Guide Book for the 53rd Midwest Friends of the Pleistocene Field Conference

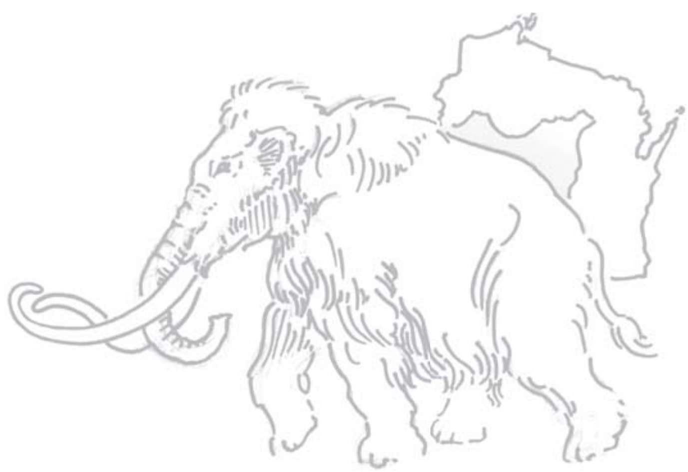
May 18–20, 2007 • Oshkosh, Wisconsin

Late-Glacial History of East-Central Wisconsin

2007

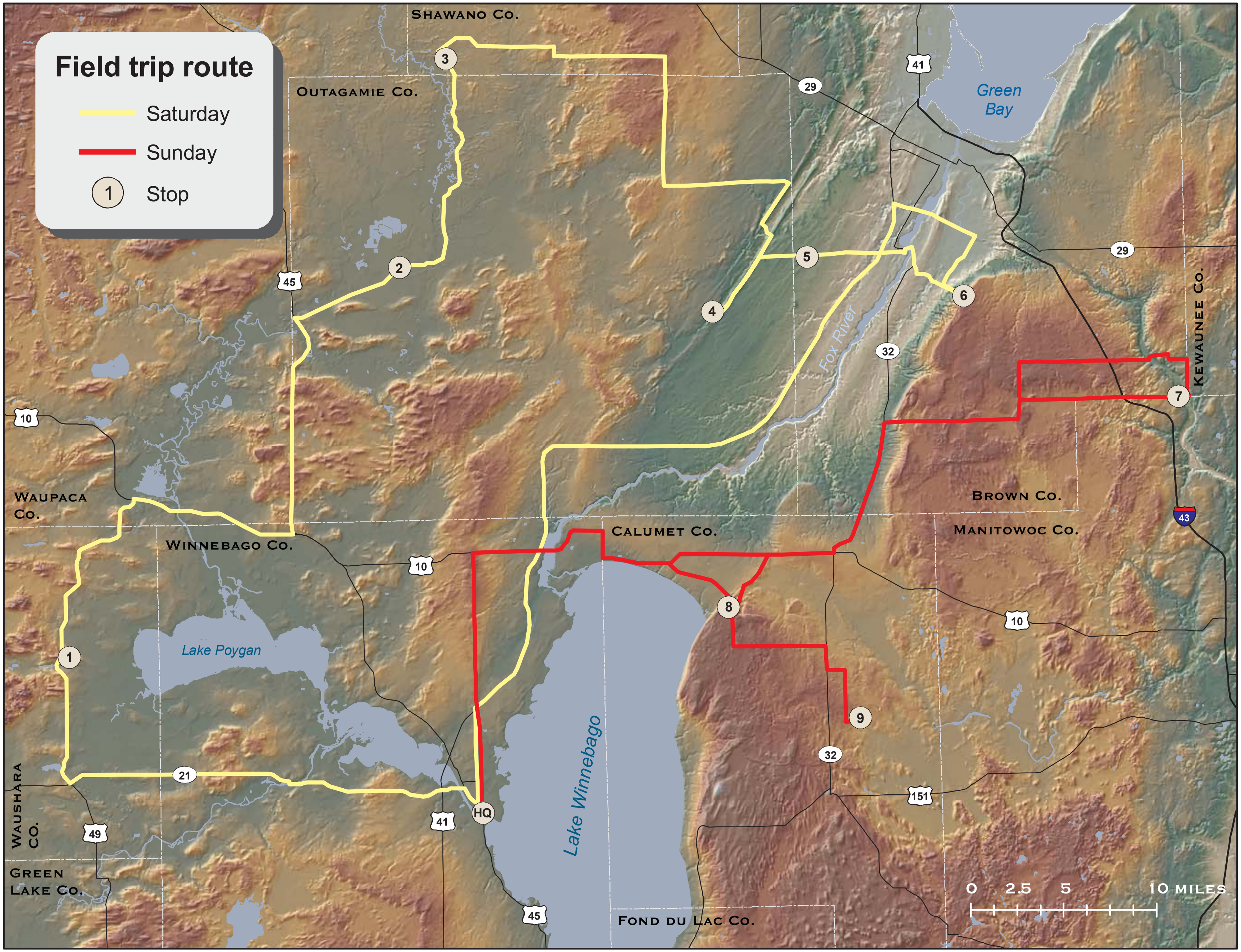
edited by

Thomas S. Hooyer



Wisconsin Geological and Natural History Survey

Open-File Report 2007-01



Field trip route
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Winnebago Co.
Calumet Co.
Brown Co.
Manitowoc Co.
Kewaunee Co.
Waushara Co.
Green Lake Co.
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with contributions from

William G. Batten
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Stephen L. Forman
Holly Gertz
Jacquelyn Gill
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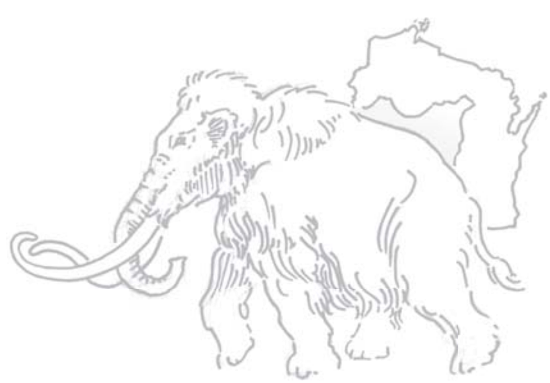

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Evolution of glacial Lake Oshkosh and the Fox River lowland

Introduction

Profoundly affected by glaciation, the landscape of east-central Wisconsin is dominated primarily by undulating glacial lake plains and landforms, such as moraines, drumlins, deltas, and sand dunes.

During the most recent glacial maximum, approximately 25,000 years ago, the southern margin of the Laurentide Ice Sheet extended southward into Wisconsin. The margin appears to have been irregular, with four lobes crossing the Superior basin directly from the north (the Superior, Chippewa, Wisconsin Valley, and Langlade Lobes) and two other lobes, the Green Bay and Lake Michigan Lobes, farther to the northeast (fig. 1). Lakes formed in front of the ice when established surface-water drainage routes were blocked (Thwaites, 1943). As the Green Bay Lobe advanced into the northern Michigan basin and Green Bay area, it blocked the northward drainage of the ancestral Fox River. A lake that formed in front of the ice margin, glacial Lake Oshkosh, occupied a region referred to either as the Fox River lowland or the Oshkosh basin (fig. 2). This lowland covers a linear region approximately 160 km long and 50 km wide that extends southwest from the city of Green Bay and includes parts of 14 counties. This area is topographically one of the lowest regions in the state, and it is the focus of a long-term mapping effort of the Wisconsin Geological and Natural History Survey (fig. 2).

Glacial Lake Oshkosh must have existed during the initial advance of the ice lobe into the basin about 30,000 years ago. However, little to no evidence remains for this early lake phase, which indicates that the ice lobe effectively eroded its substrate, including the surface of the underlying sandstone, shale, and dolomite bedrock. As the climate warmed, the Green Bay Lobe started to recede, and glacial Lake Oshkosh again formed between the ice margin and the outermost moraine. The lake persisted in front of the Green Bay Lobe until it eventually receded northward out of Wisconsin (Wielert, 1979).

Figure 1. Major landscape regions and extent of glaciation in Wisconsin. Note that the division of the Green Bay and Lake Michigan Lobes coincides with the Silurian escarpment.

Previous work

There are still many unknowns regarding the glacial history of the late Pleistocene, even though the geology of east-central Wisconsin has been studied for almost 150 years. The presence of a large proglacial lake in the Fox River valley in east-central Wisconsin has been recognized since the early field work of Whittlesey (1849) and a surficial map produced by Warren (1876). Numerous observations relevant to the geol-

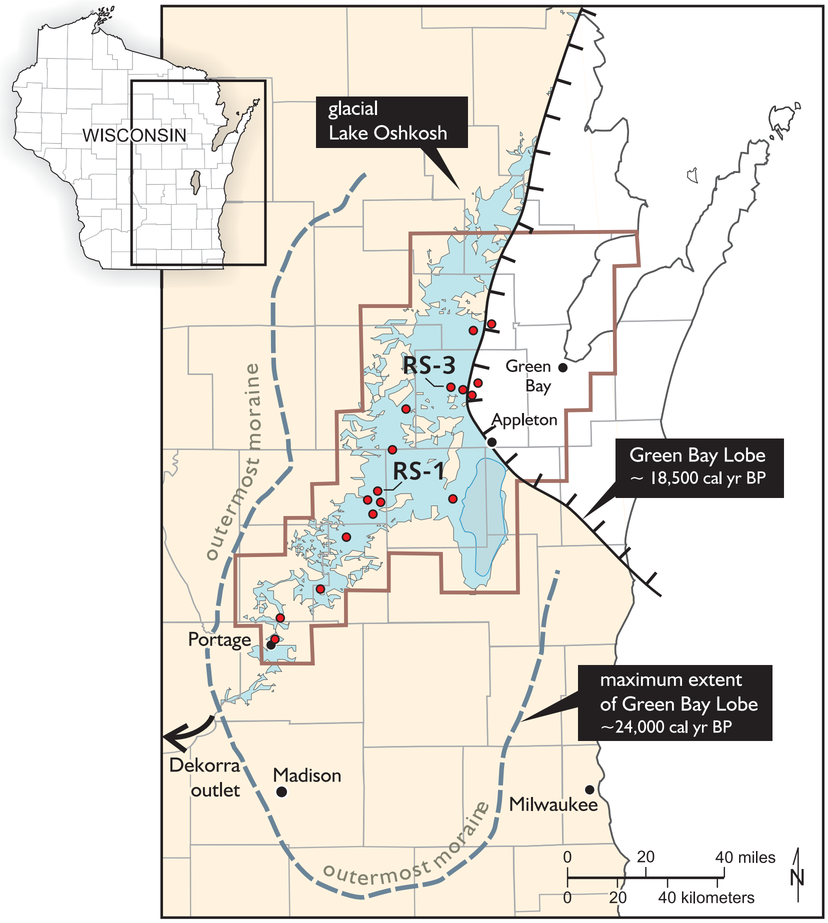

Figure 2. Map of east-central Wisconsin showing the location of glacial Lake Oshkosh and rotosonic boreholes (red circles) recently drilled. The solid line represents Fox River lowland, the area currently being mapped by the Wisconsin Geological and Natural History Survey.

ogy of the region were subsequently made by Chamberlin (1883), Upham (1903), and Weidman (1911). Alden (1918) published a map of the surficial deposits of southeastern Wisconsin that covered the southern half of the basin. Thwaites (1943) mapped the central basin and produced a manuscript that has provided the basis for our understanding of the Pleistocene geology of east-central Wisconsin. Since this work, no modern surficial maps in the basin have been made, with the exception of Brown County (Need, 1985). More recent work in the region has focused on understanding the glacial history of the area (McCartney, 1979; McCartney and Mickelson, 1982) and of glacial Lake Oshkosh (Wielert, 1979).

Glacial history of the Fox River lowland

The extent of glacial Lake Oshkosh was controlled not only by the location of the ice margin, but by the elevation of its outlets. When the margin of the Green Bay Lobe initially receded northward, the lake drained southward into the lower Wisconsin River by way of the Dekorra outlet just south of Portage (fig. 3A). Once the lobe advanced northward of the present location of the city of Oshkosh, a series of four lower outlets (Manitowoc, Neshota, Kewaunee, and Ahnapee, fig. 3B–F) opened to the east across the Silurian escarpment, which extends into the Door Peninsula; drainage through the outlets eventually lowered glacial Lake Oshkosh to the level of the water in the Michigan basin (fig. 3F). The configuration of the lake changed each time the ice receded far enough to uncover a lower outlet. A topographic profile along the drainage divide of the Silurian escarpment shows a transverse cut of the channels downstream from the outlets (fig. 3G). These channels were most likely incised by many discharge events from the Oshkosh basin to the Michigan basin.

Geologic evidence in the Oshkosh basin has indicated that the Green Bay Lobe readvanced at least twice after it had fully receded from Wisconsin; the second readvance was of lesser extent than the first (fig. 3H). During each readvance, the eastern outlets were subsequently reactivated and covered, leaving the lake to drain out of the southern outlet at Dekorra. With each recession, the outlets were used again, but in reverse order.

The complex history of the Fox River lowland associated with the recession and read-

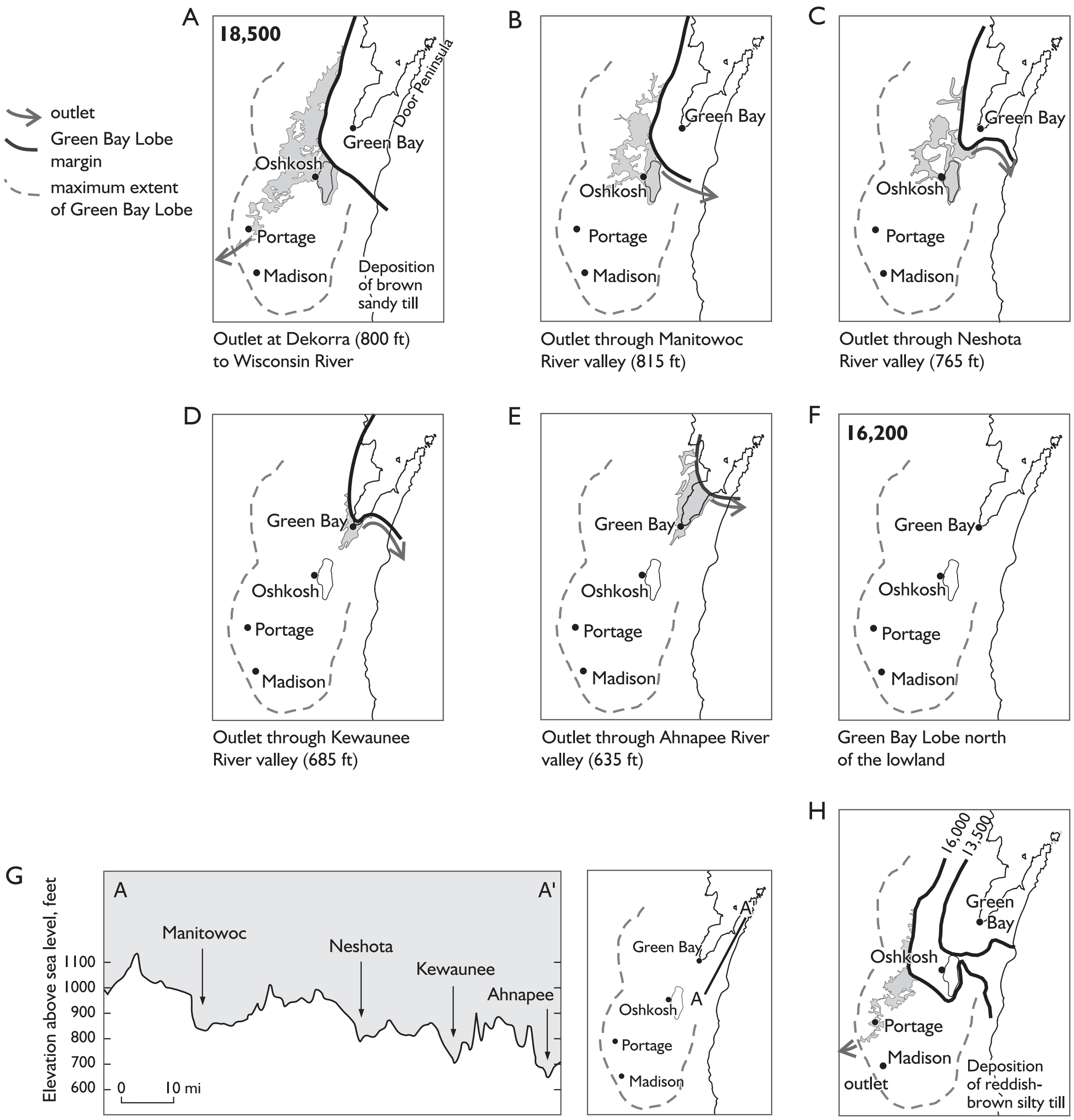

Figure 3. Glacial Lake Oshkosh formed any time the Green Bay Lobe advanced or receded into the Fox River lowland, blocking northward surface-water drainage. Upon the recession of the ice lobe, the lake initially drained southward through the Dekorra outlet to the ancestral lower Wisconsin River **(A).** Eventually, a series of four lower outlets opened eastward to the Michigan basin, lowering the level of the lake **(B–E)** until it was at the same level as water in the Michigan basin **(F).** Discharge through the outlets cut channels across the Silurian escarpment **(G).** The lobe had two major readvances to the central part of the basin approximately 16,000 and 13,500 cal yr BP, resulting in reuse of the outlets **(H).**

vances of the Green Bay Lobe is summarized in figure 4. At least four major glacial phases have been identified; they include the minor readvances/standstills associated with the Horicon Member of the Holy Hill Formation. This member, which consists primarily of sandy brown till with dolomite clasts, commonly covers the bedrock surface outside the basin. In particular, this till forms the large upland drumlin field south of the lake basin (Colgan and Mickelson, 1997). Although the topography of the basin prior to the advance of the ice sheet is unknown,

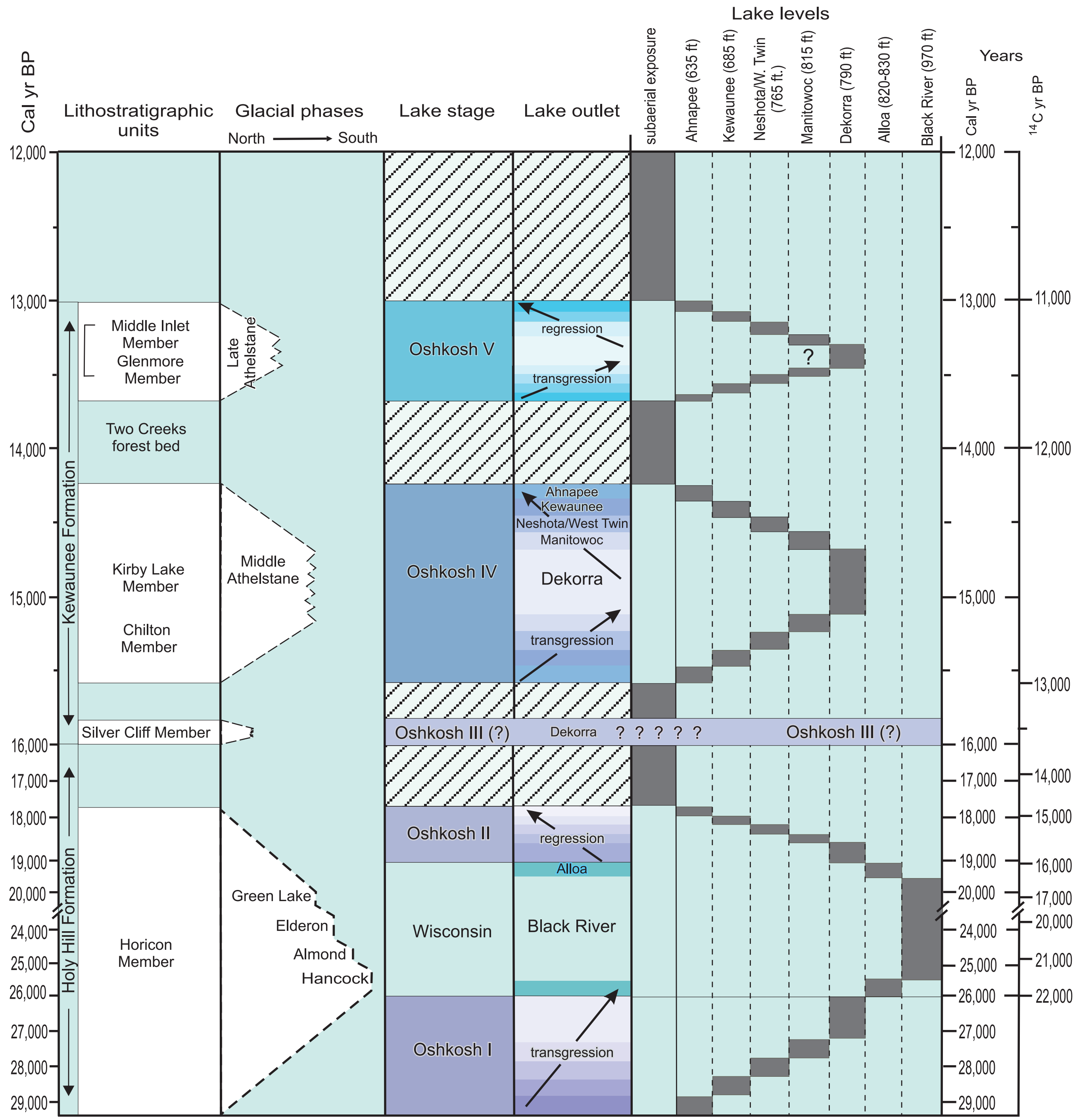

Figure 4. Chronology in the Oshkosh basin. The vertical time scale was calibrated according to Stuiver and others (1998).

the ice must have blocked the northward drainage of the Fox River and formed glacial Lake Oshkosh I. Unfortunately, there is little to no evidence of this lake, which may be the result of intense subglacial erosion.

Following recession of the lobe from its maximum extent, the basin was reopened and a thick sequence of fine-grained sediment was deposited in low-lying areas, reforming the lake (glacial Lake Oshkosh II). With continued recession of the ice margin, lower outlets opened to the Michigan basin, eventually draining the lake. Evidence to support this interpretation includes organic material that is preserved in places beneath a younger till sheet (see Mickelson and others, this volume). Pollen analyses on some of this organic material indicated that tundra conditions were prevalent (Maher, 1970; Maher and others, 1998). Shortly following this period, the Green Bay Lobe must have readvanced westward across the northern part of the basin, depositing a sandy red till (the Silver Cliff Member of the Kewaunee Formation; McCartney and Mickelson, 1982). Given the limited extent of this till, it is unclear how it relates to glacial Lake Oshkosh III because it has not been observed in cores from deep boreholes within the central and southern parts of the lake basin. This readvance (and recession?) of the Green Bay Lobe to the west was followed by a major readvance southward during the Middle Athelstane glacial phase. Lake sediment was deposited in the basin (Oshkosh IV) in front of the advancing ice margin and then buried with fine-grained till of the Kirby Lake and Chilton Members of the Kewaunee Formation, equivalent to the Valders Member till farther west. Unlike the older Horicon till, the Kirby Lake till consists primarily of reworked lake sediment and represents an important marker bed across the basin. In addition, it usually overlies thick sequences of lake sediment deposited in glacial Lakes Oshkosh II, III, and the transgression of Oshkosh IV. At the apex of the readvance, the ice formed the prominent Eureka moraine that rims the basin (see stop 1 description, this volume). The Kirby Lake till is covered with lake sediment that was deposited during the regression of glacial Lake Oshkosh IV as the ice margin receded.

As the Green Bay Lobe melted back from the basin, all four eastern outlets were probably again used to drain the lake. The climate must have warmed significantly during this time because the Two Creeks forest developed in the region; it covered parts of 11 counties (approximately 2,500 km^2). At least 40 radiocarbon dates on wood bracket the existence of the forest between 14,400 and 13,100 cal yr BP (12,400 and 11,200 ^{14}C yr BP)[1]. Part of this forest was eventually overrun by ice in a second major readvance of the Green Bay Lobe during the Late Athelstane (fig. 4); this advance terminated near the present location of the city of Appleton on the northern shore of Lake Winnebago. In areas north of this ice margin (up ice), remains of the forest bed are usually buried beneath fine-grained red till composed of reworked lake sediment of the Middle Inlet, Glenmore, and Two Rivers Members of the Kewaunee Formation (see fig. 5.2 in stop 5 description, this volume). Beyond the terminus of this ice margin, most of the forest drowned in glacial Lake Oshkosh V as the lake level rose because the four lower outlets to the Michigan basin were covered by ice. The forest bed is well preserved in east-central Wisconsin because of the fine-grained nature of the overlying lake sediment and till. In upland areas not covered by the lake or the ice, there remains little to no evidence of the forest bed.

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[1] *Most radiocarbon dates used in this paper and in the field-trip stop descriptions are expressed as calendar years before present (cal yr BP), with plus or minus one standard deviation derived from CALIB calibration. This is usually followed by the radiocarbon-year equivalent. For general reference to timing of events, calibrated dates are often used with radiocarbon-year dates in parentheses. CALIB is a program from from the Quaternary Isotope Laboratory, University of Washington. This program was originally published in* Radiocarbon *(Stuiver and Reimer, 1986), and most recently revised in 1998 (Stuiver and others, 1998).*

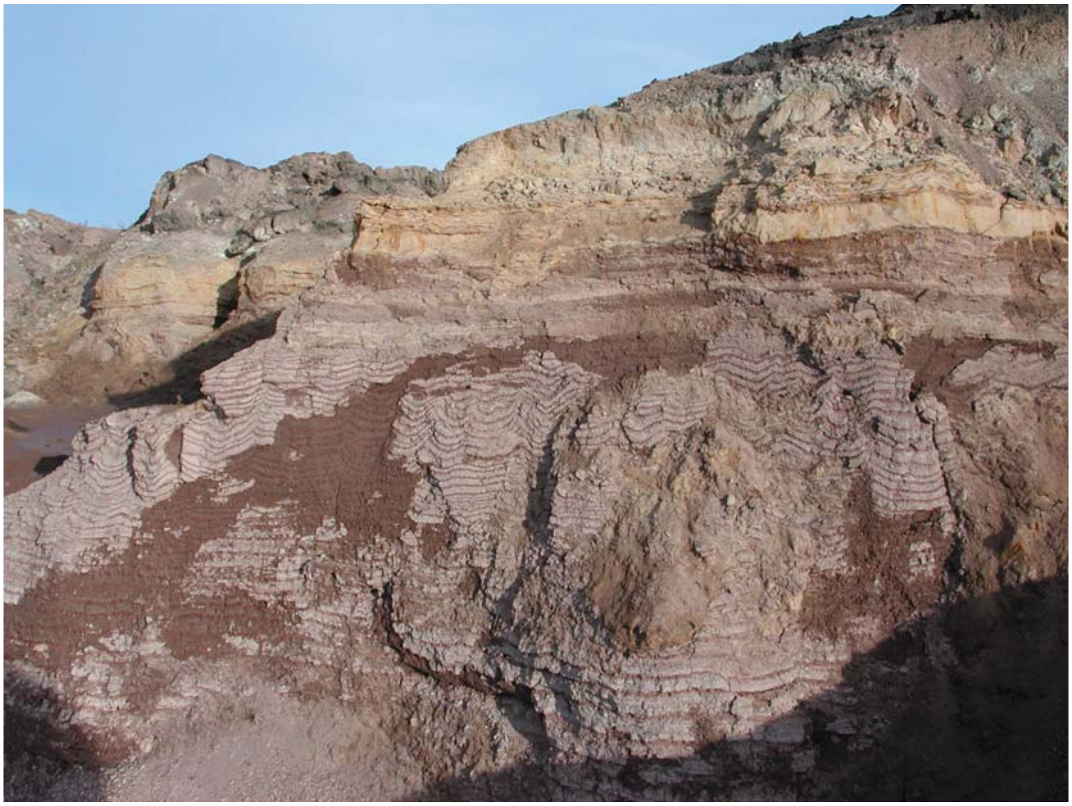


Figure 5. Photograph showing lake sediment typical of glacial Lake Oshkosh. The exposure is approximately 7 m high.

