 60	wood	CAMS#25714	vertical accretion
11	3.69	173.56	5720 ± 60	wood	CAMS#25715	older fluvial
15	1.52	177.02	1.219 ± .022%= M	charcoal	AA21151	gravel
16	1.16	176.08	6840 ± 55	leaf litter	AA22421	vertical accretion
17	0.75	176.55	6140 ± 55	bulk sed.	AA25950	palustrine
17	1.16	176.14	6610 ± 60	bulk sed.	AA25951	palustrine
17	1.45	175.85	5300 ± 90	leaf litter	AA21149	vertical accretion
17	3.19	174.11	5490 ± 70	wood	Beta-80197	vertical accretion
20	1.27	176.10	2760 ± 70	wood	Beta-77076	lateral accretion
20	2.94	174.43	5730 ± 60	wood	CAMS#25716	channel bed
21	0.39	178.17	1.222 ± 0.9%= M	wood	Beta-80198	vertical accretion
21	2.79	175.77	5370 ± 60	twig	AA21547	older fluvial
22A	0.52	176.79	2260 ± 60	wood	CAMS#25717	channel bed
24	2.70	174.68	6450 ± 70	bulk sed.	AA25953	vertical accretion
28	0.75	176.62	1260 ± 60	woody matter	AA21549	vertical accretion
28	1.46	175.91	880 ± 50	wood	AA21550	channel bed
28	2.58	174.79	5770 ± 60	wood	AA21548	channel bed
29	0.65	177.22	7060 ± 120	bulk sed	AA25952	palustrine

The Chronological and Paleoenvironmental Interpretation

The sediment cores reveal a very complex Holocene history (Fig. 5, Table 3). Reddish-brown, clay-rich diamicton and massive silt and clay are interpreted to be glacial till and glaciolacustrine sediment, respectively. No ages for glacial sediment were obtained, but Valders ice (final deglaciation) was out of Manitowoc County just north of Sheboygan County by about 12,200 years ago (Maher and Mickelson 1996).

Coarse gravel overlies glacial sediment in all cores that contain till or glacial lake sediment and occupies the lowest position in many other cores, where it is presumed to overlie glacial sediment at depth. This lowest channel-bed unit decreases in elevation in a down-valley direction (Fig. 4). Ages of sediment higher in the cores constrain this gravel to be older than 6,360 ± 60 ^{14}C years BP. It probably was deposited by the river as it flowed into Lake Michigan when it was substantially lower (Chippewa low stand). Knox and Leigh (1987) made a similar interpretation of the basal gravel in the Onion and Sheboygan river valleys.

The yellowish-red fluvial sand and silt unit overlies the lowest channel bed unit in the northern half of the valley. The most likely origin of the yellowish-red sand and silt unit is fluvial, with most of the sediment being contributed from the north bluff of the valley, which contains large lenses of glaciolacustrine sand and yellowish-red silt and clay (Bay Lakes Regional Planning Commission, 1997). Ages of material above and below this unit suggest that this fluvial activity occurred between about 6,360 ± 60 and 5,370 ± 60 ^{14}C years BP. However, ages of material above and below the unit

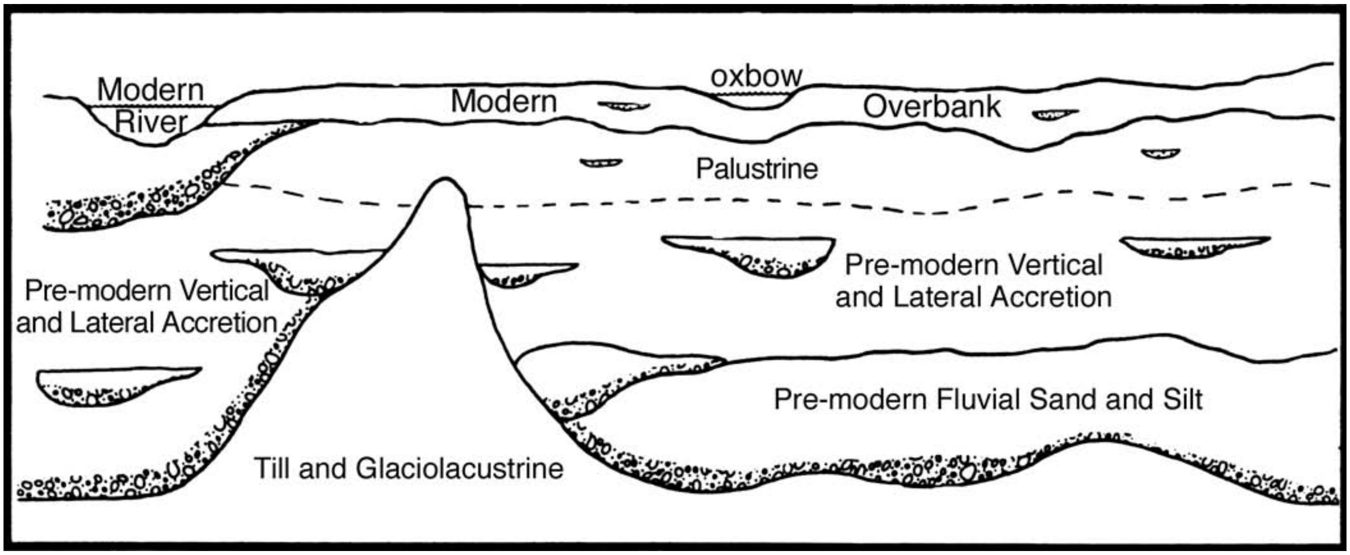

FIG. 5. Sketch illustrating the Holocene stratigraphy of the mouth-area of the Pigeon River. Coarse pattern indicates coarse channel-bed units. Pre-modern Fluvial refers to the yellowish-red sand and silt unit. Vertical and lateral accretion units are the fluvial sediments. Isolated pockets indicate meander scar cross-sections.

vary considerably. An age of 6,450 ± 70 ^{14}C years BP (Core 24) in what appears to be an organic-rich zone (soil?) grading upward from the yellowish-red unit suggests the unit is older than that.

Overlying the yellowish-red sand and silt unit is a widespread, 2–3 m thick, unit of lateral and vertical accretion deposits and associated channel-bed sediments. Ages ranging from 5,490 ± 70 ^{14}C years BP (bottom of floodplain sediments in Core 17) to 5,200 ± 60 ^{14}C years BP (top of floodplain sediments in Core 5) are stratigraphically consistent in these sediments everywhere in the valley. These floodplain sediments were probably deposited as water rose from the Chippewa low stand during the Nipissing transgression. Point-bar sediments are coarser than those on the modern river bed, and floodplain deposits of this unit are coarser than those of the modern floodplain, suggesting a higher energy stream during mid-Holocene than today.

Grading upward from mid-Holocene floodplain deposits are gray palustrine sediments interpreted to represent a rise of lake level to present levels and above, creating marshy (*i.e.*, wetter, standing water) conditions on the floodplain. All ages obtained for these sediments are older than is consistent with their stratigraphic positions, presumably due to carbonate contamination. In most cores, palustrine sediments grade upward into more clay-rich sediments, indicating deepening water. Depth of water is indeterminable from the sediment record, as is the length of time water flooded the river mouth. However, when the water in the valley was at its maximum level (184.4 m, 605 ft) during the Nipissing phase, it should have been about 7 m deep in the study area. Because an unconformity exists everywhere above the clay-rich palustrine sediment, the total amount of sediment deposited in this estuary and subsequently eroded is unknown.

This extensive unconformity and lack of sediment containing dates between about 5,000 and 2,000 ^{14}C years BP in the valley mouth-area suggest that this was a time of erosion. During this

TABLE 3. Summary of depositional chronology.

Unit/Event	Time
modern floodplain	2,000 ^{14}C yrs BP to present
erosion	5,000 to 2,000 ^{14}C yrs BP
floodplain and palustrine units	mid-5,000 to 5,000 ^{14}C yrs BP
yellowish-red fluvial sand and silt unit	mid-6,000 to mid-5,000 ^{14}C yrs BP
lowest channel bed unit (above glacial material)	prior to mid-6,000 ^{14}C yrs BP
till and glaciolacustrine sediments	oldest, but no dates

time, water level generally fell but with large fluctuations (Baedke and Thompson 2000, Larsen 1985, Dott and Mickelson 1995). According to Knox (1988, 1999) and Brakenridge (1980), there were significant periods of large floods in the Midwest during the past 3,000 years, and these, combined with declining lake level after the Nipissing high stand, caused significant erosion. Climatic interpretations from pollen, inland lake levels and varves, and fluvial activity indicate periods of cooler and more moist climate with episodes of extensive erosion in the upper Midwest especially during the latter half of this time (Knox 1988 and 1999, Dean *et al.* 1984, Winkler *et al.* 1986, Winkler 1988, Harrison 1989).