As the Green Bay Lobe receded from Wisconsin for the final time, the four eastern outlets were again uncovered, eventually draining glacial Lake Oshkosh V. Discharge from the lake cut channels downstream from each outlet. Aerial photographs and topographic maps show that the channels are wide compared to those of the modern rivers that now occupy them. This indicates that discharge events from the lake may have occurred quickly, thus incising the spillway channels (see stop 7 description, this volume) and causing a rapid fall of lake level. Depending on the location within the lake basin, such change in falling lake level would have resulted in a change in the sedimentary environment.

Field evidence

Understanding the geology of the Fox River valley is difficult because of the low relief of the landscape and limited outcrops. However, field evidence from a limited number of sand and gravel pits, shallow excavations, borings, and well-construction reports has documented the presence of extensive lake sediment and till. For example, a borrow pit in northern Winnebago County contains a sequence of fine-grained laminated lake sediment, consisting of silt and clay overlain by lake sand (fig. 5). This coarsening-upward sequence is typical of many exposures in which the laminated sediment, deposited in a quiet offshore environment, is capped by sand deposited in a higher-energy, near-shore environment as lake level fell when the lower outlets opened. An alternative explanation for the coarsening-upward sequence is that the basin was filled with sediment without changes in lake level.

The most convincing field evidence of glacial Lake Oshkosh includes a series of sediment

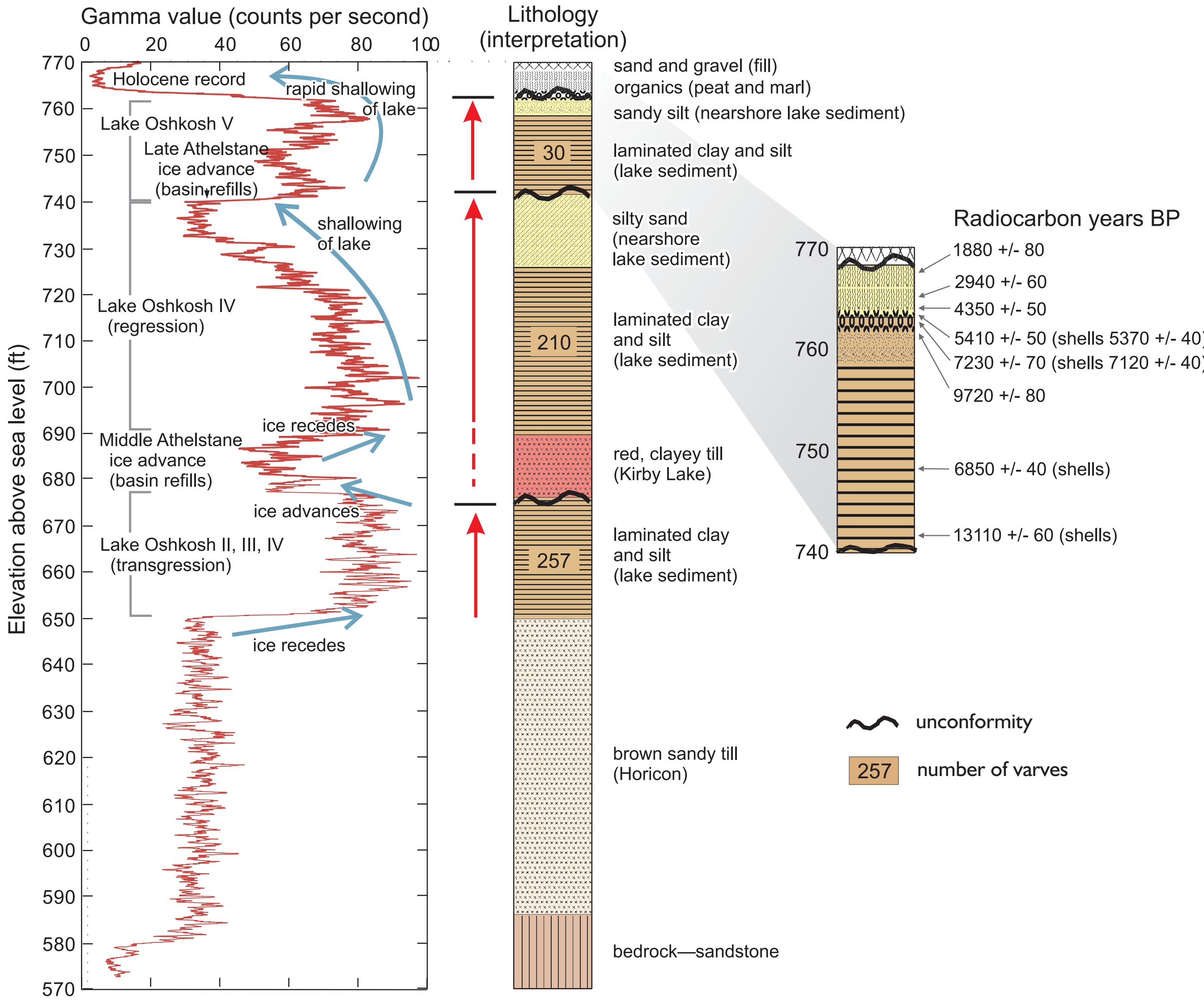


Figure 6. Downhole geophysical log of natural gamma radiation and associated lithologic log for rotosonic borehole RS-3 (see fig. 2 for borehole location), Outagamie County.

cores collected from a series of rotosonic boreholes drilled in nine counties in the Fox River lowland (fig. 2). Most of these borings were drilled in low-lying areas and encountered up to 100 m of lake sediment and till. One such borehole, RS-3, in Outagamie County (NE¼SW¼, sec. 24, T23N, R16E; fig. 2), was drilled on the edge of a marsh well below the highest level of glacial Lake Oshkosh at 771 ft (235 m) (fig. 6). The core, examined in detail, contains a complete record of late-glacial history, which includes three lake-sediment sequences documenting glacial Lake Oshkosh II through V. A downhole geophysical log of natural gamma radiation replicated the lithologic log, showing the various advances and recessions of the Green Bay Lobe as glacial Lake Oshkosh filled and emptied numerous times.

Unlike many rotosonic cores in the basin, the top of RS-3 contained a 2 m layer of peat, which included wood fragments that dated from 1,800 ±190 to 8,060 ±120 cal BP (1,880 ±80 to 7,230 ±70 ^{14}C yr BP) (fig. 6). Shells adjacent to two of these samples revealed similar

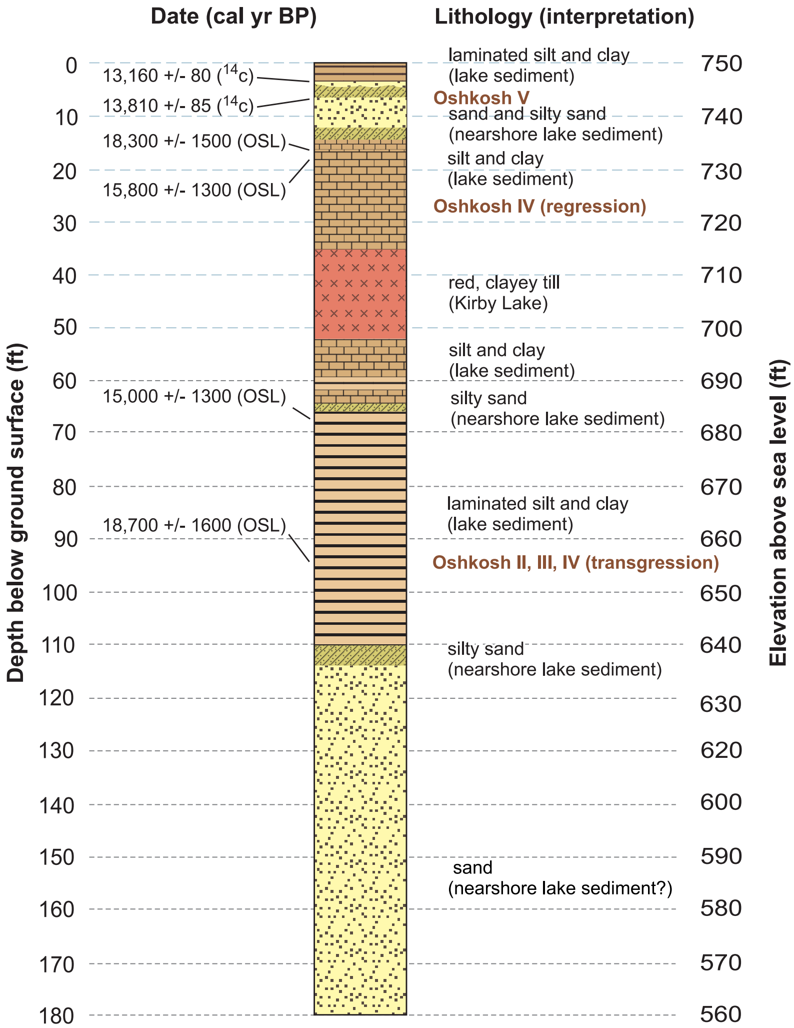


Figure 7. Lithologic log for borehole RS-1 (see fig. 2 for borehole location), Waushara County.

age dates. Immediately beneath the peat from 2.0 to 2.3 m was a 0.3 m layer of marl. Organic matter collected from this horizon dated to the early Holocene at 11,030 $\pm$120 cal yr (9,720 $\pm$80 ^{14}C yr BP). Of more interest was the rare presence of shells in lake sediment at 6 and 8 m below ground surface (elevations of 748 ft and 741 ft, respectively). The date for the shells in the 6 m sample was 7,730 $\pm$50 cal yr BP (6,850 $\pm$40 ^{14}C yr BP). It is difficult to make sense of the date for this sample because of its reverse chronological order in relation to the two samples immediately above it (fig. 6). The shells in the 8 m sample were dated at 15,505 cal yr BP (13,110 $\pm$60 ^{14}C yr BP), far older than we would expect to find at such a shallow depth.

Similar to most of the cores collected across the basin, RS-3 lacked the organic material, including fossils, to constrain the history of the lake basin (W.N. Mode, University of Wisconsin–Oshkosh, verbal communication, 2005; B.B. Curry, Illinois State Geological Survey, verbal communication, 2005). Thus, samples were selected from some cores in an attempt to date the lake sediment using optical stimulated luminescence (OSL) (Aitken, 1998). One core of particular interest, RS-1, was taken from within the margin of the Eureka moraine, but outside the Denmark moraine, which marks the maximum extent of the post-Two Creeks ice advance (fig 3H). Similar to RS-3, the core contains three sequences of lake sediment and one till that is interpreted as Kirby Lake (fig. 7). Two calibrated radiocarbon dates on wood extracted from the top lake-sediment sequence (13,160 $\pm$80 and 13,810 $\pm$85 cal yr BP [11,020 $\pm$40 and 11,480 $\pm$70 ^{14}C yr BP]) showed that the wood was Two Creeks in age. Because no organic material was in the remainder of the core, samples were collected from individual clay layers of the lake sediment for OSL dating. The dates, which ranged from 18,700 $\pm$1,600 to 15,300 $\pm$1,300 cal yr BP, indicated that the lake sediments were deposited since the late-glacial maximum. The older date from the first (deepest) lake sediment sequence indicated that glacial Lake Oshkosh II opened by approximately 19,000 cal yr BP. This was confirmed by OSL dates from another core. Although there is no age control, the lake may have drained completely shortly after 18,000 cal yr BP. As expected, the radiocarbon dates are younger than the OSL dates, given their location near the top of the core. However, the OSL

Figure 8. Photograph of wave-washed till surface in Waupaca County.

dates are not in chronological order. This is not surprising because some of the mineral grains may not have had time to "reset" prior to being deposited. In addition, the errors associated with each sample are sufficiently large, making it difficult to distinguish individual lake stages.

Beaches

Given the numerous phases of glacial Lake Oshkosh, very few beaches have been preserved in the Fox River lowland. This may be explained by fluctuating lake levels resulting from small changes in the elevation of the outlet. Such changes can cause shorelines to migrate laterally across the landscape, especially in shallow lake basins. In addition, the landscape was covered in fine-grained sediment and may have lacked enough sand to develop beaches. However, numerous wave-washed till surfaces can be seen at elevations between 800 and 860 ft (244 and 262 m) in the central and southern parts of the basin (see stop 1 description, this volume). These till surfaces are easily recognized by the large number of boulders present (fig. 8). It is apparent that the wave action along the shoreline was effective in eroding and transporting the fine-grained silt and clay, but left boulders to form a lag. In areas that have a source of sand, some baymouth bars and deltas can be observed, establishing the level of glacial Lake Oshkosh (see descriptions for stops 3 and 6, this volume).

Mechanisms of lake-level change

The change of lake level within the Fox River lowland has thus far been attributed to the opening of lower outlets resulting from the fluctuation of the glacial margin. However, changes in the volume of water entering the lake, downcutting of outlets, and differential isostatic changes in outlet elevation can also affect lake level (Hansel and others, 1985). Given that the Green Bay Lobe blocked surface-water drainage to the north, part of the ice margin always bordered the northern part of the lake. Thus, a large amount of the lake's water was probably derived from subglacial meltwater. The flux of this meltwater probably varied seasonally, but may have remained fairly steady over longer time periods. In addition, glacial Lake Oshkosh was also part of a much larger upland drainage basin that probably provided a large amount of surface water to the lake. Thus, the amount of water available to fill glacial Lake Oshkosh probably did not control lake level.

The topography surrounding each outlet indicates that downcutting was probably not a major factor in lowering lake level over an extended period of time. Most downcutting was at

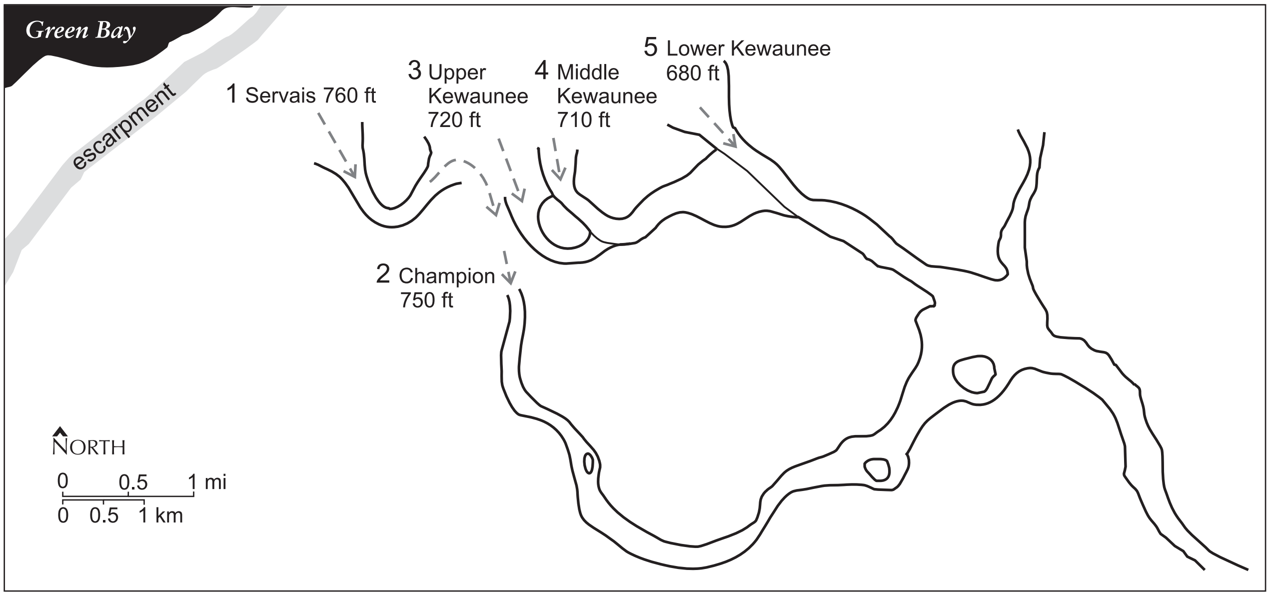


Figure 9. As the Green Bay Lobe receded, the flow of water from glacial Lake Oshkosh shifted from the channels at higher elevations (Servais, Champion, and Upper and Middle Kewaunee) to the Lower Kewaunee channel, which became the main outlet.

the Kewaunee outlet, where a series of small channels became progressively lower, from 770 to 680 ft (235 to 207 m) (fig. 9). Initially, the Servais channel was incised along with the Champion channel. This was likely followed by incision of the Upper, Middle, and Lower Kewaunee channels, respectively, as the ice sheet receded northward. All these channels were most likely cut relatively quickly in front of the receding ice margin, until the main channel stabilized at 680 ft. Once this channel was stabilized, it appears that little downcutting occurred, and lake level would have remained constant.

Glacial isostatic adjustment of east-central Wisconsin

Differential isostatic adjustment of the Earth has been known to be an important factor in changing lake level (Clark and others, 1990). This is probably not the case in the Oshkosh basin because the opening of lower outlets resulting from recession of the ice margin occurred at a much faster rate than isostatic rebound. This is not to say that isostatic adjustment of the Oshkosh basin is not important. In fact, isostatic rebound due to glacial loading has had a major effect on the Oshkosh basin and is important to understanding the history of the region.

Isostatic adjustment caused changes in land elevation, including the elevations of the outlets that controlled the lake level. The extent of glacial Lake Oshkosh at any stage varied as a function of time, depending on the amount of adjustment. For example, at any given outlet, such as Kewaunee, the extent of glacial Lake Oshkosh IV differed from that of Oshkosh V simply because the elevation of the outlets and land surface changed over time.

To help understand the glacial-isostatic adjustment of the Earth's surface in east-central Wisconsin, a numerical ice sheet model was used (Clark and others, 1990). Two important parameters of this model are the viscoelastic properties of the Earth's mantle and the thickness history of the Late Wisconsin ice sheets. Details of the model can be found elsewhere (Clark and others, 1990; Clark and others, in press).

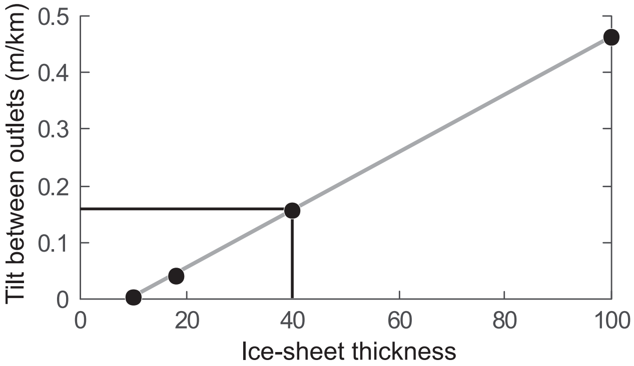


Figure 10. Magnitude of shoreline tilt between the Dekorra and Manitowoc outlets at 13,600 cal yr BP.

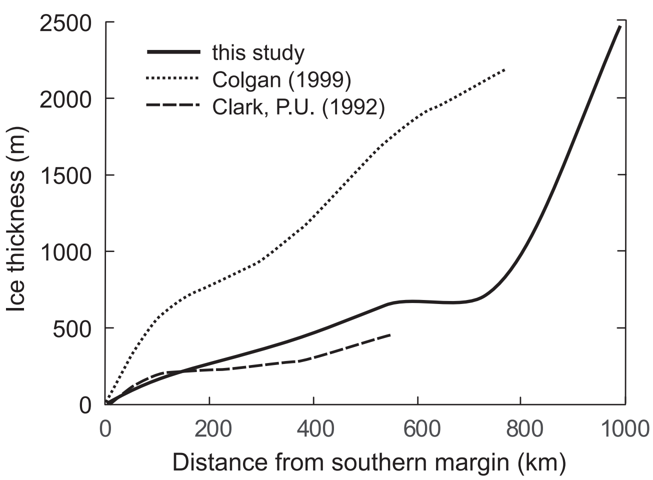


Figure 11. Calculated ice sheet profile along a north–south transect through east-central Wisconsin compared to profiles suggested by Colgan (1999) and Clark (1992).

Initial runs of the ice sheet model resulted in calculations of tilt of east-central Wisconsin three times greater than that observed across the Oshkosh basin. To address this poor fit, the thickness of the ice sheet was reduced over the Great Lakes region by a known proportion (Clark and others, in press). After several attempts, it became apparent that the calculated tilt was linearly related to the proportionality constant (fig. 10) and that only 40 percent of the ice thickness was required to fit the tilt datum. Therefore, the model used the original ice-sheet configuration everywhere except over the Great Lakes region, where the thickness is 40 percent of its initial value. This reduction in ice thickness results in the Green Bay Lobe at its maximum extent being approximately 300 m thick over east-central Wisconsin. This is similar to the ice-sheet reconstruction of Clark (1992), but thinner than that of Colgan (1999) (fig. 11).

With the ice-sheet model reasonably calibrated, deformation of the Oshkosh basin was calculated for the past 30,000 years. With the availability of high resolution digital elevation models (DEMs) and lake bathymetry, glacial Lake Oshkosh was reconstructed using only a DEM and the deformation calculations of glacial-isostatic adjustment. No assumptions about outlet locations or shoreline positions were needed for the calculation. We used the calculated deformations, relative to the present geoid, to warp the present DEM by the calculated amount through time. Hence, a series of paleo-DEMs was constructed for east-central Wisconsin, giving the topography at 1,000-year intervals for the past 30,000 years.

With the paleo-DEMs and an estimated location of the ice sheet margin, geographic information systems (GIS) were used to determine the extent of glacial Lake Oshkosh at the various stages. An ice-sheet dam was placed on the paleo-DEM and a hydrology extension of Spatial Analyst™ in ESRI® ArcView™ was used to create a drainage pattern. In the process, the GIS extension fills any sinks, and it is those filled sinks that define the flooded proglacial lake basins. The volume of glacial Lake Oshkosh at any one stage was determined by subtracting the filled from the unfilled paleo-DEM. An outlet was determined as the point on the calculated shoreline that had the greatest number of uphill cells contributing water. With this method, the extent of glacial Lake Oshkosh was calculated for numerous times (for example, 13,600 cal yr BP; fig. 12). The edge of the lake represents the shoreline location and can be loaded into a portable computer with a version of ArcView GIS and coupled with GPS to check calculations in the field and as an aid in field work. Also included in figure 12 are calcu-

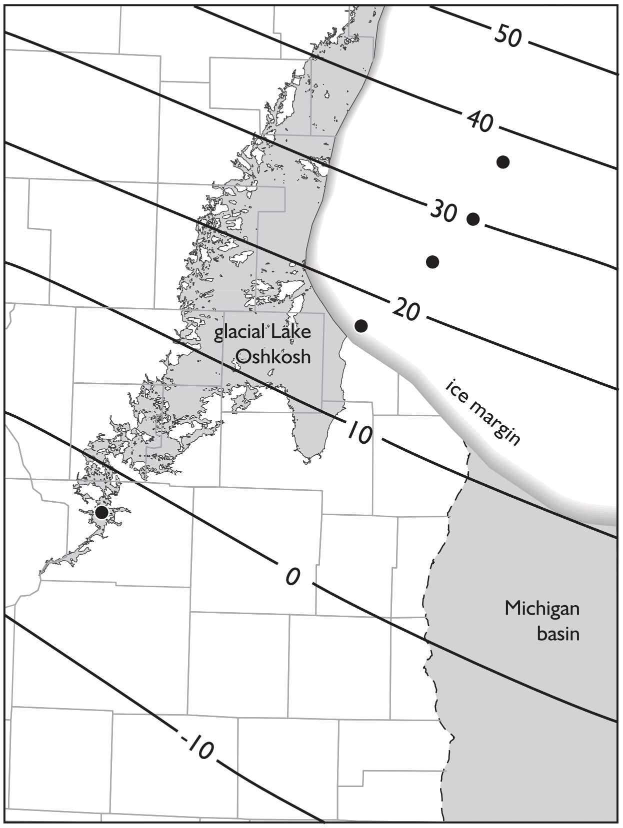
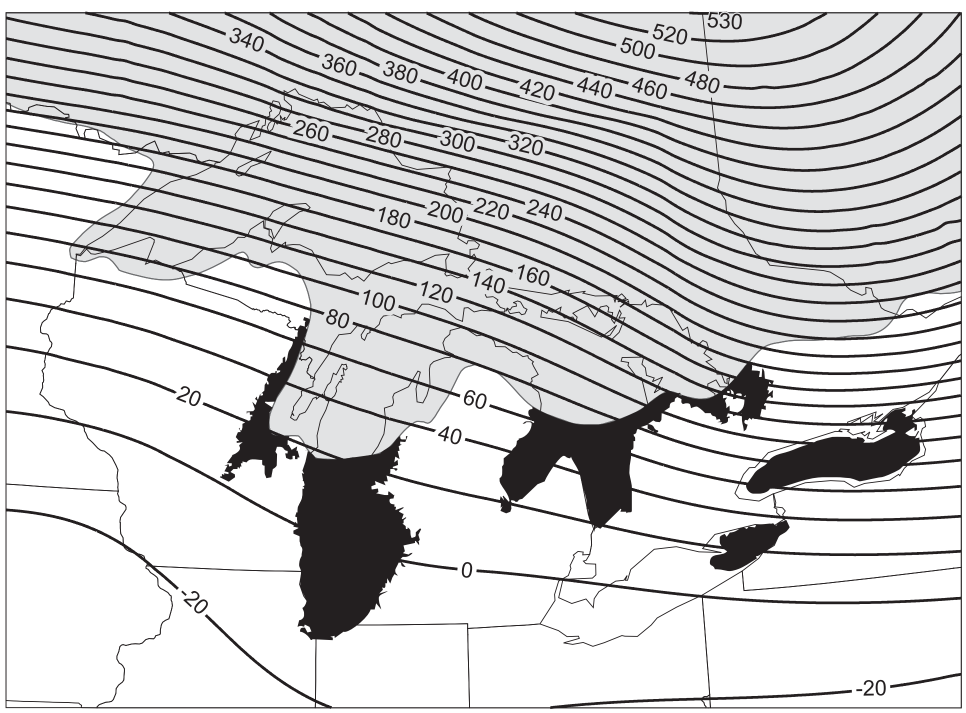


Figure 12. Isobases (in m) of deformation in relation to present geoid calculated for 13,600 cal yr BP for the Oshkosh and Michigan basins. Shorelines of glacial Lake Oshkosh formed at that time would be tilted upward 50 m toward the northeast between the southern and northern limits of the lake.

lated deformation isobases relative to the present geoid. To place these isobases in perspective, figure 13 shows isobases and calculated lakes over the entire Great Lakes region, although the DEM resolution for this large calculation was only 1,000 m; the resolution for the eastern Wisconsin calculation is 100 m.