One possible explanation for not finding evidence of minor lake-level fluctuations in the present-day Pigeon River is that, until recently, the river mouth was farther to the east under present-day Lake Michigan. Chrzastowski and Thompson (1994) suggest that during the Chippewa phase, stream incision occurred well out into Lake Michigan, and paleochannels have been found in what was a lacustrine plain now under the lake. Recent studies of lakeshore bluff erosion (Bay Lakes Regional Planning Commission 1997) show rapid bluff-recession rates along this part of the Lake Michigan shoreline during the past 20 years. Although the erosion rates would have been highly variable during the past 5,000 years, depending on the level of the lake at any given time, if the rates in this study are averaged together (conservatively, about 0.3 m/year) and extrapolated into the past, the shoreline location during the Nipissing rise could have been as much as a mile further out into the lake 5,000 years ago. Even if this distance is too large, it is clear that the location of the present-day mouth was probably well inland of the open lake. For whatever reason, erosion dominated this area until about 2,000 years ago.

The next event for which there is evidence is deposition of more vertical accretion floodplain sediments over the unconformity. The modern soil has developed in these sediments. Much younger ages (ranging in age from 1,800 ± 50 to 805 ± 70 ^{14}C years BP) in Cores 5, 6, 7, 8, and 28 suggest that a floodplain environment dominated the area for about the last 2,000 years. This floodplain is interpreted to have been created by a low-energy, meandering river because of numerous oxbow scars and because all of the sediments are relatively fine-grained. Deposits are much thinner than older floodplain deposits, and no coarse channel-bed deposits exist. These factors all suggest the valley probably has been slowly aggrading in response to subtle lake-level fluctuations or perhaps, most recently, to post-settlement alluvium.

The most recent event in the valley mouth area has been the deposition of modern beach sand over the floodplain soil in the area nearest the present-day shoreline. This beach sand does not extend inland, indicating lake level has not risen high enough for a long enough period of time in recent years to create sand deposits farther inland.

CONCLUSIONS

Sediments of the Pigeon River mouth area record evidence of major changes in Lake Michigan water levels during the Holocene. Coarse gravel over glacial and glaciolacustrine sediment is interpreted to have been deposited by the river flowing into a lower level of Lake Michigan. Yellowish-red sand and silt were deposited as lake level began to rise. Continued rise to the Nipissing level is recorded by thick point bar and floodplain sediments. These grade upward into palustrine sediment which becomes more clay-rich near the surface, suggesting that standing water eventually occupied the valley. An unconformity and a gap in the radiocarbon ages at the upper boundary of the palustrine sediments indicate that much of the sediment that accumulated in the valley during the Nipissing high stand was eroded away.

Younger floodplain sediment in which the present-day soil is formed suggests that aggradation has occurred in the valley for at least the past 1,800 years. This may have occurred quickly since European settlement or more slowly during the last 2,000 years. There appears to be no evidence in the study area of minor variations in lake level significant enough to have left a record in the sediments.

Although the sediments along the lower Pigeon River provide some evidence for major lake-level events, the resolution of inferences that can be drawn from these deposits appears to be less than that available from beach-ridge studies.

ACKNOWLEDGMENTS

The authors gratefully acknowledge the vibracoring and surveying help of dozens of field assistants from the UW-Madison, UW-Baraboo/Sauk County, and Wisconsin DNR over the course of three summers. Particle-size analyses and Chittick analyses were done by the UW-Madison Quaternary Lab, and graphite targets were made under the direction of David McJunkin at the UW-Madison Radiocarbon Lab. Discussions with Robert Dott, Jr., Curt Larsen, Louis Maher, Jr., Toni Simo, Allan Carroll, and James Knox were greatly appreciated. Special thanks are due Mary Diman who created or modified the figures. Harvey Thorleifson, Ron Stieglitz, and an anonymous referee made helpful contributions to the manuscript. The following grants provided financial help for this project: The Geological Society of America Summer Research Grants (#4967-92 and #5634-95), the UW-Madison Department of Geology and Geophysics (three Exxon summer grants #133-S317) and Sigma Xi. The University of Arizona NSF-Arizona AMS Facility provided 19 AMS dates at a very reasonable cost. Five other AMS dates were provided by USGS Great Lakes Research projects.

REFERENCES

Acomb, L.J., Mickelson, D.M., and Evenson, E.B. 1982. Till stratigraphy and late glacial events in the Lake Michigan lobe of eastern Wisconsin. *Geol. Soc. Amer. Bull.* 93:289–296.

Alden, W.C. 1918. *The Quaternary geology of southeastern Wisconsin.* U.S.G.S. Prof. Paper 106.

Anderson, T.W. 1979. *Stratigraphy, age, and environment of a Lake Algonquin embayment site at Kincardine, Ontario*. Geol. Surv. Can. Paper 79-1B: 147–152.

Baedke, S.J., and Thompson, T.A. 2000. A 4700-year record of lake level and isostacy for Lake Michigan. *J. Great Lakes Res.* 26:416–426.

Barnett, P.J., Coakley, J.P., Terasmae, J., and Winn, C.E. 1985. Chronology and significance of a Holocene sedimentary profile from Clear Creek, Lake Erie shoreline, Ontario. *Can. J. Earth Sci.* 22:1133–1138.

Bay, J.W. 1937. Glacial-lake levels indicated by terraces of the Huron, Rouge, and Clinton Rivers, Michigan. *Mich. Acad. of Arts, Sciences, and Letters Papers* 22: 411–419.

Bay-Lakes Regional Planning Commission, 1997. *Lake Michigan shoreline recession and bluff stability in northeastern Wisconsin: 1996*. preliminary draft of Technical Report No. 51.

Bettis, A.E., III, Baker, R.G., Green, W.R., Whelan, M.K., and Benn, D.W. 1992. *Late Wisconsinan and Holocene alluvial stratigraphy, paleoecology, and archaeological geology of east-central Iowa.* Guidebook Series 12, Iowa Dept. of Natural Resources.

Booth, R.K., Jackson, S.T., and Thompson, T.A., 2002, Paleoecology of a northern Michigan lake and the relationship among climate, vegetation, and Great Lakes water levels. *Quaternary Research* 57: 120–130.

Brakenridge, G. R. 1980. Widespread episodes of stream erosion during the Holocene and their climatic cause. *Nature* 283:655–656.

Cant, D.J. 1982. Fluvial facies models and their application. In *Sandstone Depositional Environments*, eds. P.A. Scholle and D. Spearing, pp. 115–137. Assoc. of Petrol. Geol.

Chapel, D.M. 2000, The Quaternary geology of central Sheboygan County. MS thesis, University of Wisconsin-Madison.

Chrzastowski, M.J., and Thompson, T.A. 1994. Late Wisconsinan and Holocene geologic history of the Illinois-Indiana coast of Lake Michigan. *J. Great Lakes Res.* 20:9–26.

Clark, J.A., Pranger, H.S., II, Walsh, J.K., and Primus, J.A. 1990. A numerical model of glacial isostasy in the Lake Michigan basin. In *Late Quaternary History of the Lake Michigan Basin*, eds. A.F. Schneider and G.S. Fraser, pp. 111–123. Geol. Soc. Amer. Special Paper 251.

Coakley, J.P., and Lewis, C.F.M. 1985. Postglacial lake levels in the Erie basin. In *Quaternary Evolution of the Great Lakes*, eds. P.F. Karrow and P.E. Calkin, pp. 195–212. Geol. Assoc. Canada Special Paper 30.

Colman, S.M., Forester, R.M., Reynolds, R.L., Sweetkind, D.S., King, J.W., Gangemi, P., Jones, G.A., Keigwin, L.D., and Foster, D.S. 1994. Lake-level history of Lake Michigan for the past 12,000 years: the record from deep lacustrine sediments. *J. Great Lakes Res.* 20:73–92.

Coordinating Committee on Great Lakes Basic Hydraulic and Hydrologic Data, 1977. *Apparent vertical movement over the Great Lakes*. Detroit, Detroit District, U.S. Army Corps of Engineers.

Cowan, W.R. 1978. Radiocarbon dating of Nipissing Great Lakes events near Sault Ste. Marie, Ontario. *Can. J. Earth Sci.*15:2026–2030.

Dalrymple, R.W., and Carey, J.S. 1990. Water-level fluctuations in Lake Ontario over the last 4000 years as recorded in the Cataraqui River lagoon, Kingston, Ontario. *Can. J. Earth Sci.* 27:1330–1338.

Dean, W.A., Bradbury, J.P., Anderson, R.Y., and Barnosky, C.W. 1984, The variability of Holocene climate change: evidence from varved sediments. *Science* 226:1191–1194.

Deevey, E.S., Jr., Marsha, S.G., Hutchinson, G.E., and Kraybill, H.L. 1954. The natural C^{14} contents of materials from hard-water lakes. *Proc. Natl. Academy Sci.* 40:285–288.