Model results revealed that the Oshkosh basin is dynamic. Although it lies south of the "hinge line" proposed by others (Hough, 1958; Leverett and Taylor, 1915), the fact that tilting is occurring over the entire region even at present, and in a pattern easily understood by the glacial isostatic adjustment process, suggests that the hinge-line concept is in error. However, the magnitude of tilting is not as great as some who have been critical of the hinge-line idea have suggested. Figure 14 displays the amount of deformation calculated along a transect through the various outlets of glacial Lake Oshkosh. It is clear that the amount of tilt and the absolute elevation of the transect points varied during the glacial advance over the region (30,000 to 18,000 cal yr BP) and the subsequent retreat. Maximum tilt occurred at 18,000 cal yr BP with a value along the transect of about 0.2 m/km. Calculations of deformation for the known outlets

Figure 13. Isobases (in m) of deformation in relation to present geoid calculated for 13,600 cal yr BP for the Upper Midwest.



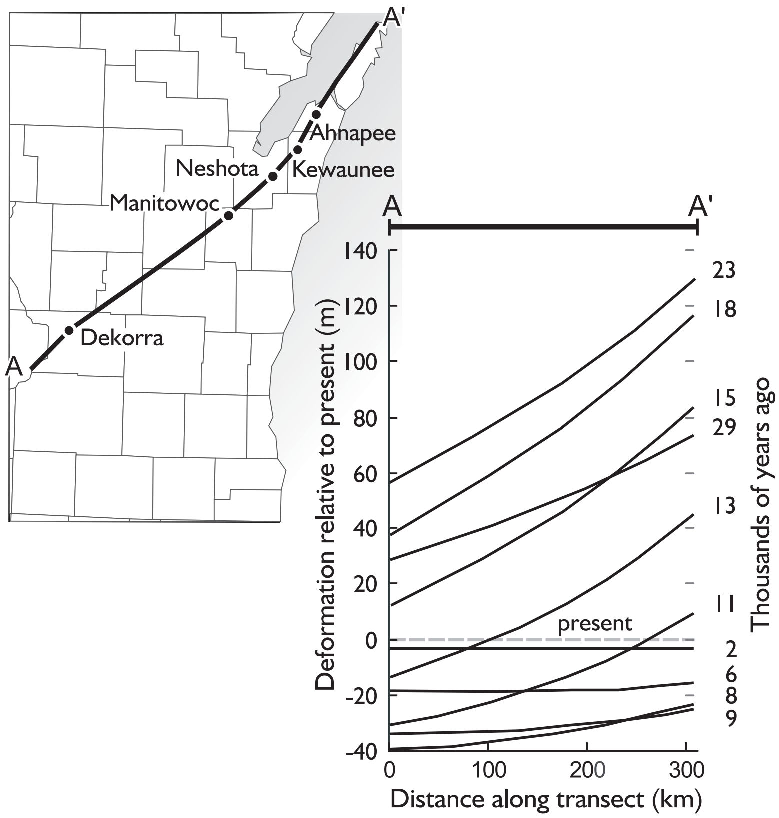


Figure 14. Deformation between 29,000 and 2,000 years ago in relation to present along a transect (A–A′) that intersects the outlets of glacial Lake Oshkosh. These deformations, when subtracted from the present DEM, give the elevations relative to present sea level. The amount of tilt and the magnitude of deformation differs at the indicated times.

of glacial Lake Oshkosh indicated that the outlets experienced initial subsidence during ice advance, followed by uplift. Then, as a migrating forebulge passed through the region, subsidence occurred after glacial Lake Oshkosh drained completely. Approximately 100 m of vertical movement affected the region.

Knowing the elevation of the outlets as they became ice free, the area and volume of glacial Lake Oshkosh can be determined at various phases. Using glacial Lake Oshkosh V as an example, the model showed that the outlets decreased in elevation as the ice receded northward (table 1). Although the Manitowoc outlet is now higher in elevation than the Dekorra outlet, model calculations indicated that it was lower than the Dekorra outlet when ice receded from the area. At its maximum extent, glacial Lake Oshkosh V was approximately 6,625 km^2 and had a mean depth of 17 m. With the opening of the Manitowoc outlet, the area of the lake did not dramatically decrease, although its volume was reduced by almost by 40 percent. The largest drops in lake level occurred when the Neshota and Kewaunee outlets opened, lowering the lake 20 m and 30 m, respectively. With each of these drops in lake level, the volume was reduced by a factor of three. The gradual recession of the ice margin to just north of the Kewaunee outlet increased the volume and area

Table 1. List of outlets for glacial Lake Oshkosh. A numerical model that calculated isostatic adjustment as a function of time was used to calculate past outlet elevation and to determine area, depth, and volume of each lake phase associated with the given outlets.

Lake outlet	Time (cal yr BP)	Outlet elevation in ft (m)		Lake area (km^2)	Mean depth (m)	Lake volume (km^3)	Volume difference (km^3)
		Present	Past				
Dekorra	13.6	784 (239)	774 (236)	6,625	17	112	–
Manitowoc	13.3	817 (249)	758 (231)	5,553	13	70	41
Neshota	13.2	768 (234)	692 (211)	1,778	21	37	33
Kewaunee	13.1	702 (214)	594 (181)	497	16	8	29
Ahnapee	12.9	637 (194)	531 (162)	1,464	15	22	-14

of glacial Lake Oshkosh as the depth increased within Green Bay. Once the ice receded past Sturgeon Bay, the level of glacial Lake Oshkosh was lowered to that of glacial Lake Chicago in the Michigan basin.

Summary

The history of glacial Lake Oshkosh is complex, given the number of outlets and lake levels associated with each advance and recession cycle of the Green Bay Lobe. No information is available to provide dates for glacial Lake Oshkosh I, but on the basis of OSL dates on lake sediment, it appears that glacial Lake Oshkosh II may have formed close to 19,000 cal yr BP and drained completely for the first time shortly after 18,000 cal yr BP. Several radiocarbon dates on organics have shown that the Green Bay Lobe readvanced during the Middle Athelstane at approximately 16,000 cal yr BP, which reactivated glacial Lake Oshkosh IV. The lobe reworked lake sediment, forming the clay-rich till Kirby Lake and Chilton tills of the Kewaunee Formation; this formation is an important stratigraphic marker in the central and southern part of the Fox River lowland. The extent of this ice advance is marked by the formation of the Eureka moraine.

Glacial Lake Oshkosh eventually drained as the ice receded northward and a forest developed in the region. Part of the forest bed was subsequently overrun by the last major readvance of the Green Bay Lobe. Beyond the ice margin, parts of the forest were drowned when glacial Lake Oshkosh V formed. Remains of this forest are well preserved; radiocarbon dates show it existed from 14,400 to 13,100 cal yr BP (12,400 to 11,200 ^{14}C yr BP). When the ice margin receded from east-central Wisconsin 13,000 cal yr BP, the outlets and associated channels that cut across the Silurian escarpment were used for the last time.

As a result of isostatic readjustment in the region, a fluctuating ice margin, and use of various outlets, glacial Lake Oshkosh varied in size. With each recession of the Green Bay Lobe, lower outlets to the Michigan basin were opened and large volumes of water must have discharged down the various spillways that cut across the Silurian escarpment. The magnitude of these discharges was estimated from changes in lake volume that at times exceeded 30 to 40 km^3. The rate at which this water discharged through each outlet is unclear, although some large boulders have been observed in various spillway gravel pits (See stop 7 description, this volume).

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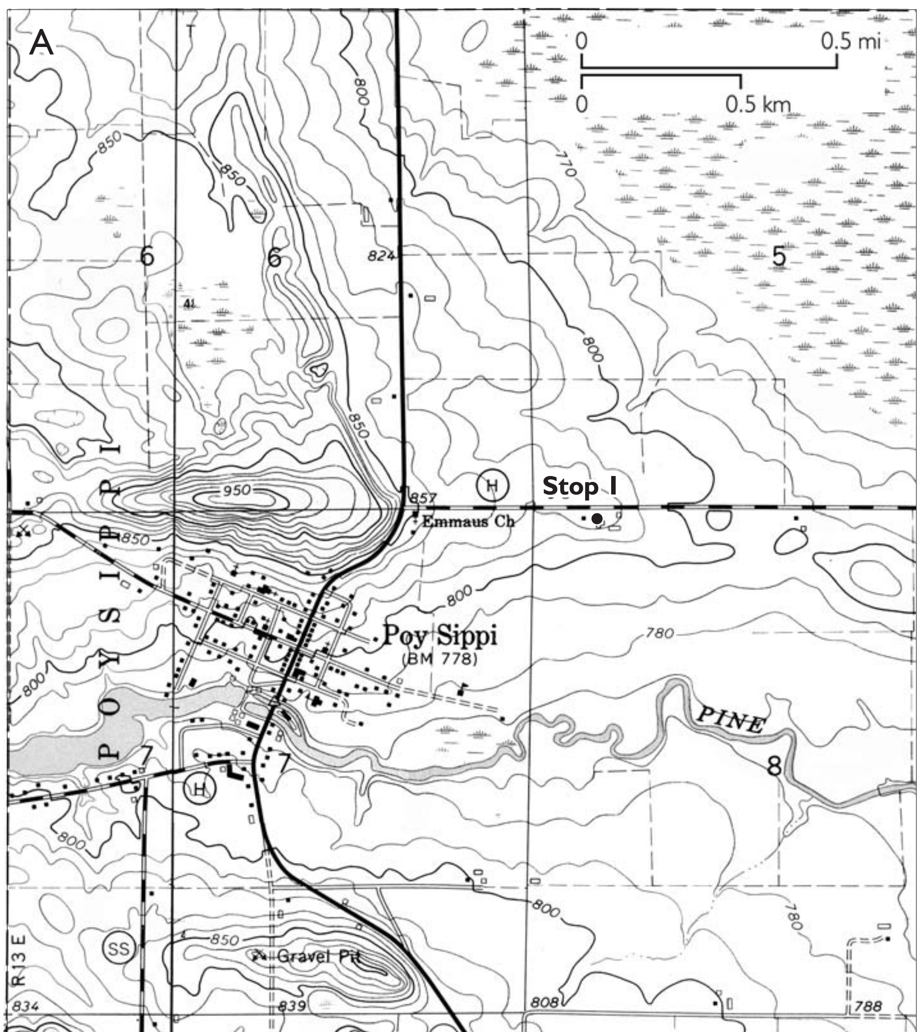
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R13E
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Stop 1

Eureka moraine and wave-washed till surface of glacial Lake Oshkosh, Buchholtz farm, Waushara County

Location: NW¼NW¼ sec. 8, T19N, R13E, Poy Sippi Quadrangle (1961), Waushara County. Stop is at a farm owned by Meryl Buchholtz. Permission is required to enter; the farm field is immediately south of the house on Highway H (fig. 1.1A).

Authors: Thomas S. Hooyer and James A. Clark

The Buchholtz farm lies along the westernmost margin of the Oshkosh basin (fig. 1.2). From this location, we can gaze south and east and visualize how low-lying areas were once filled with the water of proglacial Lake Oshkosh. This lake formed at the times that the Green Bay Lobe of the Laurentide Ice sheet advanced southward into Green Bay, blocking northward drainage of surface water from an area that spans the eastern third of the state. During the most recent glacial maximum, the Buchholtz farm was covered with about 200 to 400 m of ice (Clark, 1992; Colgan and Mickelson, 1997). Upon recession of the ice lobe, glacial Lake Oshkosh started to form between the outermost moraine and the ice margin. The lake grew as the lobe receded to the northeast, eventually uncovering the Buchholtz farm area. Initially, glacial Lake Oshkosh drained southward through the Dekorra outlet before lower eastern outlets to the Michigan basin were opened. The elevation of the Dekorra outlet—and thus lake level—was ap-

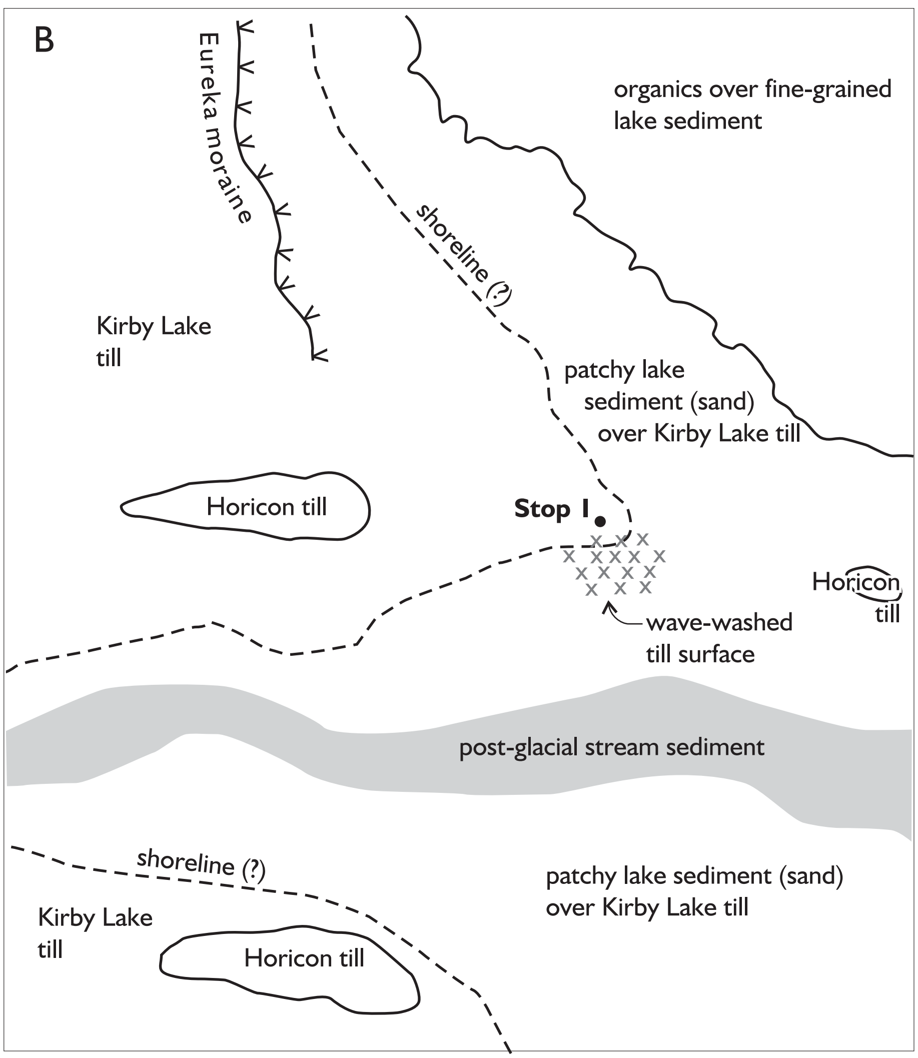


Figure 1.1. A. Part of U.S. Geological Survey Poy Sippi and Saxeville Quadrangles, Wisconsin (7.5-minute series, topographic, 1961), showing the location of the Buchholtz farm (stop 1), which lies along the western margin of glacial Lake Oshkosh. Note the break in slope at 770 ft, which also marks a change in materials from peat over fine-grained lake sediment (< 770 ft) to patchy sand over till (770 to 810 ft). **B.** Schematic showing the distribution of geologic materials, the location of the wave-washed shoreline, and a segment of the Eureka moraine.

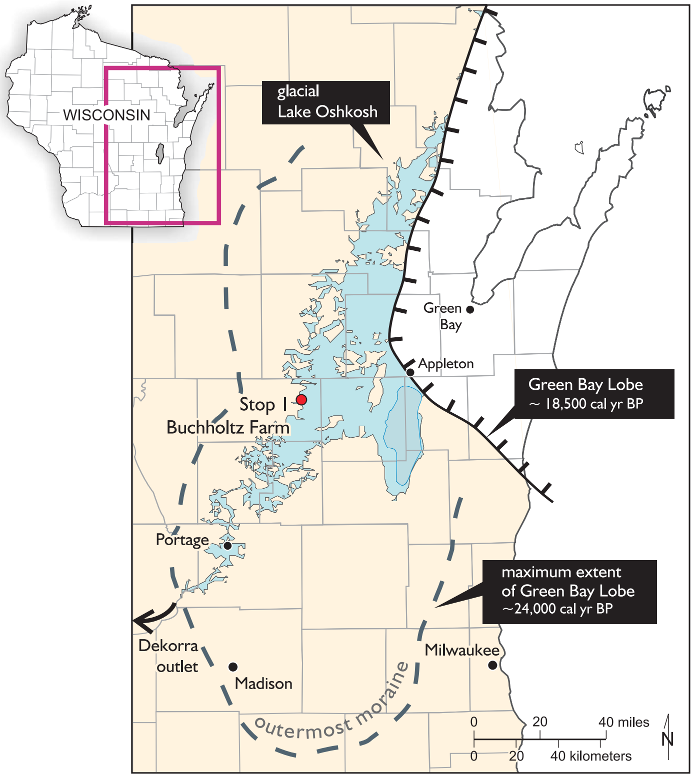


Figure 1.2. Map of east-central Wisconsin showing the approximate extent of glacial Lake Oshkosh in front of the Green Bay Lobe, the location of the outermost moraine, and the location of the Buchholtz farm (stop 1).

proximately 790 ft (241 m) in elevation. Glacial Lake Oshkosh did not completely cover the Buchholtz farm area because it is slightly higher—820 ft (250 m). Accounting for isostatic adjustment of the Earth, a shoreline should be present at this location at an elevation of approximately 810 ft (247 m).

Any evidence for a shoreline at the Buchholtz farm from this early stage of glacial Lake Oshkosh was probably modified when the Green Bay Lobe readvanced into the basin. The lobe reworked some of the fine-grained lake sediment, depositing an extensive till named the Kirby Lake Member of the Kewaunee Formation. This till is distinctively reddish brown and fine-grained compared to the brown, sandy till of the Holy Hill Formation, which was deposited when the ice initially receded from its maximum extent. The limit of the readvance into the basin (approximately 16,000 years ago) is marked by a moraine (fig. 1.3). This moraine, called the Eureka moraine, is relatively small, but prominent in the southern part of the Oshkosh basin, which is relatively flat. However, along the western margin of the basin, the moraine commonly abuts the hillier terrain, which consists of the older, sandy Horicon till. The Eureka moraine is easy to identify in this area because it is composed primarily of fine-grained till. A segment of the Eureka moraine can be observed immediately northwest of the Buchholtz farm (fig. 1.1B). The Kirby Lake till extends about 1,000 m farther west of the moraine; its extent marks the limit of a major readvance of the Green Bay Lobe. Streamlined upland areas consist of Horicon till and are interpreted as drumlins. Because no reddish-brown, Kirby Lake till covers these areas, they were probably sticking through the relatively thin ice margin.

Once the ice receded from the Eureka moraine, glacial Lake Oshkosh filled the low-lying areas against the ice margin along the western part of the basin. Given that the lake drained southward through the Dekorra outlet (fig. 1.2) at this time, the elevation of the shoreline was probably similar to that of the previous lake stage that used this outlet. Because subsequent readvances of the Green Bay Lobe did not reach as far into the basin, the shoreline developed on the Buchholtz farm should be slightly higher than the Dekorra outlet elevation when accounting for isostatic readjustment. The topographic map of the area surrounding the Buchholtz farm reveals a break in slope at the 770 ft (235 m) contour (fig. 1.1A), well below the expected shoreline elevation of 810 ft (247 m). The soil survey and geologic mapping indicate that this break in slope corresponds to a change in materials from organics over fine-grained lake sediment (<770 ft) to patchy sand over till (770–810 ft). Aerial photographs reveal no well developed beaches in the area, which may be the result of a limited sand supply; most

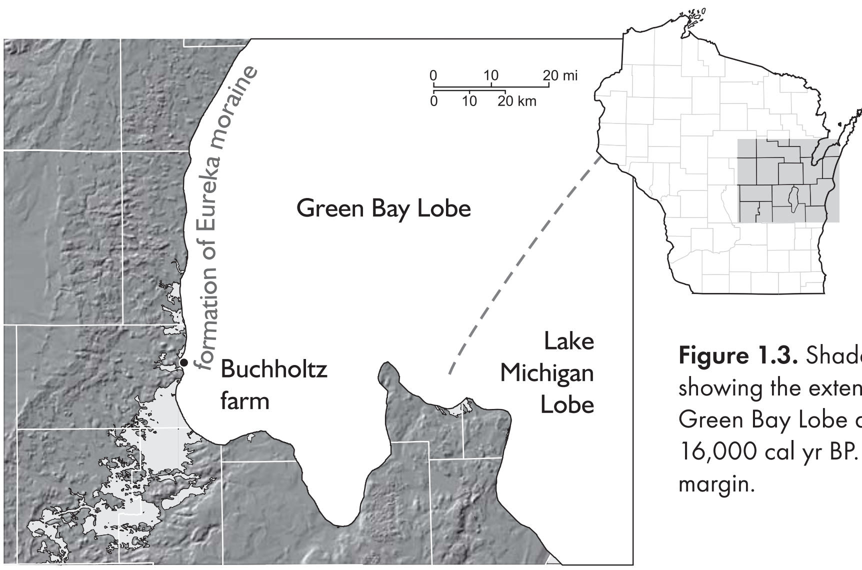


Figure 1.3. Shaded-relief image of east-central Wisconsin showing the extent of glacial Lake Oshkosh in front of the Green Bay Lobe during a major readvance approximately 16,000 cal yr BP. The Buchholtz farm is within 1 km of the ice margin.

of the shoreline in the central part of the basin terminated on fine-grained silt and clay till.

However, there is a concentration of boulders on a till surface at an elevation of approximately 810 ft. We interpreted this as a wave-washed till surface. This surface, one of many in the region, is on the south and east sides of the Buchholtz farm, where the shoreline was exposed to the greatest fetch of glacial Lake Oshkosh. It is apparent that the wave action along the shoreline was effective in eroding and transporting the silt and clay, but left the boulders to form a lag. We have interpreted the change in materials at the break in slope to represent the wave base, thus distinguishing lower- from higher-energy lacustrine environments.

Although the wave-washed till surfaces do not necessarily represent an exact level of glacial Lake Oshkosh, they were sufficient to calibrate a numerical ice-sheet model to determine isostatic adjustment (see introductory paper, this volume). Isostatic adjustment is important in understanding the late-glacial history of east-central Wisconsin, especially the elevations of outlets for the various stages of glacial Lake Oshkosh. Model results indicated that east-central Wisconsin, including the Oshkosh basin, is dynamic, although most of the area lies south of the "hinge line" proposed by others (Hough, 1958; Leverett and Taylor, 1915). Tilting of the land surface has occurred over the entire region and continues in a pattern easily explained by the process of glacial isostatic adjustment. Maximum tilt across the basin is approximately 0.2 m/km and occurred at approximately 18,000 cal yr BP. The magnitude of tilting, however, is not as great as suggested by those who have been critical of the hinge-line concept (Clark and others, 1990). The relative maximum amount of tilt between the cities of Madison and Green Bay is about 100 vertical meters. Model results of deformation for the known outlets of glacial Lake Oshkosh indicated that the outlets subsided during the ice advance, but were uplifted during ice recession and into the Holocene.

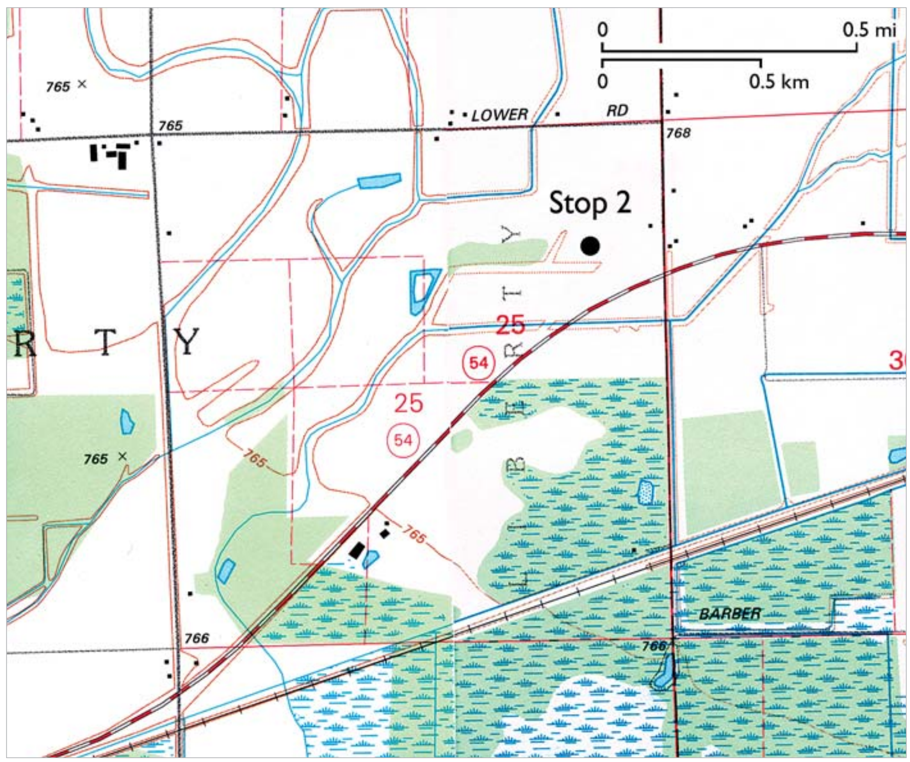


Figure 2.1. Parts of U.S. Geological Survey Schiocton and New London Quadrangles, Wisconsin (7.5-minute series, topographic, 1992), showing the location of stop 2.