DeNiro, M.J. 1983. Distribution of the stable isotopes of carbon, nitrogen, oxygen, and hydrogen among plants. In *Organic Geochemistry of Contemporaneous and Ancient Sediments*, ed. W.G. Meinschein, pp. 4-1–4-27.

Dott, E.R., and Mickelson, D.M. 1995. Lake Michigan levels and the development of Holocene beach ridge complexes at Two Rivers, Wisconsin. *Geol. Soc. Amer. Bull.* 107:286–296.

Dreimanis, A. 1962. Quantitative gasometric determination of calcite and dolomite by using Chittick apparatus. *J. Sed. Petrol.* 32:520–529.

Farrand, W.R., and Miller, B.B. 1968. Radiocarbon dates on and depositional environment of the Wasaga Beach (Ontario) marl. *Ohio J. Sci.* 68 (4):235–239.

Fisher, T.G., Loope, W., and Fuller, A.P., 2001, A Nipissing transgression record from dunedammed Silver Lake, Little Sable Point, Lake Michigan. North-Central Geological Society of America *Abstracts with Program* 33, no. 4:11.

Flint, J.E., Dalrymple, R.W., and Flint, J.J. 1988. Stratigraphy of the Sixteen Mile Creek lagoon and its implications for Lake Ontario water levels. *Can. J. Earth Sci.* 25:1175–1183.

Folger, D.W., Colman, S.M., and Barnes, P.W. 1994. Overview of the southern Lake Michigan coastal erosion study. *J. Great Lakes Res.* 20:2–8.

Fraser, G.S., Larsen, C.E., and Hester, N.C. 1990. Climatic control of lake levels in the Lake Michigan and Lake Huron basins. In *Late Quaternary History of the Lake Michigan Basin*, eds. A.F. Schneider and G.S. Fraser, pp. 75–89. Geol. Soc. Amer. Special Paper 251.

Hansel, A.K., and Mickelson, D.M. 1988. A reevaluation of the timing and causes of high lake phases in the Lake Michigan basin. *Quat. Res.* 29:113–128.

_____, Mickelson, D.M., Schneider, A.F., and Larsen, C.E. 1985. Late Wisconsin andHolocene history of the Lake Michigan basin. In *Quaternary Evolution of the Great Lakes,* eds. P.F. Karrow and P.E. Calkin, pp. 39–53. Geol. Assoc. of Canada Special Paper 30.

_____, Chrzastowski, M.J., Riggs, M.H., Miller, M.V., and Follmer, L.R. 1990. C-14 ages on Late Holocene peat accumulations in beach-ridge swales at Illinois Beach State Park (abst). In *2nd Annual Southern Lake Michigan Coastal Erosion Study Workshop*, eds. D.W. Folger, S.M. Colman, and P.W. Barnes, p. 4. U.S.G.S. Open File Report 90-272.

Harrison, S.P. 1989. Lake levels and climatic change in eastern North America. *Climate Dynamics* 3: 157–167.

Johnson, T.C., Stieglitz, R.D., and Swain, A.M. 1990. Age and paleoclimatic significance of Lake Michigan beach ridges at Baileys Harbor, Wisconsin. In *Late Quaternary History of the Lake Michigan Basin*, eds. A.F. Schneider and G.S. Fraser, pp. 67–74. Geol. Soc. Amer. Special Paper 251.

Johnston, J.W., and Thompson, T.A. 2000, Strandplain evidence for Late Holocene lake level in Tahquamenon Bay, southeastern Lake Superior. North-Central Geological Society of America, *Abstracts with Program* 32, no 4:19–20.

Karrow, P.F. 1986. Valley terraces and Huron basin water levels, southwestern Ontario. *Geol. Soc. Amer. Bull.* 97:1089–1097.

Keough, J.R., Thompson, T.A., Guntenspergen, G.R., and Wilcox, D.A. 1999. Hydrogeomorphic factors and ecosystem responses in coastal wetlands of the Great Lakes. *Wetlands* 19:821–834.

Kiesel, D.S. 1998. Holocene stratigraphy of the river-mouth sediments of the Pigeon River, eastern Wisconsin: implications for lake-level fluctuations. Ph.D. dissertation, Univ. of Wisconsin, Madison, WI.

Klotz, J.A. 1981. Nature and origin of the Maumee River terraces, northwestern Ohio. M.S. thesis, Bowling Green State Univ., Bowling Green, OH.