▶ **Figure 2.2.** Idealized cross section, showing the extent of the Cambrian–Ordovician sandstone aquifer and the overlying confining unit, which consists of shale of the Maquoketa Formation and dolomite of the Sinnipee Group. The entire sequence is capped by relatively impermeable fine-grained glacial deposits. (Modified after Conlon, 1998; Batten and Bradbury, 1996).

Stop 2

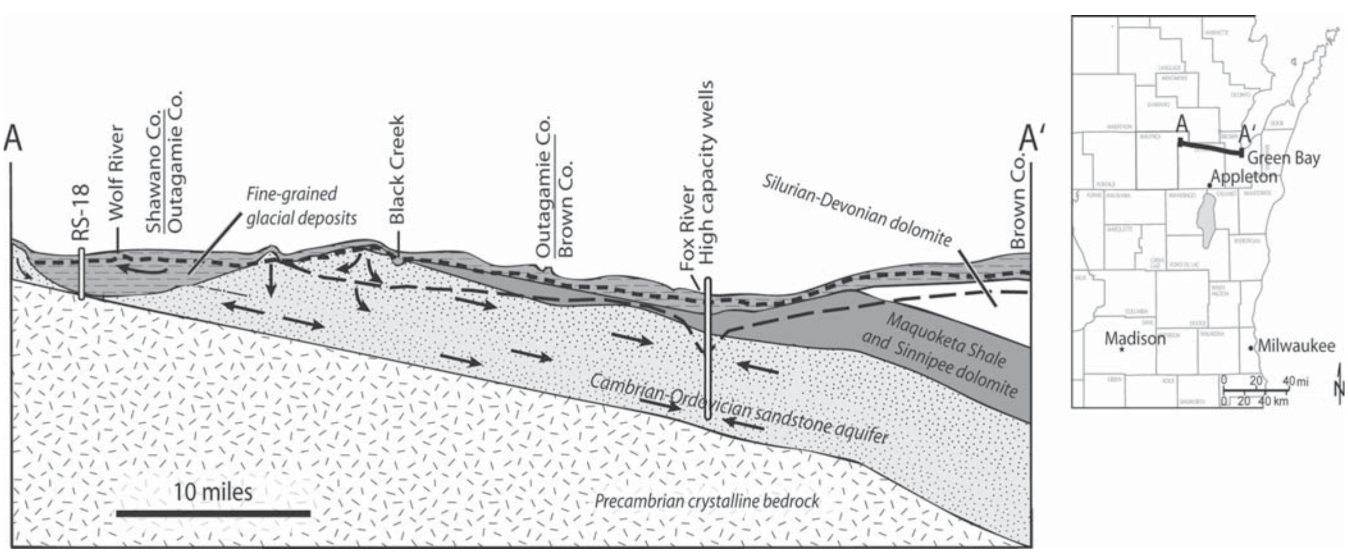
Investigating recharge to bedrock aquifers through fine-grained glacial deposits in east-central Wisconsin, Van Straten property, Outagamie County

Location: SW¼NE¼ sec. 25, T23N, R15E, Outagamie County (1992) (fig. 2.1). Stop is adjacent to Highway 54 near its intersection with Boelter Road.

Authors: Carolyn A. Moeller, Thomas S. Hooyer, and William G. Batten

Groundwater in the Fox River lowland has been of interest for the past 50 years because of concerns about this resource's availability in growing metropolitan areas, such as Green Bay, Appleton, Menasha, and Oshkosh (LeRoux, 1957; Olcott, 1968; Krohelski, 1986; Batten and Bradbury, 1996). Historically, discussion of the regional groundwater flow system has focused on the Cambrian–Ordovician sandstone aquifer, although the sand and gravel aquifer is present in isolated areas (Olcott, 1968). To gain a better understanding of groundwater recharge to the sandstone aquifer in the region, we are 1) characterizing the physical properties of the glacial sediment, including grain size and hydraulic conductivity, 2) examining the vertical distribution of hydraulic head in the glacial sediment by instrumenting two deep boreholes with multilevel well systems, and 3) determining the relative age of groundwater using stable isotope analyses on sediment pore water.

The sandstone aquifer consists of Cambrian (the Mount Simon, Eau Claire, and Wonewoc Formations of the Elk Mound Group, the Tunnel City Group, and the St. Lawrence and Jordan Formations of the Trempealeau Group) and Ordovician units (the Prairie du Chien Group and the St. Peter Formation of the Ancell Group). This aquifer is confined by dolomite of the Sinnipee Group and shale of the Maquoketa Formation (fig. 2.2) (Krohelski, 1986; Batten and Bradbury, 1996; Conlon, 1998). The primary source of recharge to the sandstone aquifer has been assumed to occur west of the Maquoketa and Sinnipee subcrop (fig. 2.2). This projected recharge area, however, is covered with fine-grained glacial sediment that has a relatively



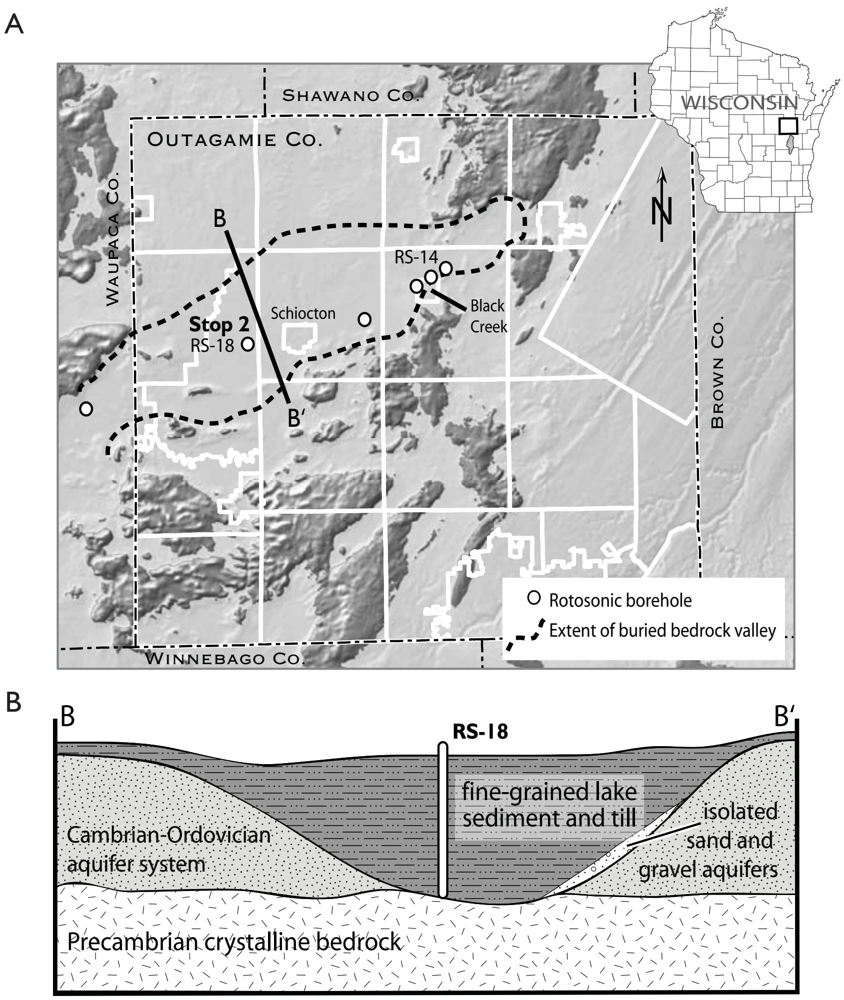


Figure 2.3. A. Shaded relief map of Outagamie County, showing the lateral margins of the buried bedrock valley and locations of rotosonic borings. The lighter gray shaded surface was once covered by glacial Lake Oshkosh. **B.** Conceptual diagram of the buried bedrock valley, with the location of rotosonic borehole RS-18.

low vertical and horizontal hydraulic conductivity. Thus, recharge to the bedrock aquifer may be largely controlled by the thickness of the fine-grained glacial sediment. This sediment varies in thickness across the region because of the undulating bedrock surface of a network of buried bedrock valleys; the bedrock surface is usually between 5 and 150 m below ground surface.

The axis of the most prominent buried bedrock valley extends westward across Outagamie County (fig. 2.3A) into eastern Waupaca County. This valley is more than 100 m deep at its center and is filled with fine-grained sediment; in the upland areas adjacent to the valley, the sediment is much thinner and composed of till (fig 2.3B). The thick sequence of lake sediment in the buried valley makes it unlikely that the bedrock aquifer receives recharge in this area. Recharge may take place on the uplands, where the till is relatively thin and potentially fractured (Helmke and others, 2005).

We drilled one rotosonic borehole, RS-18, just west of Shiocton (fig. 2.3A) through 100 m of sediment before reaching the Precambrian bedrock surface (fig. 2.4). The core consists entirely of fine-grained sediment, with the exception of some thin zones of sand and gravel near the bottom of the borehole. The borehole is between two moraine systems that represent the terminus of two major readvances of the Green Bay Lobe. We interpreted the red till in this core to be Kirby Lake Member of the Kewaunee Formation; we interpreted the remainder of the core to be lake sediment, most of it laminated.

We collected several intact core samples from this borehole as well as several other cores and outcrops for consolidation testing to determine the porosity and hydraulic diffusivity of the sediment. Results of the consolidation tests thus far have shown a vertical hydraulic conductivity of about 1 x 10^{-9} m/s for the lake sediment. Preliminary grain-size analyses of the laminated lake sediment showed this sediment to be very fine-grained with only slight variation in grain size between the silt and clay beds (fig. 2.5) (J.E. Rawling, UW–Platteville, written communication, 2007). Each silt layer is coupled with a clay layer, forming what we interpreted as a

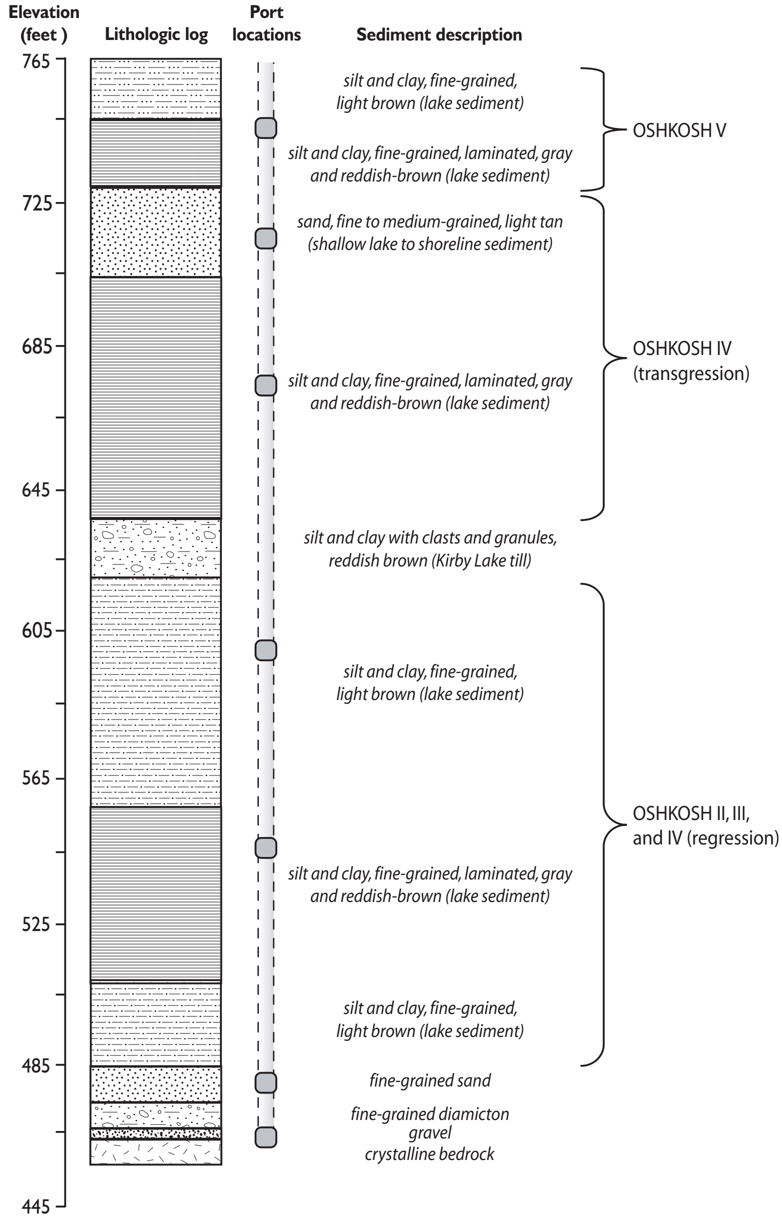


Figure 2.4. Lithologic log/interpretation and elevation of multilevel port placement for borehole RS-18.

varve. In any given core, we usually found hundreds to thousands of varves.

To measure the vertical head distributions in the glacial sediment sequence, we also instrumented RS-18 with a multilevel well system of seven piezometers, each with a diameter of 2.54 cm. Each piezometer screen is open to a sand-packed interval of 1.2 m. Screened intervals are isolated from each other by alternating lifts of bentonite within the borehole. Water-level data showed an upward gradient at this site (fig. 2.6). Using this gradient and the hydraulic conductivity determined from laboratory experiments, a calculation of vertical flux shows water moves approximately 3.9 x 10^{-9} m/s. With such a small flux, it would take more than 8,000 years for groundwater to advect through 100 m of sediment. Thus, diffusion is probably the primary mechanism of flow. This low vertical velocity indicates that relatively old water might still be present in the lake sediment.

To assess the age of groundwater in the glacial sequence, we extracted pore water from the RS-18 core in the laboratory by squeezing samples in stainless-steel vessels under high pressure. We are analyzing the pore-water samples for stable isotopes of oxygen and hydrogen, which can be used to interpret the climatic regime at the time the water was emplaced (Dansgaard, 1964). Oxygen isotope values of about -30‰ are typical of glacial-age water; values of -9‰ are indicative of modern precipitation. Unfortunately, sample analyses from RS-18 have not yet been completed. However, analyses of pore-water samples from rotosonic borehole RS-14 in the village of Black Creek, 14 miles east of RS-18 (fig. 2.3A), have been completed. Both boreholes are along the axis of the buried valley and consist of similar sequences of fine-grained sediment (figs. 2.4 and 2.7A).

Results of the oxygen-isotope analyses from RS-14 show a bow-shaped curve typical of modern oxygen isotope values (-9‰), diffusing downward from the surface (approximately 45 m depth) and upward from the bedrock into the clay aquitard (fig. 2.7B). Such a curve is typical of chemical diffusion with limited advection caused by differences in hydraulic head. The

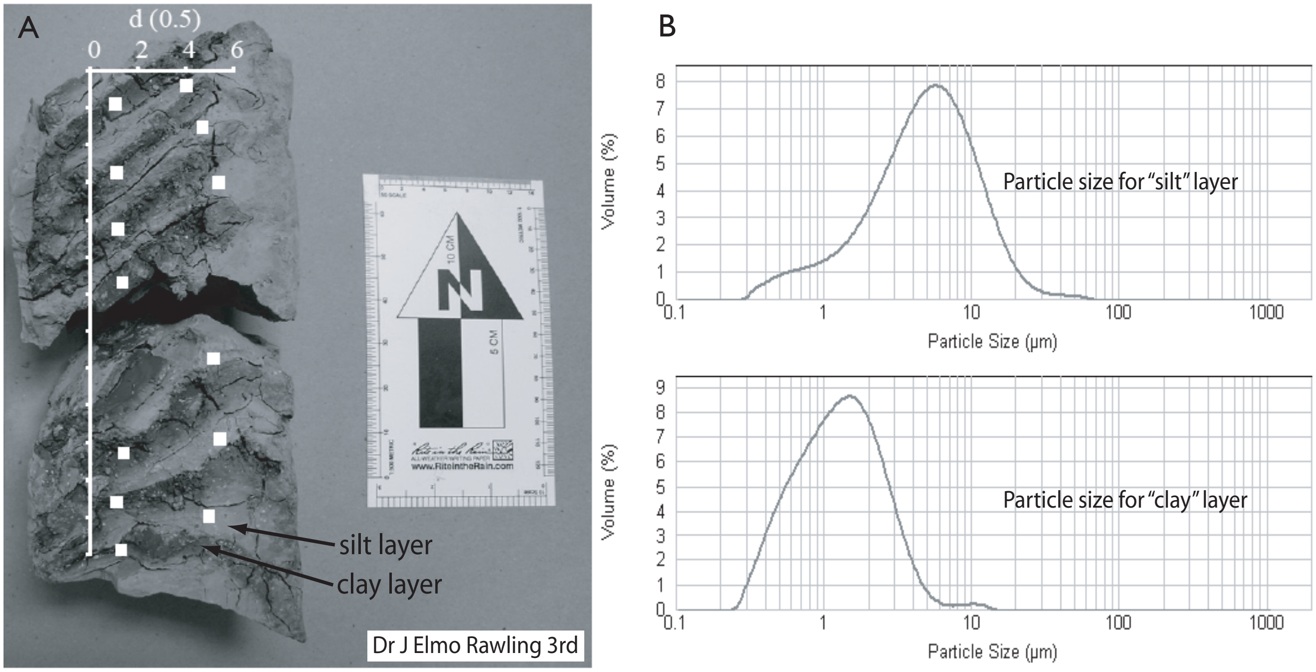


Figure 2.5. A. Photograph of rotosonic core showing laminated lake sediment, Outagamie County. The dark bands consist of clay and the lighter bands of silt. The squares represent sampling locations of the 25 cm long piece of core. The d(0.5) represents the median grain size in micrometers. **B.** Typical particle-size distributions for a silt and clay layer.

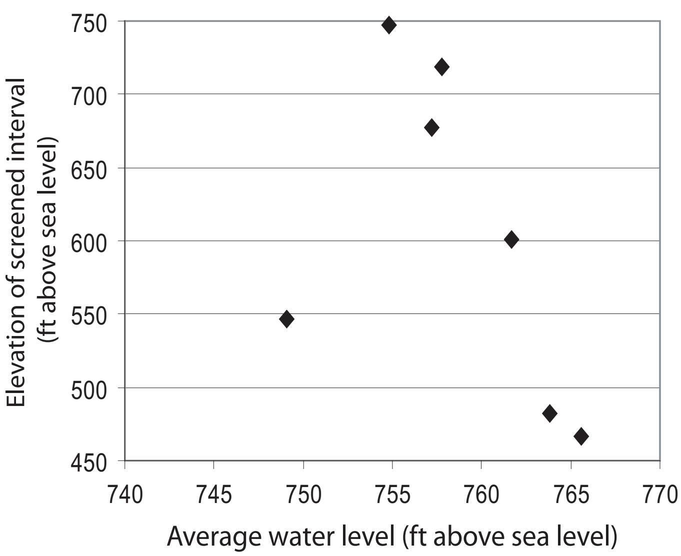


Figure 2.6. Average water levels by port elevation for the multi-level well system at borehole RS-18.

decrease in oxygen-isotope values (-15‰) toward the middle of the sequence suggests the presence of water from precipitation that could be glacially derived. Because the pore water in the lake sediment at the time it was emplaced would have been derived from the proglacial lake, and not directly from glacial meltwater, it is possible that the oxygen isotope values of the original pore water would be much lower than -30‰. Isotopic analyses from other locations in the lake basin where the lake sediment is thinner show that this "glacial" signal is not present.

The physical properties of the sediment indicate that little groundwater recharge takes place vertically through the glacial lake sediment, and the sandstone aquifer is probably recharged farther west, beyond the lake basin. Preliminary results of stable isotope

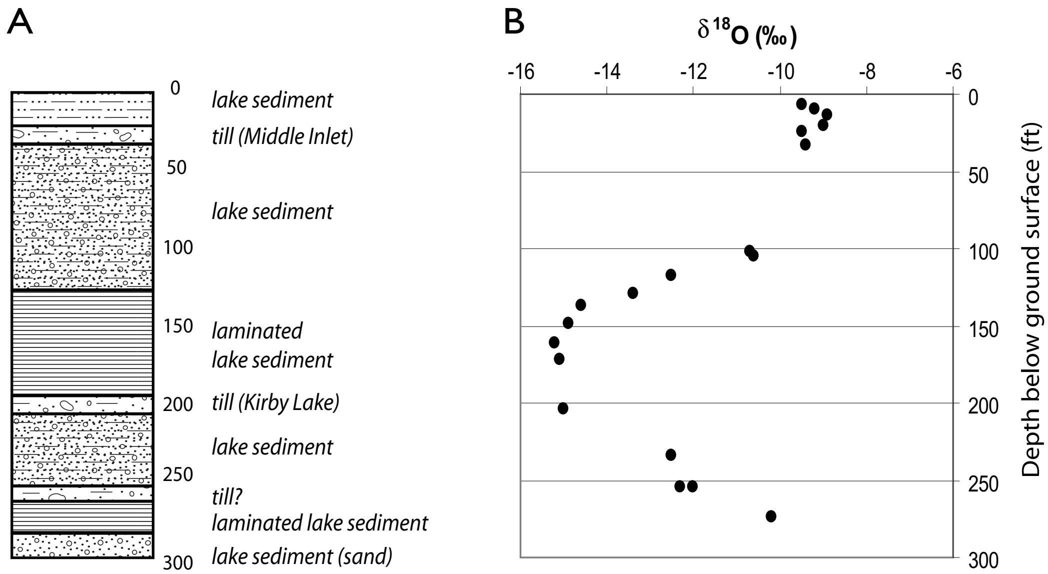
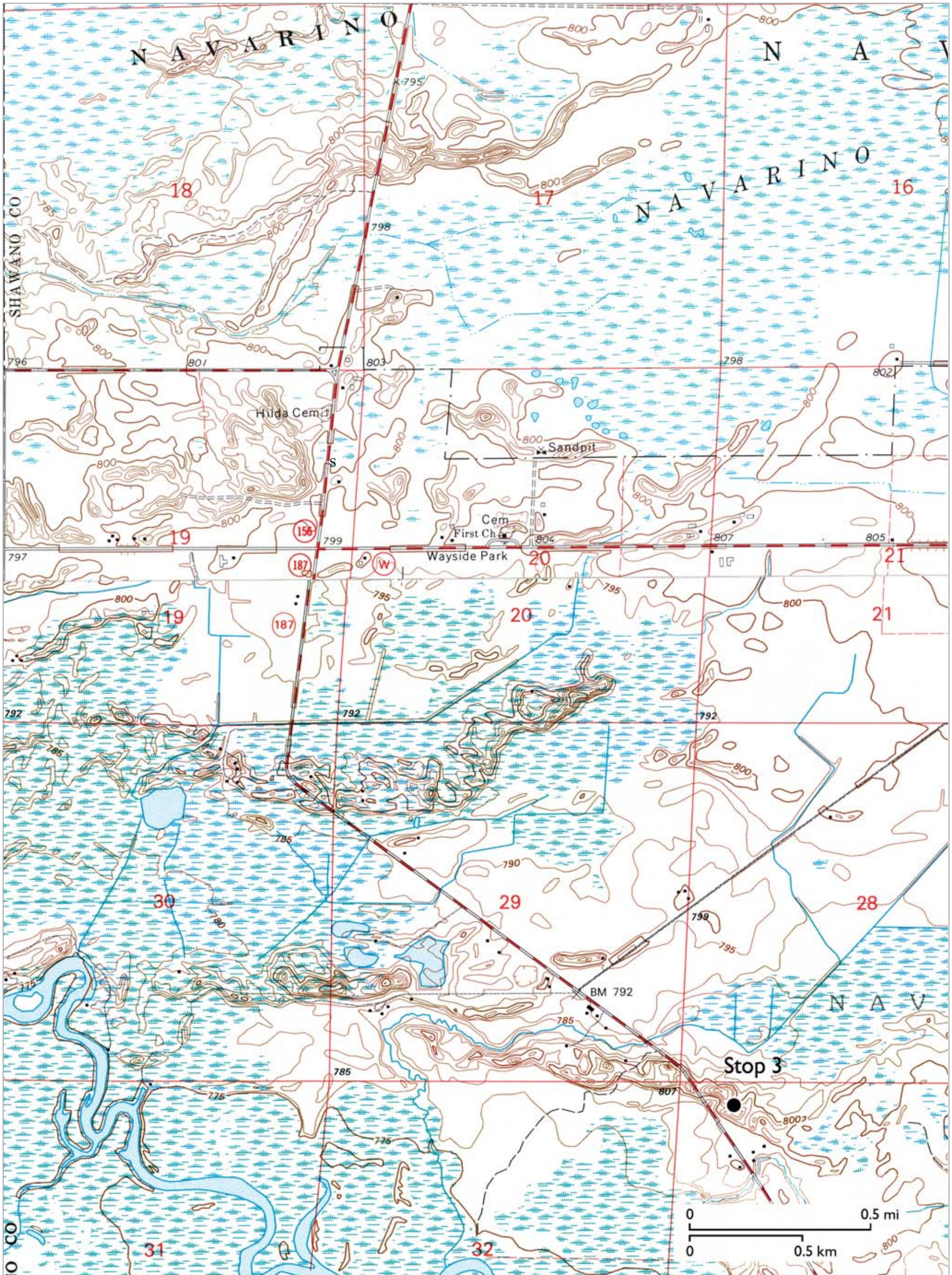


Figure 2.7. Lithologic log **(A)** and results of oxygen-isotope analysis **(B)** of pore water collected from a continuous core of fine-grained sediment from borehole RS-14, Black Creek, Wisconsin. Oxygen isotope analyses provided by John Cherry, University of Waterloo, Ontario.

analyses indicate a relatively old, potentially glacial, source for the pore water. This further indicates recharge to underlying bedrock aquifers is limited, especially in areas that have thick glacial lake sediment. However, in upland areas where the sediment is relatively thin (5 to 15 m), the bedrock aquifers are significantly closer to the surface. One possible hypothesis is that groundwater recharge occurs preferentially in these upland areas (LeRoux, 1957; Olcott, 1968). We plan to continue monitoring our existing deep wells and to install new monitoring wells where the sediment is significantly thinner and potentially fractured, allowing more rapid recharge to the sandstone aquifer.

The presence of relatively thick glacial lake deposits likely plays an important role in the flow of surface water, infiltration, and groundwater flow throughout the Fox River valley. Recharge to the bedrock aquifers may simply be a function of the hydraulic conductivity and thickness of the glacial deposits. The results of this project will not only contribute to our understanding of groundwater flow within the lake basin, but will also have implications for other regions, particularly where fine-grained glacial sediments overlie important aquifers.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Stop 3

Wolf River dune field, Shawano and Outagamie Counties

Location: NW¼NW¼, sec. 33, T25N, R16E, Shawano County (fig. 3.1). Stop is in a sand pit owned and operated by Marcks Trucking. The sand pit is off Highway 187; permission is required for entrance.

Authors: Stephen L. Forman and Thomas S. Hooyer

Eolian deposition during glacial cycles is well documented by ubiquitous loess and paleosol sequences in the midwestern United States. Other eolian systems were also active during ice-sheet expansion and recession, particularly during the formation of dune fields associated with outwash surfaces and proglacial lake margins. These dune fields reflect sand availability and supply and the dominant wind field, providing new insights about land-cover conditions and synoptic climatology in relation to ice-sheet advance and recession. At this stop we examine dune forms, stratigraphy, inferred paleo-wind directions, and optically stimulated luminescence (OSL) ages for the Wolf River dune field, an erg that formed with the drainage of glacial Lake Oshkosh. The OSL ages for dune sand provide chronologic control on changes in the levels of glacial Lake Oshkosh and the final disappearance of this proglacial system.

The Wolf River dune field is composed of blowout and single parabolic dune forms (fig. 3.1). The parabolic dunes are not fully formed, showing an asymmetry, with one arm longer than the other. Compound parabolic dunes have distinctive forms either in an en echelon progression of two to four dunes, with coalescence arms or with overlapping parallel noses forming an asymmetric sinuous ridge in places. This truncated and coalescenced morphology of parabolic dunes suggests a limitation of sand supply and/or time for formation. The orientations of more than 300 dunes were measured; results indicated a paleowind direction from the north to northwest (fig. 3.2). This is counter to the reconstructed wind field associated with the glacial anticyclone (Kutzbach and others, 1993).

Most of the dune field probably formed when glacial Lake Oshkosh drained for the final time, following the readvance of the Green Bay Lobe that buried the Two Creeks forest bed (Mickelson and others, this volume). Although the climate was warming, cold conditions must have persisted adjacent to the receding ice margin, prohibiting the growth of vegetation and subsequent stabilization of the landscape. An ample source of sand was adjacent to the ancestral Wolf and Embarrass Rivers, which were prograding into glacial Lake Oshkosh, forming a large delta that covers more than 150 square miles (fig. 3.3). This delta lies at an elevation of 800 ft (244 m) and is covered with parabolic dunes. A series of anastomosing channels at the end of the delta was most likely cut as the base level of glacial Lake Oshkosh continued to fall with the opening of lower outlets. Today, the Wolf River is incised into the original surface of the delta.

The sand pit (fig. 3.4) is in the nose of a parabolic dune in a field of compound para-

◀ **Figure 3.1.** Part of U.S. Geological Survey Leeman Quadrangle, Wisconsin (7.5-minute series, topographic, 1992), showing the location of the Marcks Trucking sand pit (stop 3) and the form of the parabolic dunes.

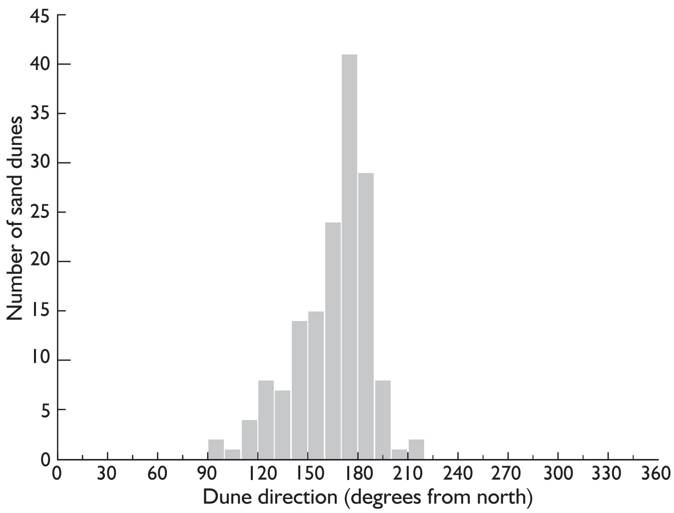


Figure 3.2. Vertical bar chart showing the direction of parabolic dunes in the Wolf River dune field at 10 degree intervals.

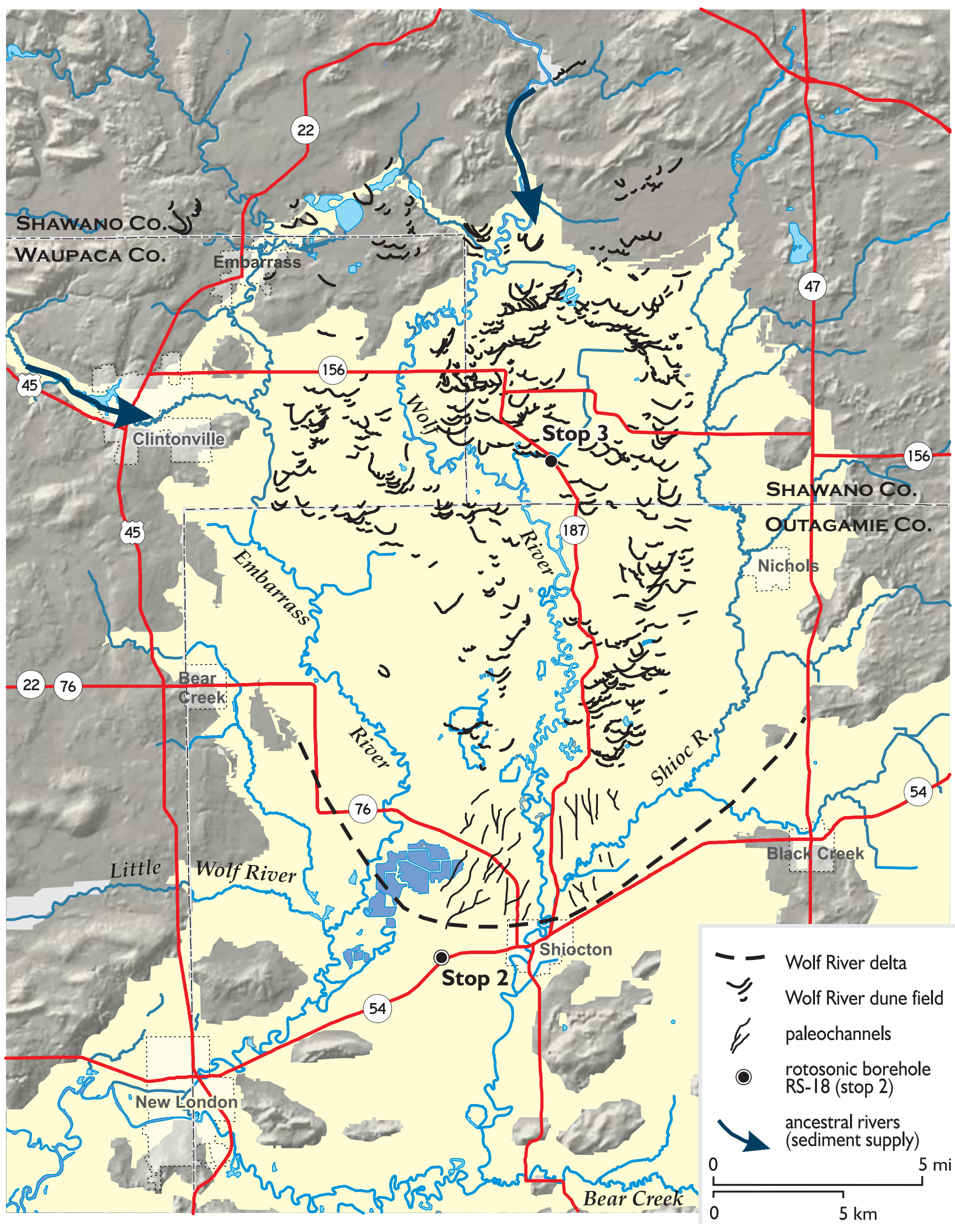


Figure 3.3. Shaded-relief map showing the extent of the Wolf River dune field and delta formed by drainage of the ancestral Wolf River into glacial Lake Oshkosh (800 ft [244 m] in elevation). The area shown in yellow represents the extent of glacial Lake Oshkosh V during its high stand.

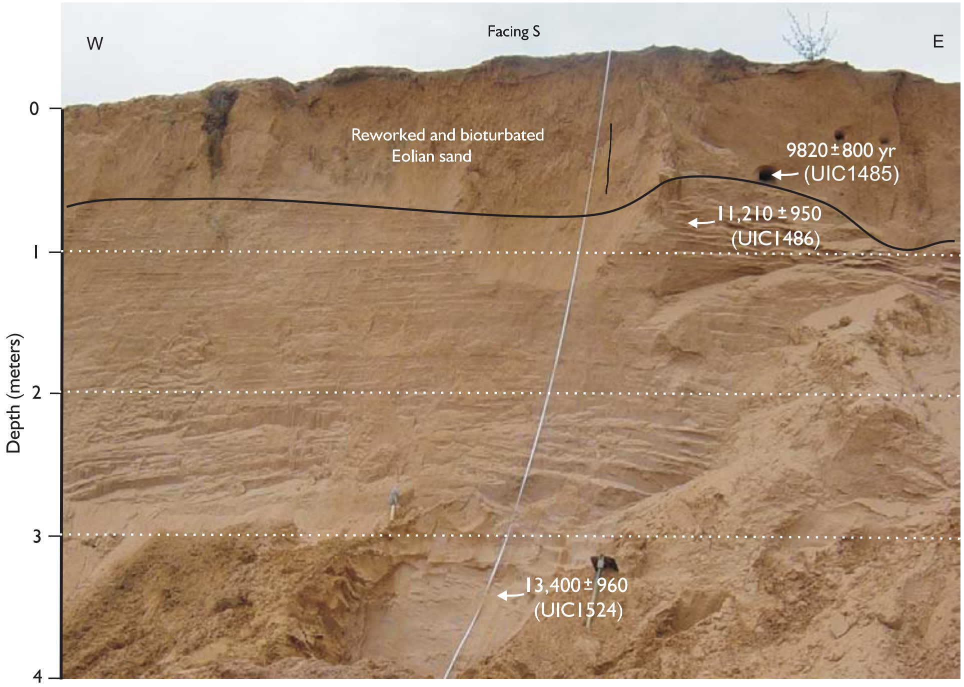


Figure 3.4. Photograph of Marcks Trucking sand dune, showing the locations and values of the OSL dates.

bolic dunes. This excavation exposes more than 4 m of very well sorted fine- to medium-grained sand deposited over glacial Lake Oshkosh sediment. Millimeter- to centimeter-scale bedding is evident and ranges from sub-horizontal to low- and high-angle beds (5º–25º); some are asymptotic and inclined north–northwest, reflecting avalanche processes associated with an advancing dune. The OSL ages from the top and base of the dune indicate that migration of this dune occurred approximately 13,500 to 11,000 cal yr BP. The upper meter is structureless, very well sorted, medium- to fine-grained sand; this layer reflects colluviation and pedogenic mixing of original eolian material and yielded an age of 9,800 cal yr BP.

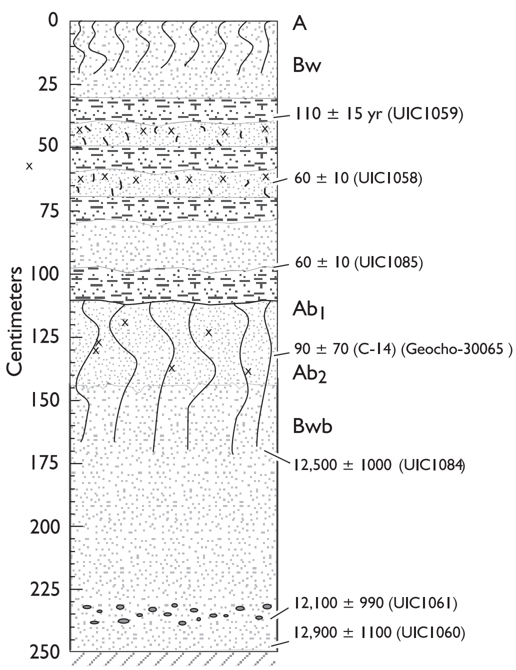


Figure 3.5. Lithologic log of the Walrath dune showing the locations and values of the OSL dates. The letters correspond to various soil horizons.

We may visit the Walrath site, which contains an exposure of the nose and arm of a parabolic dune. The basal sediment exposed at this site is very well sorted, medium-grained sand that has sub-horizontal to horizontal millimeter- to centimeter-scale laminations (fig. 3.5). Quartz grains from this unit yielded OSL ages of approximately 12,000 cal yr BP and provided a robust, minimum limiting age on the final draining of glacial Lake Oshkosh.

The upper boundary of this eolian sand is demarcated by a buried soil that shows clear and well preserved pedogenic morphologies with cumulic A and cambic B horizons. This soil is buried by more than 1 m of thick, horizontally laminated, well sorted, medium-grained sand to silty sand. The finer beds are dark (10YR 3/3), organic rich, and probably reflect resedimentation of A horizon material. In contrast, the sandy interbeds are brown to red (7.5YR 5/6) and similar in color to the subjacent buried cambic horizon. The OSL ages from this unit yielded surprisingly young ages of 100 to 60 cal yr BP. These OSL dates are confirmed by a radiocarbon date of 83 $\pm$60 cal yr BP (90 $\pm$70 ^{14}C yr BP) on dispersed organics. This nineteenth and twentieth century cover sand is banked up against the nose of a parabolic dune form, indicating deposi-

Figure 3.6. Photograph of reactivated sand dune encroaching on field in Oconto County, SE¼SW¼ sec.15, T28N and R17E. (Photograph by F.T. Thwaites, circa 1930.)

tional winds from the east to southeast. We interpreted this surficial unit to be a cover of sand associated with human-induced landscape denudation and disturbance. Some of this disturbance must have occurred during the early 1930s, when annual precipitation in east-central Wisconsin was less than 650 mm for several years in a row; figure 3.6 shows reactivation of sand dunes during the 1930s.

Stop 4

Duck Creek ridges: Long drumlins of the Green Bay Lobe, Outagamie County

Location: SE¼NW¼ sec. 11, T22N, R18E, Outagamie County (fig. 4.1). Stop is in a gravel pit owned by Murphy Concrete & Construction. Entrance to the gravel pit is off Center Valley Road; permission is required for entrance.

Authors: Lee Clayton, Thomas S. Hooyer, and William N. Mode

The Duck Creek ridges are two linear hills south of the city of Green Bay (figs. 4.1 and 4.2). These hills are up to 21 m high; they are from 13 to 30 km long and generally less than 0.8 km wide, giving them a length to width ratio of approximately 40:1. These ridges are on a broad plain that consists primarily of fine-grained till and lake sediment. On aerial photographs, this flat plain exhibits a longitudinal texture in the direction of ice flow between the sets of more conspicuous ridges. Upland areas to the east and west of this plain usually consist of thin, fine-grained till over Ordovician and Silurian dolomite. The Duck Creek ridges are the most prominent ridges and are along the western margin of the plain (fig. 4.2).

On our way to this field-trip stop, we travel down Seminary Road, along the long axis of the westernmost Duck Creek ridge, south of the town of Oneida; we can see the adjacent ridge to the east (left). Both ridges vary slightly in width and height along their longitudinal axis. The Duck Creek ridges are the longest streamlined subglacial landforms in Wisconsin. We can compare these ridges to the large drumlin field 100 miles south, toward the outermost margin of the Green Bay Lobe in southeastern Columbia County and west-central Dodge County (fig. 4.3A). These large, stubby drumlins are shaped like overturned canoes—they commonly are 10 to 30 m high and range from 0.2 to 0.4 km wide and 0.5 to 2 km long, giving them a length to width ratio of 2:1 to 10:1 (fig. 4.3B). The shorter drumlins tend to grade into incipient drumlins that are palimpsest pre-drumlin hills of various origins. Smaller yet are the inconspicuous spindly drumlins that are best observed on large-scale aerial photographs and difficult to note on topographic maps (fig. 4.4). The smallest are a few meters or less high, 10 m or less wide, ten to hundreds of meters long, and commonly have length to width ratios greater than 100:1.

Excavations at various places along the Duck Creek ridges expose a variety of Pleistocene sediment, including sand and gravel and fine-grained clay and silt (fig. 4.5) (Piette, 1963). In many places the sand and gravel is cross-bedded with channel cut and fill sequences that are interpreted as fluvial deposits. We have interpreted the fine-grained sediment to be till and lake sediment of glacial Lake Oshkosh. The upper and lower tills are probably part of the Kirby Lake and Middle Inlet Members of the Kewaunee Formation. Although it appears that the Middle Inlet till, which was deposited after Two Creeks time, does not always cap the fluvial deposits, the extensive soft-sediment deformation was probably caused by glacial loading (fig. 4.6), indicating that the ice lobe covered the ridges during the most recent glacial readvance.

Between the two Duck Creek ridges is a 40 mm thick forest-litter layer that includes fragments of tamarack wood sandwiched between lake sediment and till; this layer provides

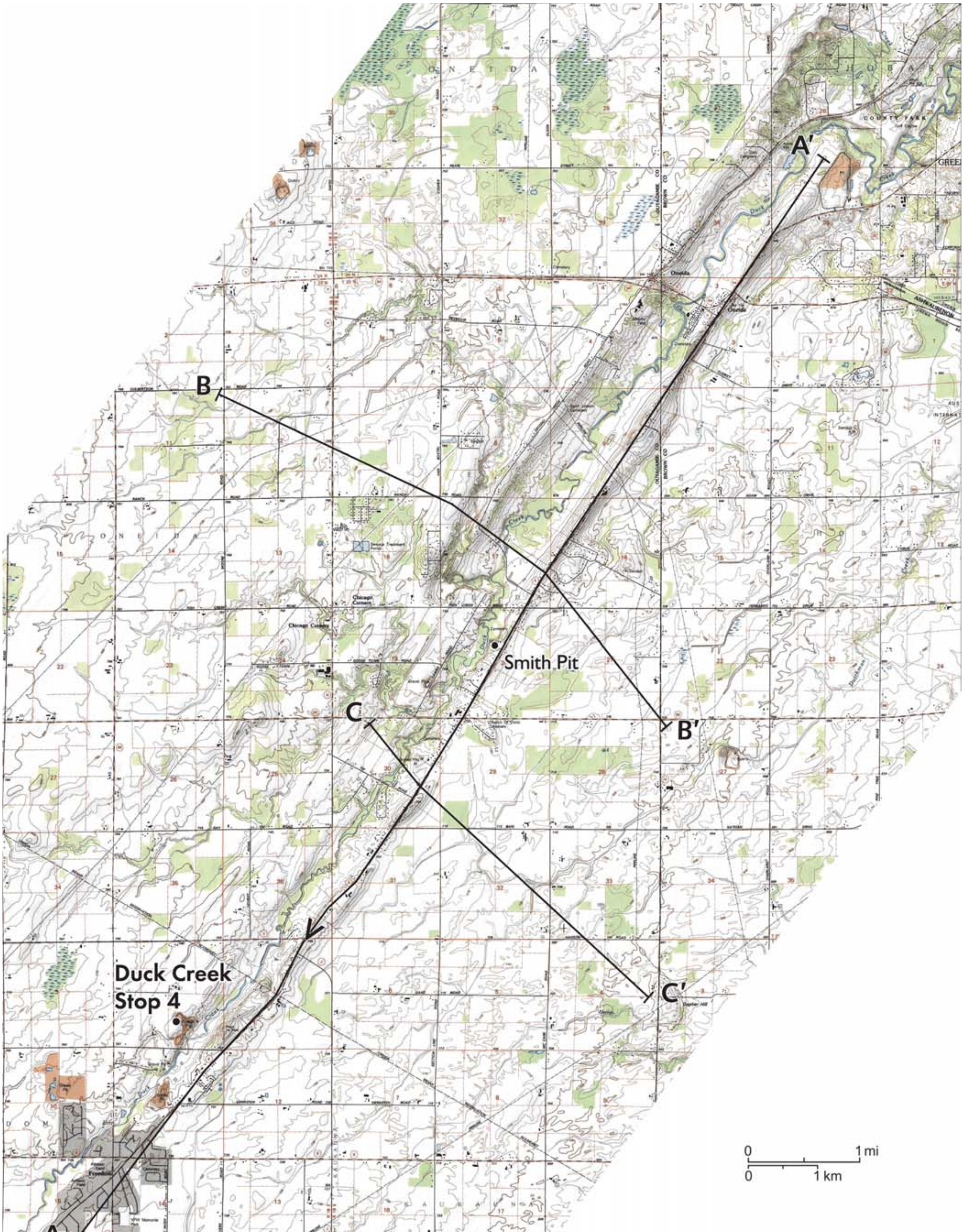


Figure 4.1. Part of U.S. Geological Survey Freedom Quadrangle, Wisconsin (7.5-minute series, topographic, 1992), showing the prominent Duck Creek ridges. Lines represent cross sections shown in figure 4.5.

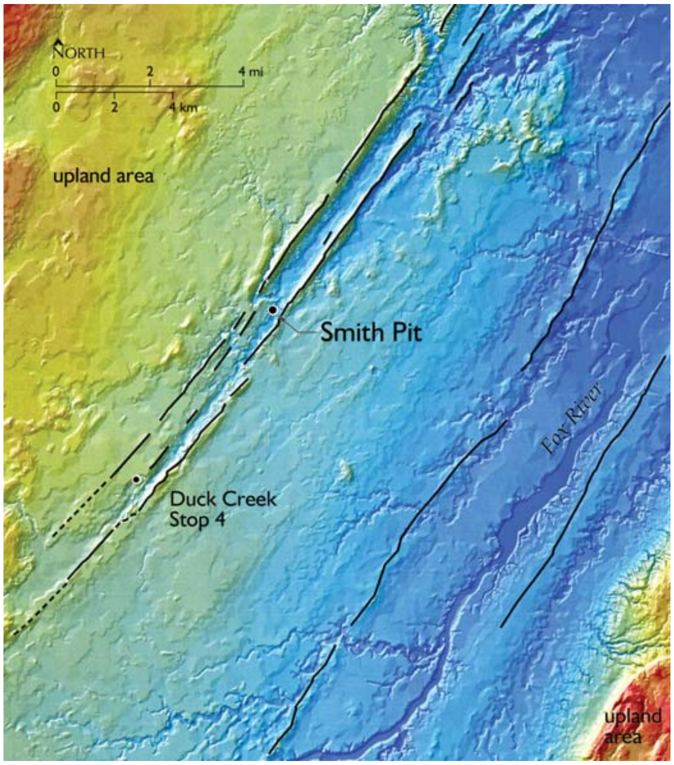


Figure 4.2. Shaded-relief image of the Duck Creek Ridges. The black lines define the longitudinal axis of each major ridge in the area. Note the low-relief lineations between the main ridges and the short, stubby drumlins northwest of the ridges.

additional evidence that the ridges were covered with ice during the most recent readvance of the ice lobe. The wood fragments have been radiocarbon dated to 13,505 ±340 cal yr BP (11,640 ±350 ^{14}C yr BP), similar to many other sites in the area (Piette, 1963). Pollen analyses of the organic layer of the same age found nearby indicated the dominance of spruce (*Picea)*; Cyperaceae (sedge) pollen seems to dominate the top of the underlying lacustrine sediment. The abundant spruce pollen is characteristic of many Two Creeks pollen diagrams (for example, West, 1961; Schweger, 1969; Maher, 1970).

The Murphy Concrete & Construction pit in which we stand lies at the southwest end of a prominent ridge (fig. 4.2) and consists of 10 m of sand and gravel interspersed with thin layers of finer-grained sediment with clasts (fig. 4.7A). The pit was capped by red till, which was recently removed. We have interpreted the sand and gravel sequence as proglacial outwash with interspersed debris flows (fig. 4.7B). It is unclear whether this sequence could have been deposited subaqueously as the Green Bay Lobe readvanced into glacial Lake Oshkosh or whether the deposition occurred subaerially. It is also unclear when this sequence was deposited. We interpreted the red till that capped this pit as Middle Inlet till that was deposited subglacially during the last readvance of the ice lobe.

On the basis of the geology and morphology of the Duck Creek ridges, it is difficult to assess whether they were formed during the last readvance of the Green Bay Lobe, which overran the Two Creeks forest bed, or whether they were formed by numerous ice advances during the last glaciation. Thwaites (1943) interpreted the ridges to be eskers; because eskers typically are sinuous and composed of sand and gravel, it is difficult to accept this interpretation using existing geological information. Piette (1963) determined that the ridges were crevasse fill deposits. However, it is unclear how longitudinal crevasses would extend through several hundred meters of actively flowing ice. An alternative interpretation is that the ridges are drumlins. Because the formation of drumlins is poorly understood, it is difficult to assess whether the ridges were formed by the glacier eroding the surrounding glacier bed or whether the ridges are depositional features.