Knox, J.C. 1988. Climatic influence on upper Mississippi valley floods. In *Flood Geomorphology*, eds. V.R. Baker, R.C. Kochel, and P.C. Patten, pp. 279–300. New York: Wiley.

_____. 1999. Long-term episodic changes in magnitudes and frequencies of floods in the Upper Mississippi River Valley. In *Fluvial Processes and Environmental Change*, eds A.G. Brown and T.A. Quine, pp. 255–282. Chichester: Wiley.

_____, and Leigh, D.S. 1987. *Late Wisconsin and Holocene geomorphic history of the Lower Onion River Valley, Sheboygan County, Wisconsin.* Unpublished report, Dept. of Geography, University of Wisconsin, Madison, WI.

Larsen, C.E. 1985. *A stratigraphic study of beach features on the southwestern shore of Lake Michigan: new evidence of Holocene lake level fluctuations.* Illinois State Geol. Surv., Environmental Geology Notes 112.

_____. 1994. Beach ridges as monitors of isostatic uplift in the upper Great Lakes. *J. Great Lakes Res.* 20: 108–134.

Larson, G., and Schaetzl, R. 2001. Origin and evolution of the Great Lakes. *J. Great Lakes Res.* 27:518–546.

Lichter, J. 1995. Lake Michigan beach-ridge and dune development, lake level, and variability in regional water balance. *Quat. Res.* 44:181–189.

_____. 1997. AMS radiocarbon dating of Lake Michigan beach-ridge and dune development. *Quat. Res.* 48: 137–140.

Maher, L.J., Jr., and Mickelson, D.M. 1996. Palynological and radiocarbon evidence for deglaciation events in the Green Bay lobe, Wisconsin. *Quat. Res.* 46: 251–259.

McDowell, P.F. 1980. Holocene Fluvial Activity in Brush Creek Watershed, Wisconsin. Ph.D. dissertation, Univ. of Wisconsin, Madison, WI.

_____. 1983. Evidence of stream response to Holocene climatic change in a small Wisconsin watershed. *Quat. Res.* 19:100–116.

Olson, E.A. 1963. The problem of sample contamination in radiocarbon dating. Ph.D. dissertation, Columbia University,

Petty, W.H., Delcourt, P.A., and Delcourt, H.R. 1996. Holocene lake-level fluctuations and beach-ridge development along the northern shore of Lake Michigan, USA. *J. Paleolimnology* 15:147–169.

Platt, N.H., and Wright, V.P. 1992. Palustrine carbonates and the Florida Everglades: towards an exposure index for the fresh-water environment? *J. Sed. Petrol.* 62:1058–1071.

Principato, S.M. 1999, The Quaternary geology of northern Sheboygan County. M.S. thesis, University of Wisconin-Madison.

Richardson, D.B. 1991. River valley terraces of the lower Onion and lower Sheboygan rivers, Sheboygan County, Wisconsin. M.S. thesis, Univ. of Wisconsin, Madison, WI.

Schneider, A.F., and Need, E.A. 1985. Lake Milwaukee: an "early" proglacial lake in the Lake Michigan Basin. In *Quaternary Evolution of the Great Lakes*, eds P.F. Karrrow and P.E. Calkins, pp. 55–62. Geological Association of Canada Special Paper 30.

Smith, D.G. 1984. Vibracoring fluvial and deltaic sediments: tips on improving penetration and recovery. *J. Sed. Petrol.* 54:660–663.

Taylor, R.E. 1987. *Radiocarbon Dating: An Archeological Perspective.* Orlando, FL: Academic Press.

Thompson, T.A. 1992. Beach-ridge development and lake-level variation in southern Lake Michigan. *Sed. Geol*. 80:305–318.

Tushingham, A.M., and Peltier, W.R. 1991. ICE-3G: a new global model of late Pleistocene deglaciation based upon geophysical predictions of post-glacial relative sea level change. *J. Geophys. Res.* 96: 4497–4523.

Wilkinson, B.H., and Fuks, K.H., 2001. Holocene sedimentation in two western Michigan estuaries. North-Central Geological Society of America, *Abstracts with Program* 33, 4:12.

Winkler, M.G. 1988. Effect of climate on development of two Sphagnum bogs in south-central Wisconsin. *Ecology* 69:1032–1043.

_____, Swain, A.M., and Kutzbach, J.E. 1986. Middle Holocene dry period in the northern Midwestern United States: lake levels and pollen stratigraphy. *Quat. Res.* 25:235–250.

Submitted: 18 February 2005
Accepted: 4 June 2005
Editorial handling: Harvey Thorleifson