Some have suggested that large drumlins (mega-lineations) such as the Duck Creek ridges are the result of ice-streaming or fast-flowing ice (for example, Stokes and Clark, 2001, 2002). Although this may have been the case, this argument assumes that elongation is controlled by the magnitude of ice sliding past bedforms rather than by the overall strain field of the basal ice, which is difficult to determine for Pleistocene ice masses (Iverson and Hooyer,

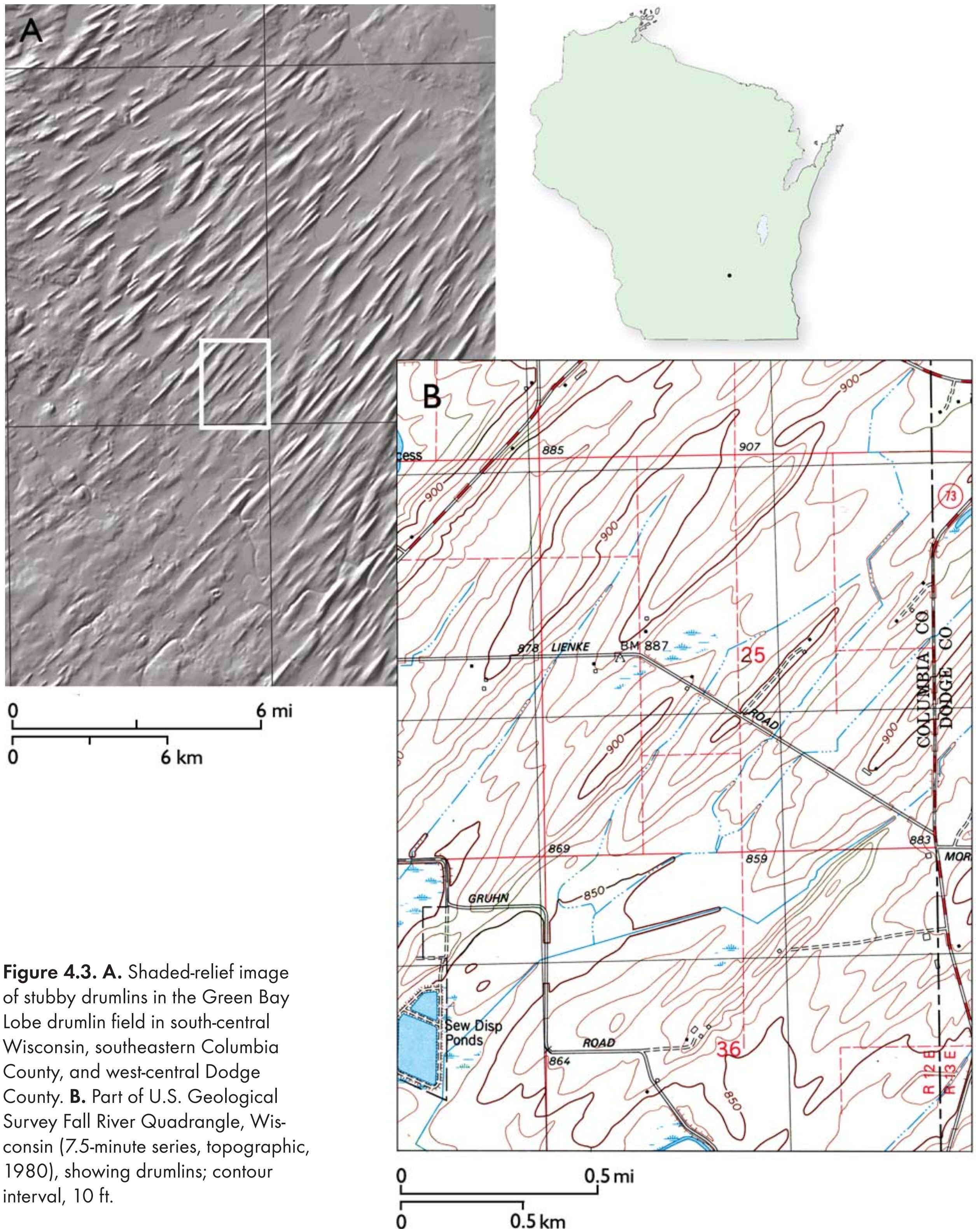


Figure 4.3. A. Shaded-relief image of stubby drumlins in the Green Bay Lobe drumlin field in south-central Wisconsin, southeastern Columbia County, and west-central Dodge County. **B.** Part of U.S. Geological Survey Fall River Quadrangle, Wisconsin (7.5-minute series, topographic, 1980), showing drumlins; contour interval, 10 ft.

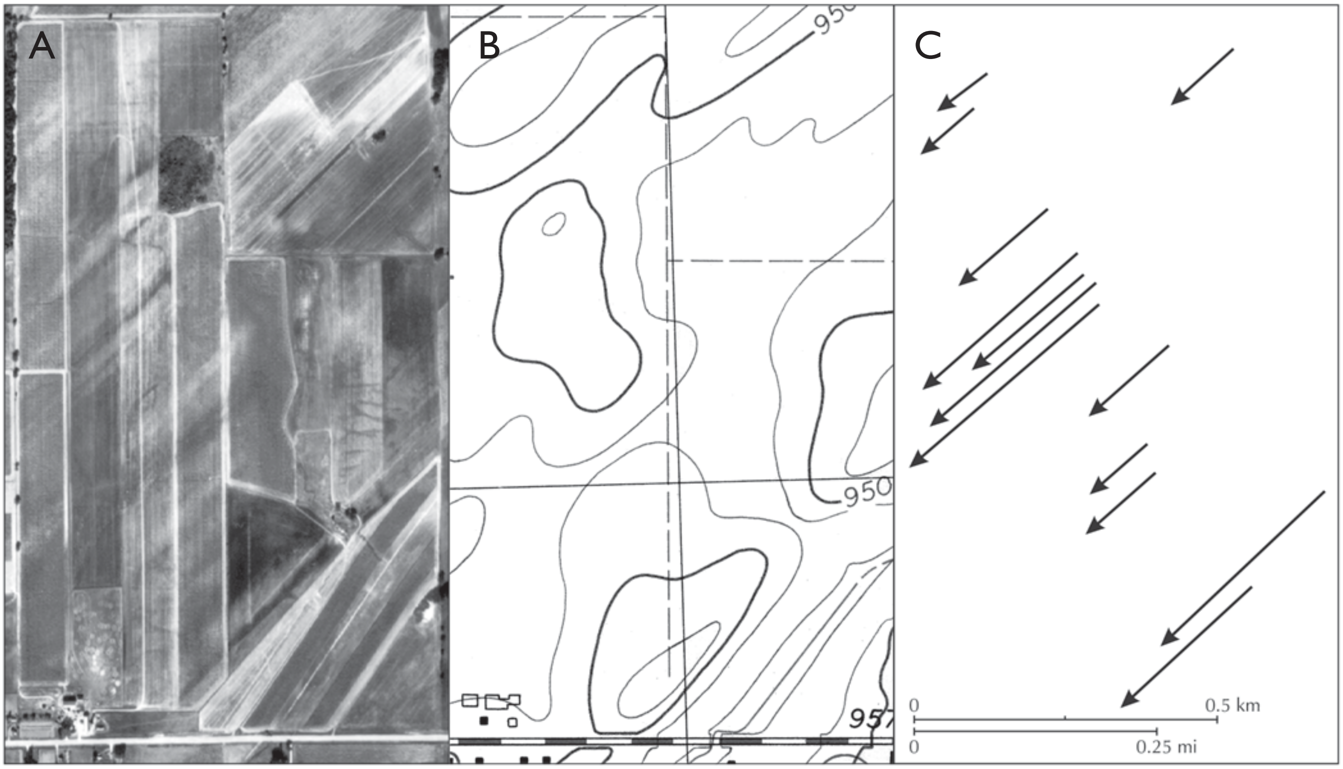


Figure 4.4. Aerial photographs **(A)** of drumlins in northeastern Dane County that are too small to be recognized on **(B)** the U.S. Geological Survey Marshall Quadrangle, Wisconsin (7.5-minute series, topographic, 1981), with 10-ft contour intervals. Sketch map of the same area **(C)** shows the more conspicuous drumlins. (U.S. Department of Agriculture aerial photograph WU-6P-5.)

2004). This argument also assumes that the magnitude of displacement at the glacier bed depends primarily on sliding speed rather than on the duration of glacier occupation. Even if these assumptions are correct, bedform elongation does not provide a quantitative estimate of velocity, making any interpretation that the ridges are the result of ice streaming uncertain.

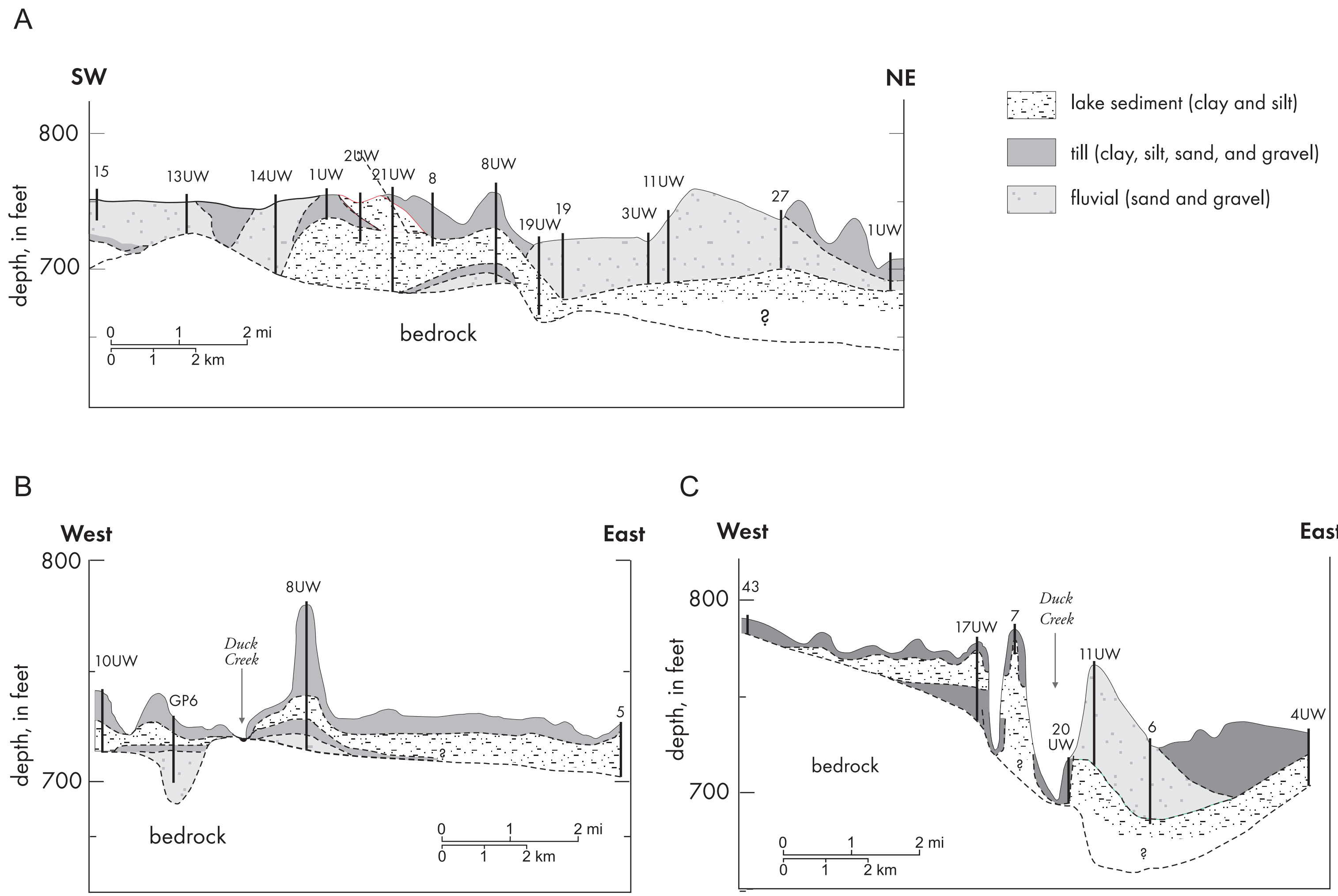


Figure 4.5. Longitudinal **(A)** and transverse **(B, C)** cross sections of the Duck Creek ridges shown in figure 4-1 (modified from Piette, 1963).

Figure 4.6. Photographs showing **(A)** a 3 m thick layer of deformed sand at the apex of a ridge in the Smith pit (fig. 4.2), and **(B)** deformed pods of sand with rotational structure (height of photograph approximately 1 m).

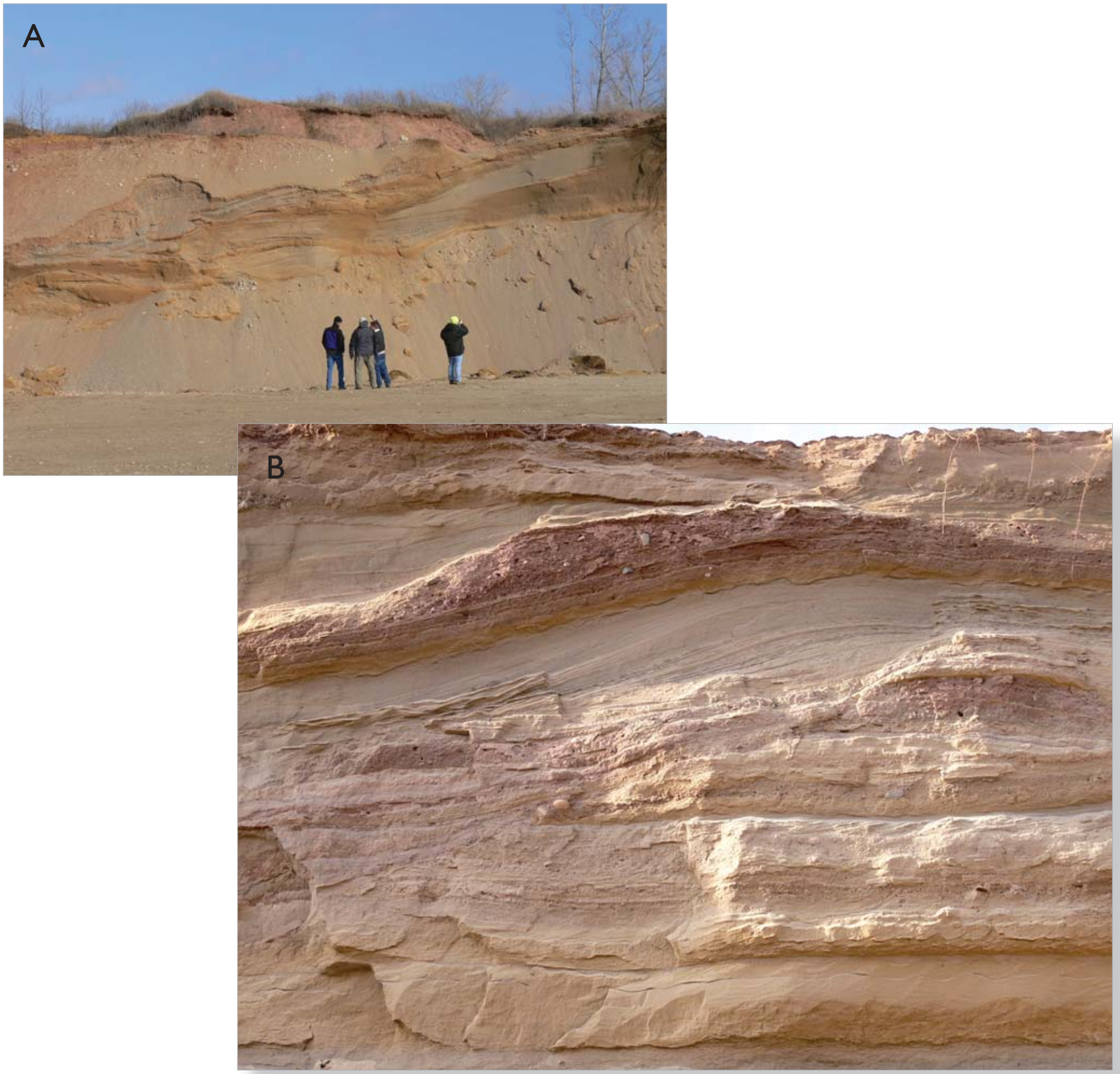


Figure 4.7. A. Photograph of sand and gravel sequence capped by red fine-grained till. **B.** Photograph showing two layers of till within the sand and gravel sequence.

Stop 5

Two Creeks forest bed, Ebben Quarry, Brown County

Location: SW¼NE¼ sec. 27, T23N, R198E, Brown County (fig. 5-1). Stop is in a rock quarry owned by Daanen and Janssen, Inc. Entrance to the quarry is off Highway EE; permission is required for entrance.

Authors: Thomas S. Hooyer and D.M. Mickelson

The Two Creeks forest bed has been a well known stratigraphic marker in east-central Wisconsin since it was described 100 years ago (Goldthwait, 1907). This bed is of interest to scientists because it documents a period of time when the Laurentide Ice Sheet was receding and the climate was changing dramatically. The presence of a boreal forest represents a warming climate that a readvancing Green Bay Lobe seems to belie. The forest was regionally extensive—wood fragments and other organic material have been found in at least 11 counties (see Mickelson and others, this volume). Extensive geologic mapping in the region has indicated that the forest developed soon after the recession of a major readvance (Middle Athelstane) of the Green Bay Lobe, which deposited the tills of the Kirby Lake, Chilton, and Valders Members of the Kewaunee Formation.

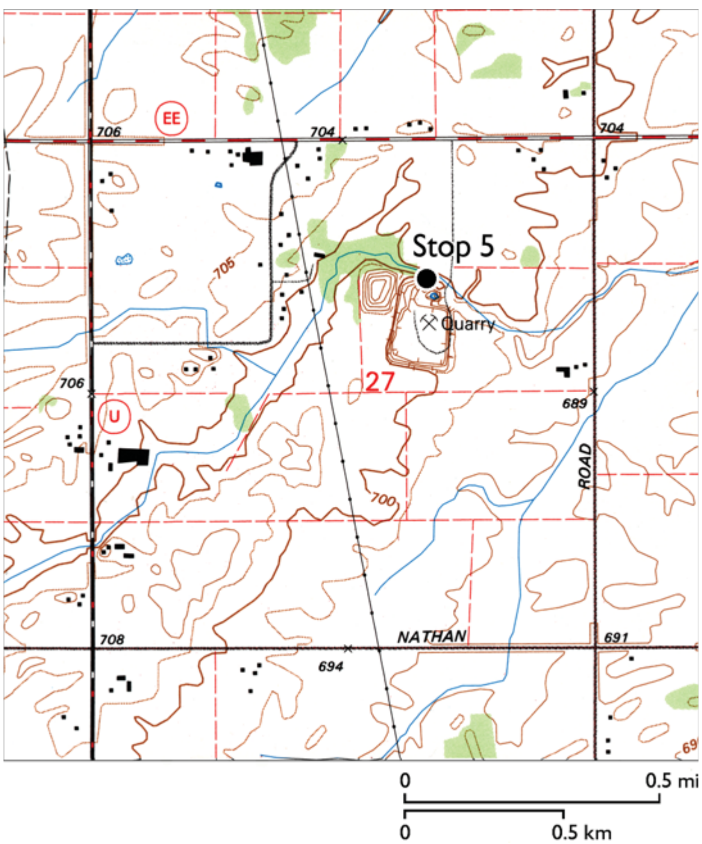


Figure 5.1. Part of U.S. Geological Survey Oneida South Quadrangle, Wisconsin (7.5-minute series, topographic, 1992), showing the location of the Ebben Quarry (stop 5).

Glacial Lake Oshkosh must have completely drained because the forest bed has been discovered in the low-lying areas of the Fox River lowland, indicating that the ice must have receded northward of Sturgeon Bay in the Door Peninsula. The Green Bay Lobe eventually readvanced up the Fox River lowland. The ice extended as far as the city of Appleton, where it deposited a broad moraine. In areas north of this ice margin (up ice), the forest bed is usually buried beneath a layer of red till that consists primarily of reworked lake sediment of the Middle Inlet, Glenmore, and Two Rivers Members of the Kewaunee Formation; in low-lying areas south and west of this ice margin, the forest bed is usually buried beneath sediment of glacial Lake Oshkosh. In upland areas not covered by the lake or the ice, there remains little to no evidence of the forest bed. Eventually, the ice receded northward, allowing reactivation of lower outlets of glacial Lake Oshkosh (see introductory paper, this volume; Mickelson and others, this volume).

Dates for wood from the Two Creeks forest bed ranged from about 14,400 to 13,100 cal yr BP (12,400 to 11,200 ^{14}C yr BP), with an average of

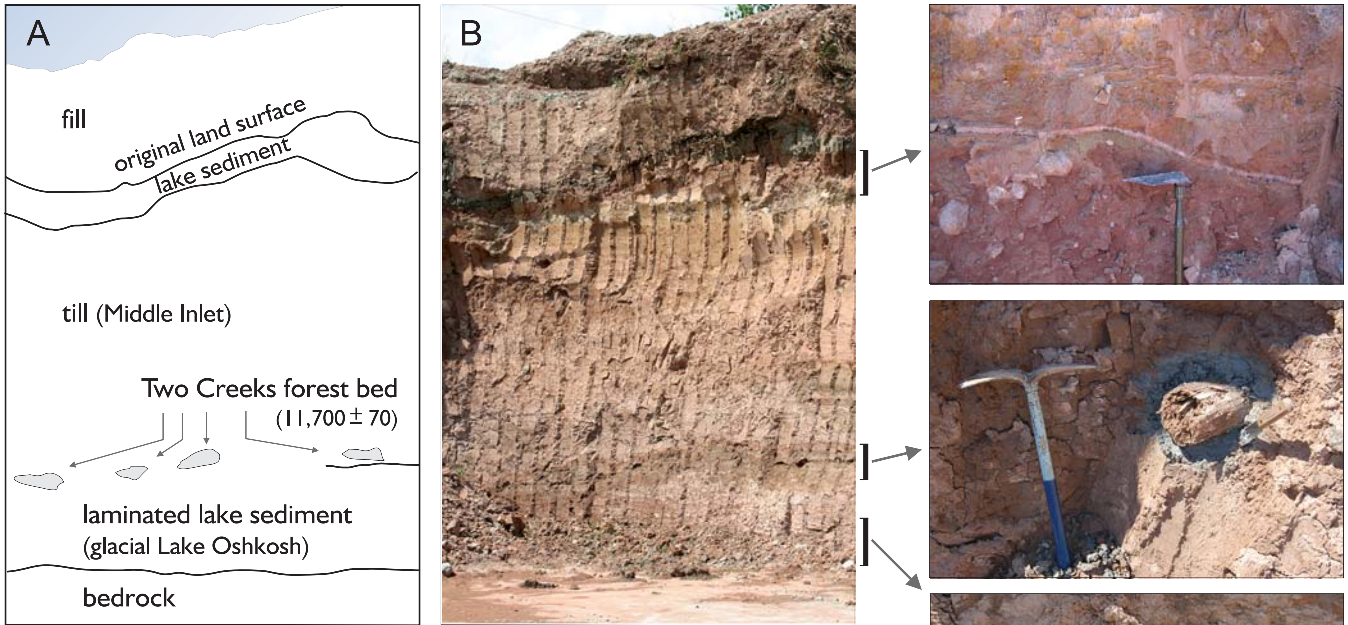


Figure 5.2. Schematic (A) and photographs (B) showing the interpretation and sequence, respectively, of the 6 m tall exposed section.

13,800 cal yr BP (11,850 ^{14}C yr BP (Broecker and Farrand, 1963; Black and Rubin, 1967-68; Kaiser, 1994; Mickelson and others, this volume). Studies of buried trees from the forest bed showed that the longest record is 234 annual rings (Kaiser, 1994). More recent work showed an almost 1,000-year record of tree rings from about 14,000 to 13,000 cal yr BP (Leavitt and others, 2006; Leavitt and Panyushkina, this volume).

The Ebben Quarry is one of many locations in which wood and organics from the Two Creeks forest bed are present (fig. 5.1). In this quarry the bed lies just above the laminated lake sediment at the base of the overlying red till (fig. 5.2). The till is capped by more lake sediment in which the modern soil has developed. The laminated lake sediment at the bottom of the section lies on bedrock; it was probably deposited in glacial Lake Oshkosh as the ice receded from the lowland just prior to the growth of the Two Creeks forest. Immediately above the lake sediment is 4 m of reddish brown, fine-grained till of the Middle Inlet Member. Incorporated into the till at its basal contact are dispersed organics, including pieces of wood. Three pieces of wood removed from this layer have been identified as spruce *(Picea);* one sample was dated to 13,550 ±90 cal yr BP (11,700 ±70 ^{14}C yr BP), typical of Two Creek dates.

Although there is not a continuous layer of forest litter at the contact between the lake sediment and till, the outcrop indicates that either the Green Bay Lobe overran the forest at this location or the forest was killed by rising waters of glacial Lake Oshkosh before the site was covered with ice. Immediately above the till, the lake sediment ranges from 1 to 2 m thick and contains interesting layers of carbonate-enhanced clay-rich sediment. The carbonate in these layers most likely precipitated from percolating water as its velocity was reduced by layers containing finer-grained sediment. The lake sediment that forms this unit was probably deposited in glacial Lake Oshkosh as the ice sheet receded northward from this location.

Stop 6

Prograding delta along the eastern edge of glacial Lake Oshkosh, Tower Hill gravel pit, Brown County

Location: SW¼NW¼ sec. 6, T22N, R21E, Brown County (fig. 6.1), in a gravel pit owned by Bernard Van Straten. Entrance to the quarry is off Highway X or through the Van Straten's yard; permission is required for entrance.

Authors: D.M. Mickelson and William N. Mode

The Tower Hill gravel pit is in a till-capped gravel terrace that extends several miles southwest along the edge of the Silurian escarpment (fig. 6.1). Sand and gravel in the terrace has been mined for many years. Just south of Highway X is the Brown County East Landfill, and between the pit and Highway X is a smaller paper-mill waste landfill. The demolition debris landfill just east of the gravel pit entrance covers an exposure of the Two Creeks forest bed. The overall stratigraphy of the site is shown in figure 6.2.

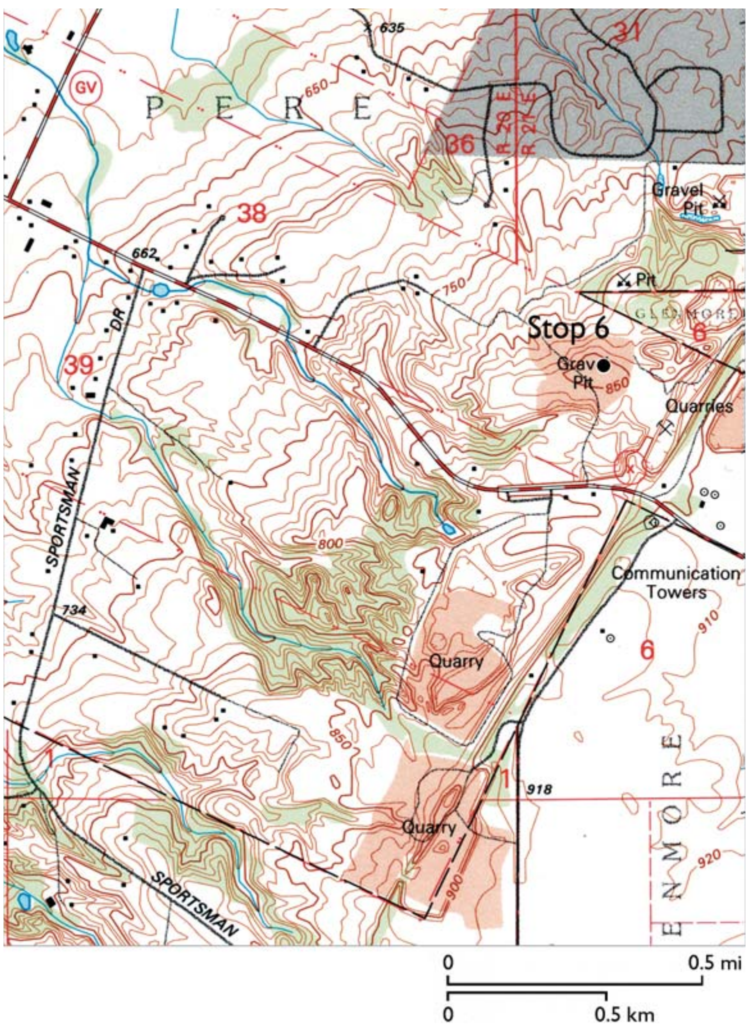


Figure 6.1. Part of U.S. Geological Survey De Pere Quadrangle, Wisconsin (7.5-minute series, topographic, 1971), showing the location of the Tower Hill gravel pit (stop 6).

Borings for the landfill south of Highway X indicate that below the sand and gravel is red, clayey diamicton above a more gravelly, grayish-brown diamicton, which lies on bedrock. We assume that these layers are the Chilton Member of the Kewanee Formation and the Holy Hill Formation, respectively, but we have seen no samples and only have driller's logs to demonstrate this. Above the clayey diamicton is sand that is approximately 20 m thick. We assumed that this sand unit is the sand that can be seen near the base of the pit at this stop. The remainder of this description is based on observations during the past 20 years, and not all of the features can be seen at this time.

The northern end of the pit contains ripple-laminated sand near the base (fig 6.3). The beds are mostly tabular and dip gently to the southwest. Climbing ripples and small cross-bed sets indicate water flow toward the southwest and fairly rapid sedimentation. We have interpreted this unit to have been deposited in a shallow-water river–lake environment (Ashley, 1988). This sediment contains small balls of reddish-brown diamicton, apparently derived from the underlying Chilton till. Above this is an abrupt

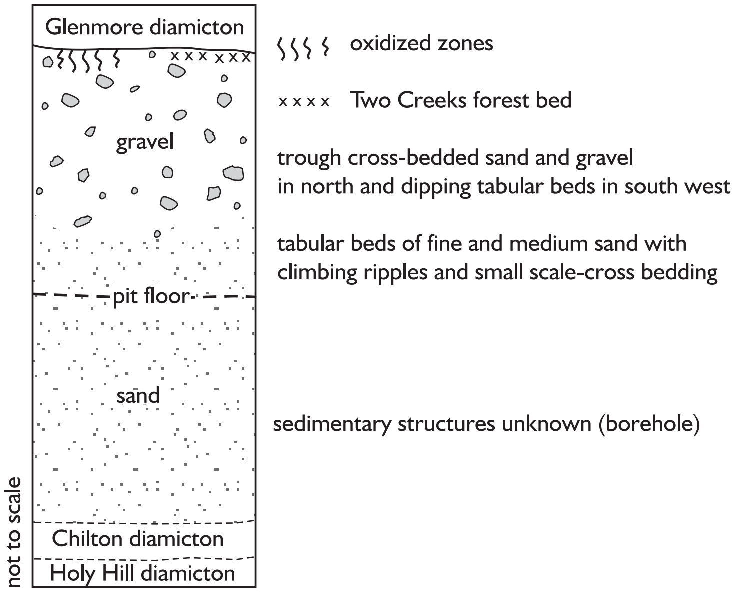


Figure 6.2. Schematic showing the stratigraphy at the Tower Hill gravel pit. Note that this is a sketch and that it is not to scale. Features shown are in different parts of the deposit.

transition to trough-bedded sand and gravel. Trough orientations also indicate flow direction toward the southwest. We have interpreted this change in the nature of the sediment upward in the section as a change from a river–lake environment to a braided stream. In the southwest corner of the pit, beds of sand and fine gravel dip toward the southwest at a low angle; we have interpreted these to be foreset beds. Near the northwest corner of the pit, a layer of brown, sandy till of the Holy Hill Formation and a layer of red clayey till of the Chilton Member appear to have been sheared into the stratified sequence. The layer dips toward the northwest.

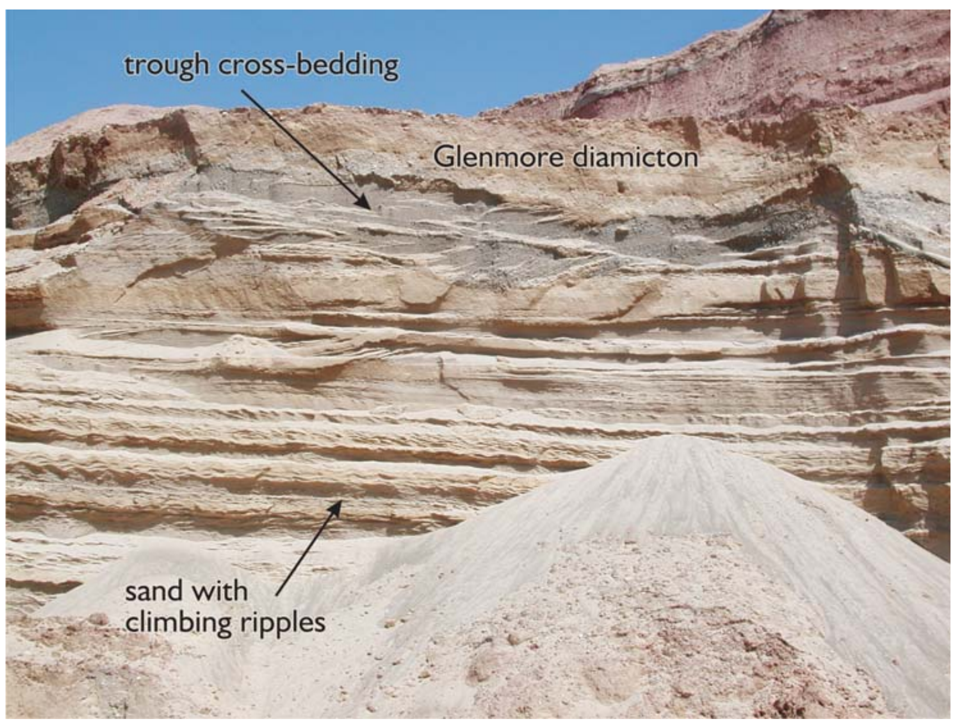


Figure 6.3. Climbing ripples in sand in the lower part of the Tower Hill gravel pit. Photograph taken in 2004 in north-central part of pit, 6 m above base. The exposure is approximately 10 m high.

The Two Creeks forest bed was exposed for a number of years in the southeast corner of the pit. It included a thin layer of spruce needles, wood and what appeared to have been a rooted stump. Above the forest bed layer was approximately 3 m of Glenmore till. This till is now exposed above sand and gravel along the north side of the pit, but we have no evidence of the Two Creeks buried forest in that area. Presumably, it was removed by ice of the Glenmore advance.

The stratified sequence is a long, narrow delta built between the ice and the Silurian escarpment. Foreset beds in the southwest part of the pit were deposited in a deeper part of the lake; the lake shallowed toward the northeast. The delta prograded toward the southwest. Sediments exposed near the base of the pit in the northeast corner were deposited in a shallow river–lake environment. As sediment filled the lake, the water depth became shallower until a braided stream flowed across the surface, depositing the sand and gravel in the upper part of the sequence. Deposition of the stratified sequence had to have taken place during recession of the Chilton ice, if our interpretation of Chilton till beneath the stratified section is correct. It cannot have been deposited during the advance or recession of Glenmore ice because the Two Creeks buried forest lies stratigraphically above it.

Was the stratified section at the Tower Hill pit deposited in glacial Lake Oshkosh? The top of the stratified sequence is at an elevation of approximately 800 ft (244 m). The thickness of the upper gravel is about 6 m, indicating the transition from shallow lake to braided stream was between elevations of approximately 800 and 820 ft (247 and 253 m). If this delta had been deposited in glacial Lake Oshkosh, it must have been built when the Manitowoc River outlet was functioning because that was the lowest ice-free outlet. (The Neshota outlet is approximately 8 km farther northeast and was presumably blocked by ice at the time the stratified sequence was deposited.) The present elevation of the floor of the Manitowoc outlet is between about 810 and 820 ft (247 and 250 m). Thus, it is only approximately 3 m below the water depth indicated in the Tower Hill pit. Given that the Tower Hill pit is approximately 32 km northeast of the Manitowoc outlet, differential rebound could easily explain the present difference in elevation. In addition, there would have been some water in the channel, so the water surface would be expected to be higher than the channel bottom. Thus, it seems likely that the stratified sequence at the Tower Hill pit was built into glacial Lake Oshkosh when it was using the Manitowoc River outlet during recession of the Chilton ice.

Stop 7

Neshota outlet and drainage spillway associated with glacial Lake Oshkosh, Denmark Quarry, Brown County

Location: NW¼SE¼ sec. 36, T22N, R22E, Brown County (fig. 7.1). Stop is in a gravel pit adjacent to a rock quarry; both are owned by Payne and Dolan. Permission is required to enter the gravel pit, which lies directly north of the rock quarry along Cooperstown Road.

Authors: James A. Clark, Thomas S. Hooyer, and Lee Clayton

The Neshota outlet (fig. 7.1), named for the underfit river that now occupies the former drainage channel, was one of five outlets that drained glacial Lake Oshkosh (see introductory paper, this volume). Four of the outlets are in east-central Wisconsin and cut across the Silurian escarpment. The existing elevation of these outlets decreases progressively to the north. All the channels are deeply incised into the surrounding landscape, which consists primarily of dolomite covered with a thin layer of glacial sediment. Two conspicuous drainage channels are associated with the Neshota and Kewaunee outlets (fig. 7.2). Because the topography of this area is dominated by bedrock, incision of these channels probably occurred over numerous glacial cycles, most recently during the late Pleistocene by the draining and filling of glacial Lake Oshkosh as the Green Bay Lobe fluctuated.

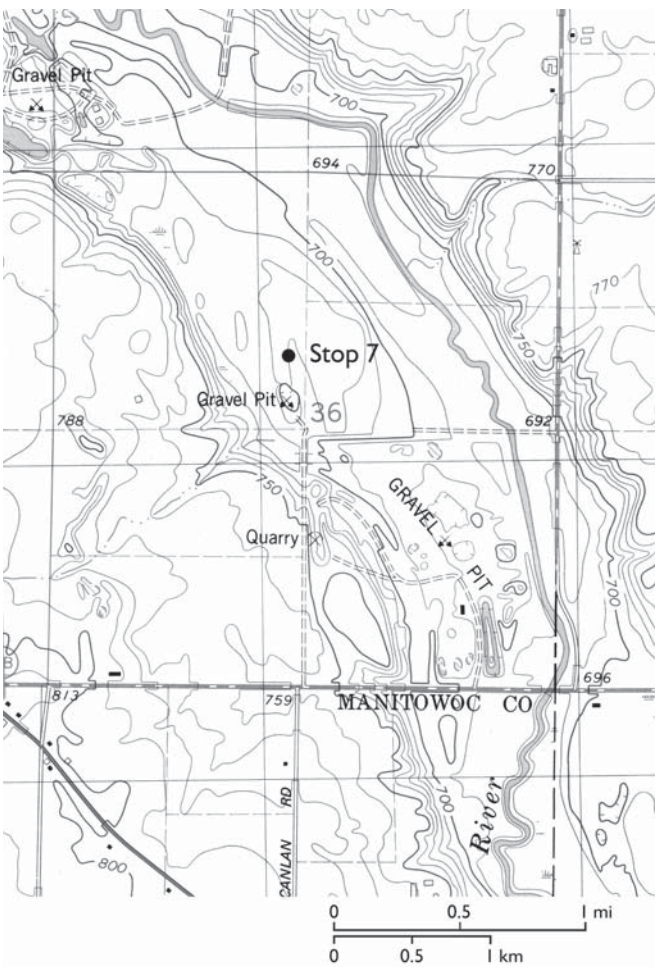


Figure 7.1. Part of U.S. Geological Survey Denmark Quadrangle, Wisconsin (7.5-minute series, topographic, 1978), showing the location of the Payne and Dolan gravel pit (stop 7) within the Neshota drainage channel.

To determine the changing elevations of the outlets and the area and volume of the lake during various phases, we used a numerical model to calculate isostatic adjustment of the land surface due to glacial loading and unloading as a function of time (see introductory paper, this volume). Model results showed that just prior to the opening of the Neshota outlet, the area of glacial Lake Oshkosh was approximately 5,500 km^2, with a mean depth of approximately 13 m and a volume of 70 km^3. Clearly, glacial Lake Oshkosh was a relatively shallow lake. With the opening of the Neshota outlet, base level dropped approximately 20 m, resulting in a volume reduction of about 50 percent. The area of the lake was also significantly reduced to a third of its original size (1,778 km^2). Although it is difficult to assess how quickly the lake drained through the outlet, it must have been quite dramatic, given the incision observed in the channel today. If the lake drained quickly, it could have transported

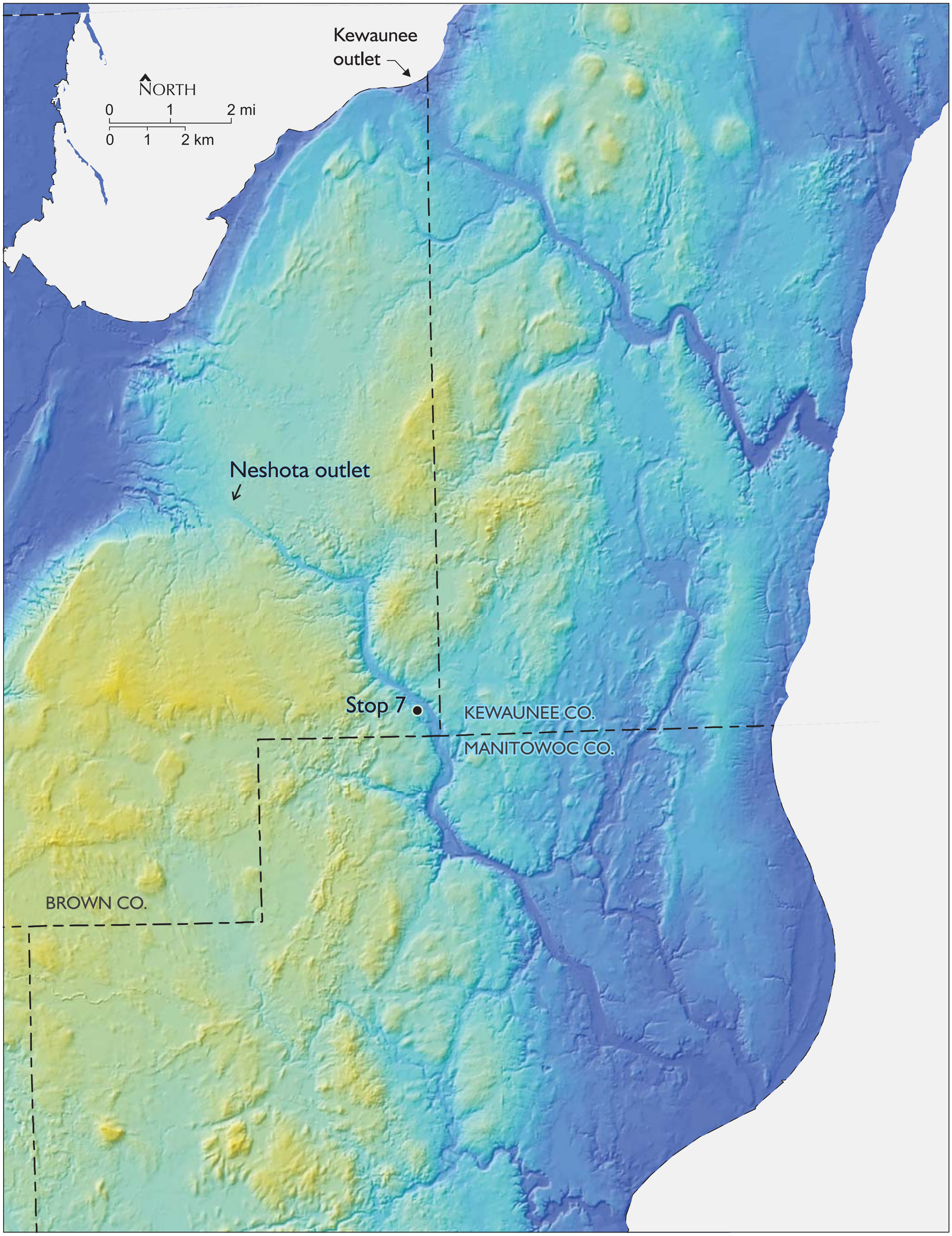


Figure 7.2. Shaded-relief image showing the Neshota and Kewaunee outlets and associated drainage channels.

Figure 7.3. Photograph of exposure showing the coarse-grained nature of the deposit.

large volumes of sediment downstream.

The coarse sediment carried down the Neshota channel by the discharge from glacial Lake Oshkosh can be observed in a gravel pit approximately 13 miles downstream of the Neshota outlet (fig. 7.3). The exposure consists of 2 to 4 m of sand, gravel, and boulders of various sizes. We interpreted this sequence to have been deposited during the late Pleistocene, when there were large discharges down the spillway. The relatively small size and low gradient of the existing Neshota River would not be capable of transporting such a large bed load.

Given the coarse nature of the deposit and the known volume of water that must have drained through the Neshota outlet, we can predict the magnitude of flow through the drainage channel. As glacial Lake Oshkosh drained rapidly from a higher to a lower stage, lake volume *(V)* changed (fig. 7.4A). The empirical relationship for lake stage *(s)*, $s = g(V)$, will differ for every lake and is dependent on how much of the lake basin is filled with ice. For example, when the ice sheet uncovered the Neshota outlet, lake stage dropped from 771 ft to 705 ft (235 m to 215 m) and lake volume decreased about 34 km^3.

For the existing channel configuration, we calculated discharge using the Hydrologic

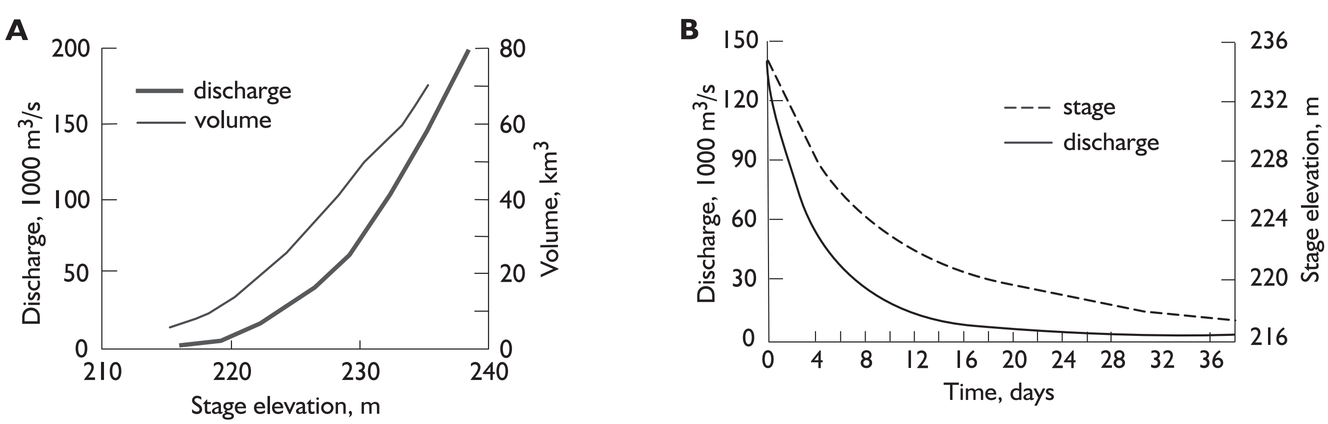


Figure 7.4. A. Empirical relationship between lake volume, discharge, and stage elevation at 13,300 cal yr BP as calculated by the HEC–RAS model for the Neshota outlet and channel. **B.** Time-dependent change in discharge and stage elevation as a function of time calculated from a numerical solution.

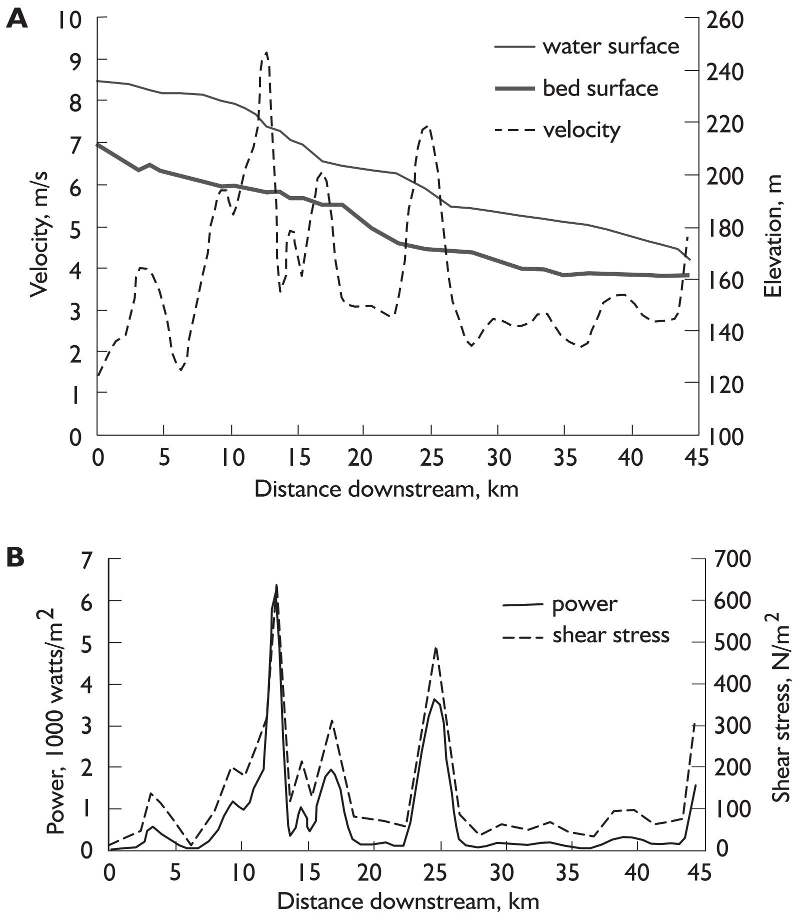


Figure 7.5. The HEC–RAS model calculations for **(A)** water-surface elevation and water velocity and **(B)** stream power and shear stress when the Neshota outlet first became active at 13,300 cal yr BP. Distances are measured downstream from the Neshota outlet. Bed-surface elevation in part A is defined by the digital elevation model corrected for isostatic rebound.

Engineering Center–River Analysis System (HEC–RAS) model of the Hydrological Engineering Center of the Army Corps of Engineers (2005). The HEC–RAS model was designed to be a floodplain-management tool that calculates stage elevation for a prescribed discharge, but we used it to calculate the discharge, Q, for any stage height prescribed in the uppermost reach of the river valley. Required HEC–RAS input includes topographic cross sections of the river valley at representative reaches. We also assumed that the existing land surface, corrected for isostatic rebound, contained the Neshota channel as it exists today. This assumption implies that much of the channel erosion occurred during the initial phases of the flood shortly after the outlet was opened. With this assumption, the relationship between discharge and stage over time (t) can be calculated (fig. 7.4B):

$$Q = dV/dt = f(s)$$

Substitution of the lake volume relationship into the discharge relationship yielded an ordinary differential equation, $dV/dt = f(g(V))$, that can be solved readily with numerical methods for the time-dependent lake volume, V, using the empirical relationships of figure 7.4A. The resulting lake volume is then known through time, $V(t)$, and hence the lake stage, $s(t)= g(V(t))$. The Neshota outburst event is calculated to have lasted only about 10 days with a peak discharge of 140,000 km^3/s (fig. 7.4B).

For a given channel, HEC–RAS can also be used to calculate stream parameters that are useful in characterizing a flood event. River depth, velocity, shear stress at the bed, and stream power are of the most interest because they control transport and deposition of coarse sediment load. Figure 7.5 shows how dramatically these variables change along the channel profile during initial flood times. Where velocity and stream power are high, boulders as large as 3 m in diameter may be transported (O'Connor, 1993). Deposition of this very coarse load would occur in adjacent reaches with lower velocity and stream power. Thus, the boulders observed at this stop could have easily been transported in a flow event that drained glacial Lake Oshkosh through the Neshota outlet.

Stop 8

Evidence of tundra plants overridden by ice approximately 16,000 cal yr BP, Sherwood, Wisconsin, Calumet County

Location: SW¼SE¼ sec. 30, T20N, R19E, Calumet County (fig. 8.1). Stop is in a quarry owned by J & E Construction, south of Clifton Road, approximately 0.8 km west of the intersection of Highways 114 and 55 in Sherwood, Wisconsin. Permission is required to enter the quarry.

Author: Betty J. Socha

At the J & E Construction quarry, operations have exposed about 4.5 to 6 m of glacigenic sediment in a small end moraine. The uppermost sediment is reddish-brown clay-rich till of the Chilton Member of the Kewaunee Formation. In some places, the Chilton till is underlain by up to 1.2 m of gray silt of the High Cliff Member of the Hayton Formation; this site is the type section for the High Cliff. Underlying this silt at some places in the quarry are thin patches of grayish brown silty-sand till of the Cato Falls Member of the Hayton Formation. In other parts of the quarry, the Chilton till is underlain by yellow-brown sandy till of the Horicon Member of the Holy Hill Formation.

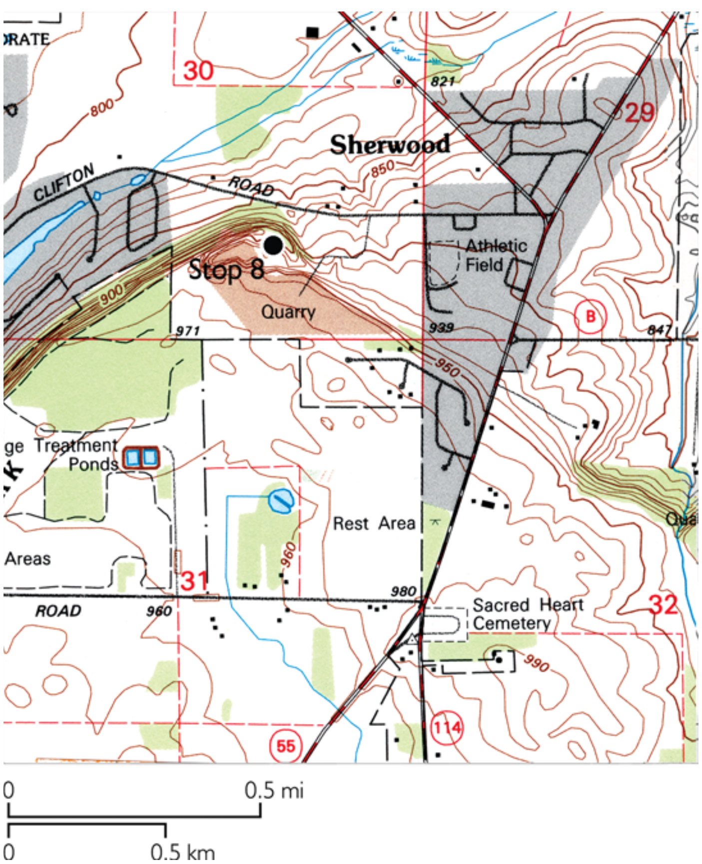


Figure 8.1. Part of U.S. Geological Survey Sherwood Quadrangle, Wisconsin (7.5-minute series, topographic, 1992), showing the location of the J & E Construction quarry (stop 8).

Organic remains from this site indicate that a tundra environment existed before the ice of the Green Bay Lobe readvanced over the area and deposited reddish-brown clay-rich Chilton till, approximately 16,000 cal yr BP (13,500 ^{14}C yr BP). This date is one of a few recently obtained in east-central Wisconsin that help constrain a major readvance of the Green Bay Lobe and document relatively cold environmental conditions.

Chilton Member of the Kewaunee Formation

The upper part of the Chilton till is fractured, leached of carbonate, and partially oxidized to brown and yellowish brown. The underlying reddish-brown (generally 5YR 4/4), clay-rich diamicton (predominantly basal till), is generally massive, but contains a few laminae and stringers of silt (fig. 8.2). The matrix of the five till samples analyzed was 9 to 25 percent sand, 44 to 63 percent silt, and 22 to 39 percent clay. Locally, clast concentrations and stone lines are visible. Dispersed in the Chilton till, and also associated with the gray silt stringers, are small pieces of plant remains. Pebble-fabric analysis indicated a strong

Figure 8.2. Photograph of gray stringer of fine sand and silt in reddish-brown Chilton till. Lower gray layer is silt and fine sand with dispersed organics and silty-sandy diamicton. Bedrock is vuggy dolomite and has a polished and striated surface. Scale is 16.5 cm long.

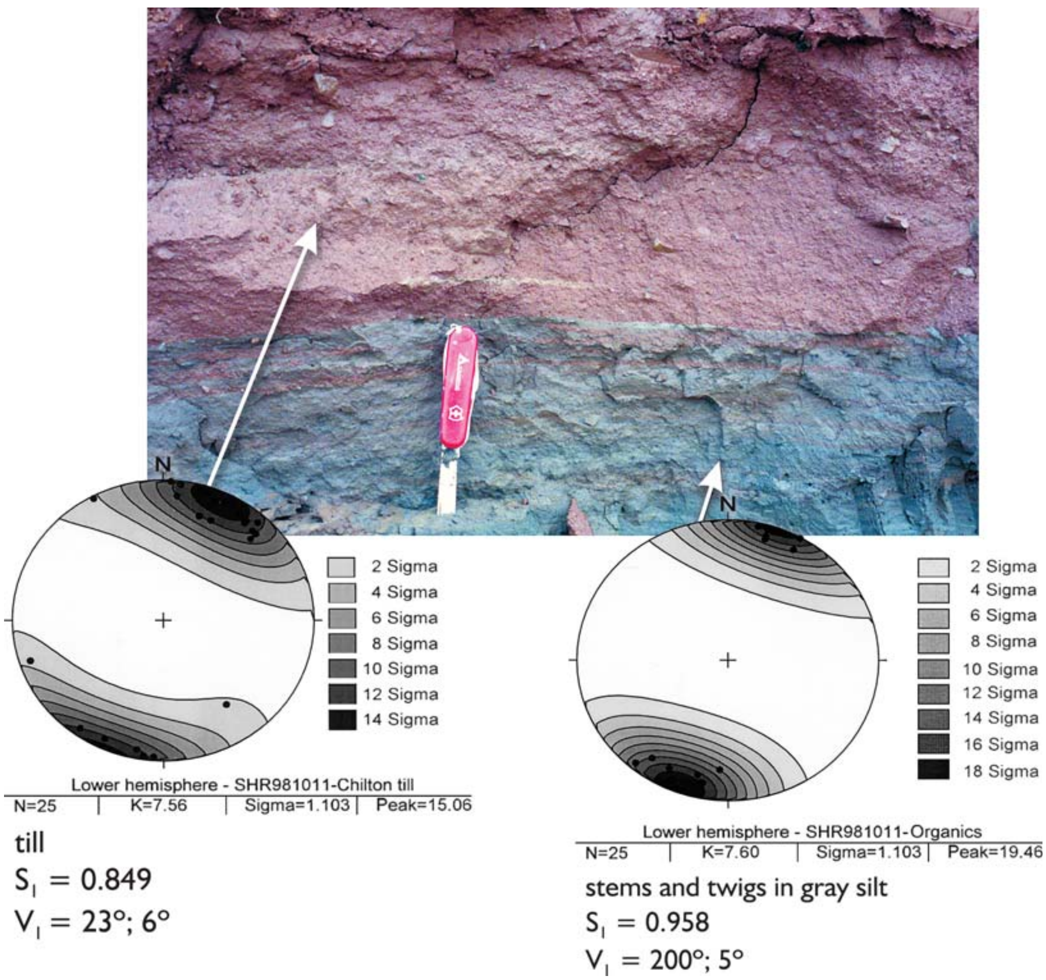


Figure 8.3. Photograph showing strong pebble fabric in uniform, reddish-brown clayey diamicton (Chilton till) and strong fabric of stems and twigs in the underlying laminated gray silt (High Cliff Member). S_1 is the eigenvalue, V_1 is the eigenvector. Knife handle is about 10 cm long.

Figure 8.4. Photograph of mat of moss, stems, and leaves on stoss side of bedrock bump. Knife handle is 10 cm long.

preferred orientation of clast long axis (fig. 8.3); ice-flow direction was generally from the northwest, but varied from northwest to northeast at this location.

Horicon Member of the Holy Hill Formation

The pale brown (10YR 6/3), silty-sand diamicton of the Horicon Member is generally massive or crudely stratified basal till. Two samples of the matrix of the Horicon till from this location were 49 percent sand, 40 percent silt, and 10 percent clay; and 48 percent sand, 37 percent silt, and 15 percent clay, respectively. The till is composed mainly of dolomite in all size fractions, indicating a predominantly local origin. Igneous and metamorphic rocks are present, but are not as abundant as in the till of the Hayton Formation.

High Cliff Member of the Hayton Formation

The gray silt of the High Cliff Member has an interbedded contact with the Chilton till (fig. 8.2). The silt varies from massive to diffusely laminated. The gray silt is primarily eolian, deposited by wind near the receding ice margin. In places, the gray silt may have been deposited in water or slumped into shallow ponds.

Dispersed in the gray silt are small pieces of plant remains: leaves, stems, mosses, and twigs. Locally, mats of plant remains have been found on the gray silt (fig. 8.4). The plant taxa indicate a cold, arctic environment. There was much moss and no evidence of trees; *Dryas integrifolia* leaves are abundant, and seeds of *Silene acaulis* are common (Richard Baker, written communication, 1998). Barrens and cliffs are the general habitat for *D. integrifolia* and *S. acaulis* (Maher and others, 1998). The plant material is dated at 15,890 $\pm$210 cal yr BP (13,370 $\pm$90 ^{14}C yr BP); these plants were overridden by the Chilton ice advance.

Cato Falls Member of the Hayton Formation

At the quarry, till of the Cato Falls Member generally fills in low areas in the dolomite surface. Striations on the bedrock and pebble-fabric analysis indicate ice-flow direction from the north. The matrix of one sample of Cato Falls till was about 38 percent sand, 39 percent silt, and 22 percent clay.

Regionally, the till is grayer and has more silt in the matrix than the till of the Holy Hill Formation. The Hayton till is grayer, and has more sand than the reddish-brown clayey tills of the Kewaunee Formation. The till is generally thin along the Silurian Escarpment; it is at least 24 m thick at the type section at Valders Quarry in central Manitowoc County.

Stop 9

Late-glacial and early Holocene paleoecology: Schneider farm, Calumet County

Location: SW¼SE¼ sec. 29, T19N, R20E, Calumet County (fig. 9.1). Stop is north of Lau Road in a field owned by the Schneider family. Access to the site is by a small dirt road behind the farm house and barn. Permission is required for entry.

Authors: William N. Mode, Irina P. Panyushkina, Steven W. Leavitt, John W. Williams, Adianez Santiago, Jacquelyn Gill, Cole Edwards, and Holly Gertz

Pond sediment and peat exposed in a farm-pond excavation (fig. 9.2) on the Schneider farm in Calumet County (fig. 9.3), east-central Wisconsin, contain abundant plant macrofossils, including many trunks and branches of trees that preserve beaver tooth marks (fig. 9.4). Sediment began accumulating at the site about 14,000 cal yr BP (12,000 ^{14}C yr BP) (table 9.1) in a small (approximately 30 m by 50 m) kettle. Sediment continued accumulating until some time after about 10,740 cal yr BP (9,500 ^{14}C yr BP). An unknown amount of this material, probably no more than 1 m, had been excavated before we visited the site, so we have not yet obtained dates for the uppermost sediment in the excavation.

Approximately 185 large pieces of wood, mainly spruce *(Picea)* and tamarack (larch, *Larix)* collected from the site (145 pieces in situ and 40 pieces as float) are being analyzed at the Laboratory of Tree-Ring Research. At two levels in the sediment column, tree trunks and branches were woven together in a structure that resembled a beaver dam (fig. 9.5). In other parts of the exposure, concentrations

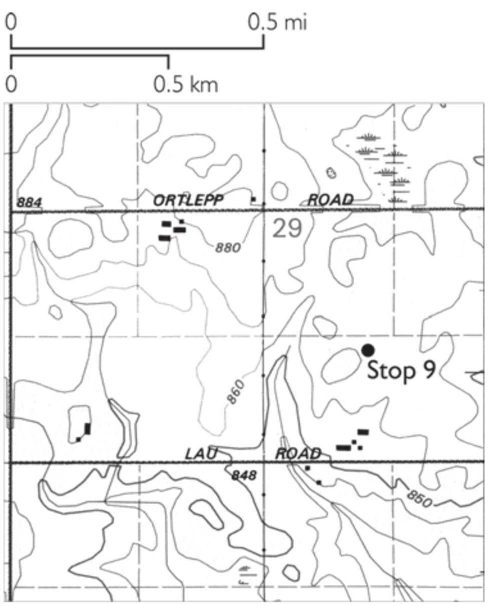


Figure 9.1. Part of U.S. Geological Survey Chilton Quadrangle, Wisconsin (7.5-minute series, topographic, 1992), showing the location of the Schneider farm (stop 9).

Figure 9.2. Low-level aerial photograph of Schneider farm pond excavation. (Photograph courtesy of L.J. Maher and D.M. Mickelson.)

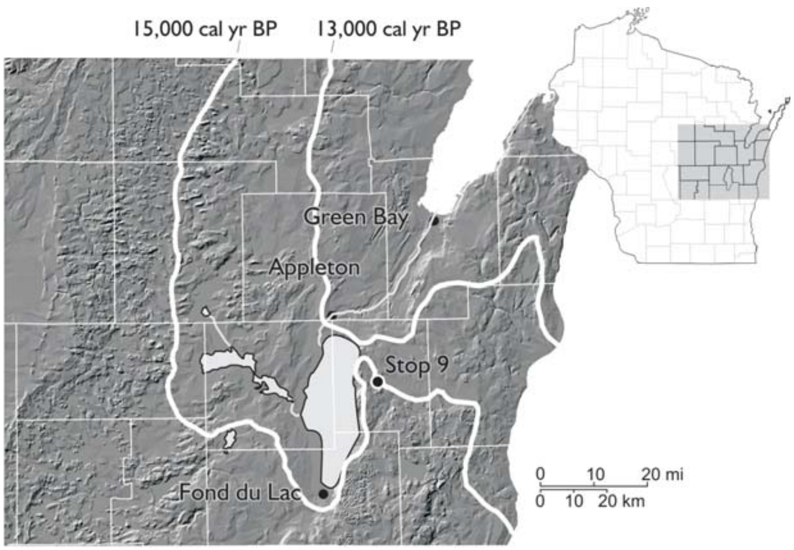


Figure 9.3. Location of ice margin during the two major readvances of the Green Bay Lobe into east-central Wisconsin approximately 16,000 and 13,500 cal yr BP. Location of the Schneider farm is indicated.

Figure 9.4. Photograph of piece of wood with beaver-tooth marks.

Figure 9.5. Cluster of aligned, beaver-chewed pieces of wood in the pond sediment that might be part of a beaver dam.

of beaver-chewed twigs and sand and gravel suggest the remains of beaver lodges. A column of sediment was analyzed for its plant macrofossil content in the University of Wisconsin–Madison Department of Geography. Other common plant macrofossils include spruce and tamarack needles and cones and seeds from the aquatic plants yellow water lily *(Nuphar)* and waternymph *(Najas)*. Invertebrates preserved include water flea *(Daphnia)* egg cases (ephippia), gastropods, and bivalves (fingernail clams, *Pisidium)*. A muskrat *(Ondatra zibethicus)* femur also was found.

Geologic setting and stratigraphy

The Schneider farm is just inside the glacial limit formed by the Green Bay Lobe when it readvanced approximately 16,000 cal yr BP (fig. 9.3) (see Mickelson and others, this volume). During this readvance, the lobe deposited the Chilton Member of the Kewaunee Formation, the equivalent to the Valders and Kirby Lake Members of the Oshkosh and Michigan basins. The till is reddish brown and clay rich and is exposed beneath the pond sediment and peat. Deposition of the pond sediment began when ice receded from

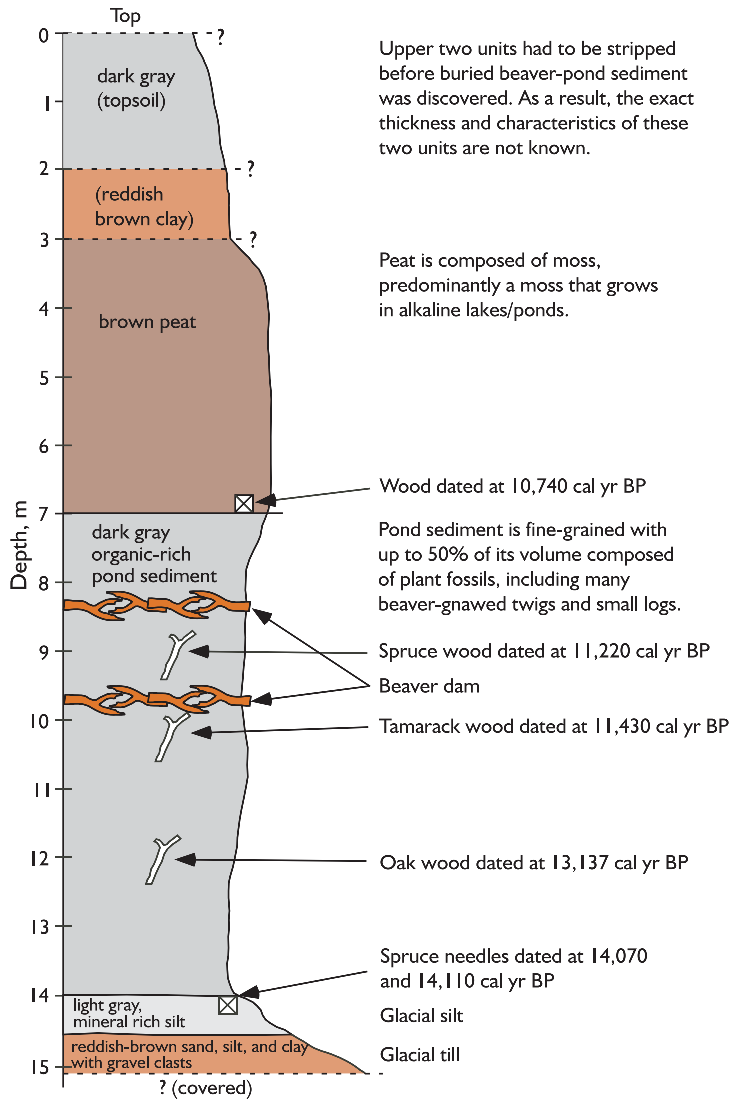


Figure 9.6. Stratigraphic section of sediment exposed in farm-pond excavation at the Schneider farm.

this limit. After several hundred years of recession (the Two Creeks interval), the Green Bay Lobe readvanced approximately 13,500 cal yr BP to within a few miles of the Schneider farm and deposited till of the Glenmore Member of the Kewaunee Formation, which is equivalent to the Middle Inlet and Two Rivers Members. There is no obvious change in the physical character of the pond sediment that might coincide with this readvance.

Resting on the Chilton Member till is basal pond sediment, which is light gray, minerogenic silt. Two radiocarbon dates for spruce needles from this basal pond sediment average 14,090 ±100 cal yr BP (12,060 ±100 ^{14}C yr BP) (table 9.1). Above the thin (several centimeters) silt, the pond sediment quickly becomes dark gray and rich in organic material. This gyttja reaches a thickness of approximately 2 m, and it is this unit that has yielded most of the macrofossils. Its upper contact is sharp and overlain by slightly humified, moss-rich peat. Wood from just above this contact was dated at 10,740 ±100 cal yr BP (9,520 ±100 ^{14}C yr BP). We are uncertain about the total thickness of the peat as well as what might have overlain it prior to excavation. The hypothetical stratigraphy of the upper part of the section (fig. 9.6) is based on the Schneider's description of what they had removed before we arrived at the site.

Tree rings

We studied 40 fossil timbers that were collected as float from the sediment excavated before we were informed about the site. The additional samples that were collected in situ are now under analysis in our lab. We sampled 40 cross sections of the logs and measured tree-ring widths from their polished surfaces using a Hanson measuring system. The tree-ring width series was cross-dated using the Time Series Analysis and Presentation software program (Rinn, 2003). The results of this analysis are presented in Panyushkina and Leavitt (this volume). Spruce specimens greatly outnumbered other tree species (84% of the total number of cross sections). The timbers showed characteristics of normal age growth from an open-canopy stand. We found no evidence for subsequent deposition and preservation of the wood in this wetland

Table 9.1. Radiocarbon dates from the Schneider farm.

Depth[1] ft (m)	Material	Conventional (^{14}C yr BP)	Calibrated[2] (cal yr BP)	Laboratory number	Significance
Float	oak wood	625 ±35 675 ±35	580 ±35 660 ±35	Arizona	age of 82-yr tree-ring series from oak
Float	spruce wood	710 ±70	675 ±70	Arizona	age of 130-yr tree-ring series from spruce
7 (2.1)	wood	9,520 ±100	10,740 ±100	Beta-208987	age of contact between pond sediment (below) and peat (above)
Float	spruce wood	9,790 ±70	11,220 ±70	Arizona	age of 133-yr tree-ring series from spruce
Float	tamarack wood	9,940 ±70	11,430 ±70	Arizona	age of 45-yr tree-ring series from tamarack
14 (4.3)	spruce needles	12,000 ±90	14,070 ±90	Beta-207064	age of basal pond sediment
14 (4.3)	spruce needles	12,120 ±110	14,110 ±110	Beta-207065	age of basal pond sediment

[1] Depths are approximate because of uncertainty about original thickness of deposit.

[2] All radiocarbon dates were calibrated using the methods of Stuiver and Reimer (1998).

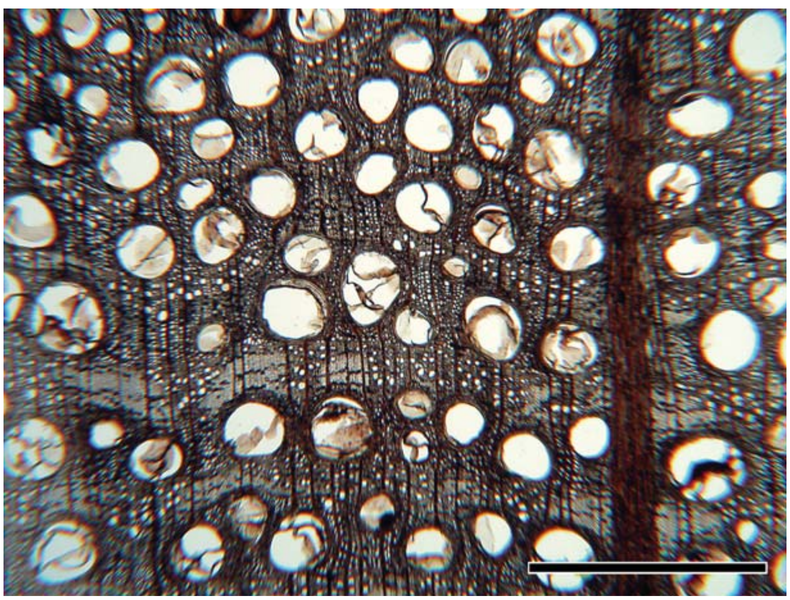


Figure 9.7. Photomicrograph of thin section of well preserved specimen (transverse surface) of sub-fossil wood from the Schneider Farm site. The large, rounded openings are the vessels common in oak *(Quercus),* which is known as a "ring porous" species. The vessel size contains ecophysiological information that can be used for environmental reconstructions. It could be related to temperature or moisture availability during the growth season. (Photograph by Alex Wiedenhoeft.)

environment. Changes in tree species and variation in tree growth were possibly caused by shifting hydrological conditions from a drier to wetter environment.

Tree-ring widths were cross-dated within wood horizons and overlapped into four chronologies: 130 (spruce), 133 (spruce), 45 (tamarack) and 82 (oak) years. The spruce (133 yr) and tamarack (45 yr) chronologies date to between 14,110 ±110 cal yr BP (12,120 ±110 ^{14}C yr BP) to 10,740 ±100 cal yr BP (9,520 ±100 ^{14}C yr BP), and the spruce (130 yr) and oak (82 yr) chronologies date to 675 ±70 to 580 ±35 cal yr BP (710 ±70 to 635 ±35 ^{14}C yr BP) (table 9.1). Each group of cross-dated specimens has distinctive morphological features related to growth conditions and age of trees. Most trees had a very slow growth rate, which suggests unfavorable growth conditions. Only the tamarack seem to have grown in a more favorable environment and showed fast rates of radial growth, sometimes with annual rings containing two bands of late wood cells (so-called false rings) that are commonly associated with a drought in a growing season.

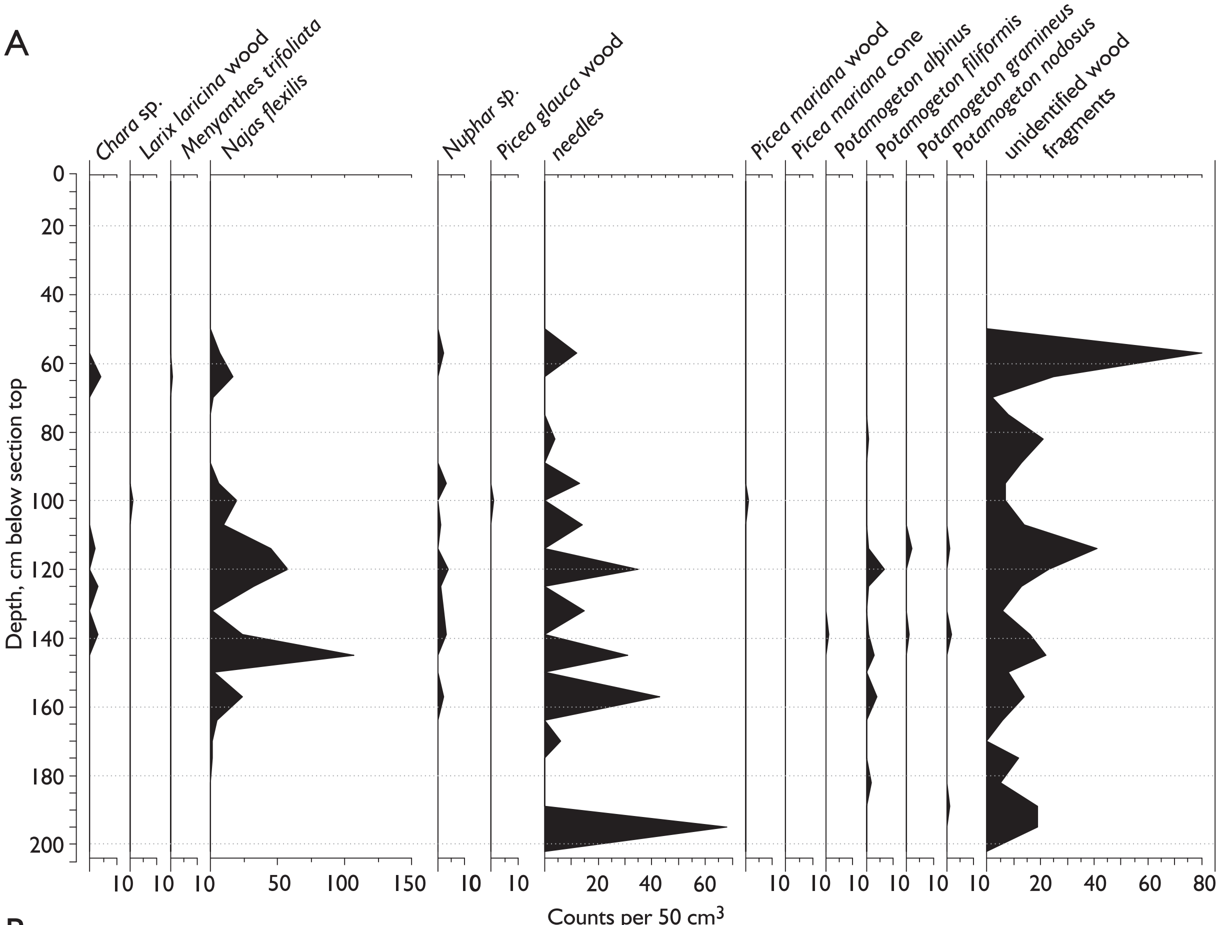
Radiocarbon dating indicates the spruce tree rings fall into two groups that are approximately 10,000 years apart. The average width of tree rings from spruce at the top of the pond sediment was 20. This record was preliminarily placed at about 700 cal year BP. Only these trees had well preserved, complete cross sections with pith, most outer rings, and even fragments of bark. One stand had two spruce generations represented by trees less than 50 years old and up to 120 years old. The stand dynamic (reaction wood, suppressed growth bands) and wood anomalies (resin duct bands and scars) suggest environmental disturbance of spruce growth lasting for approximately 15 years. Growth decline and tree death occurred within the final seven years; during that time the trees were gradually killed by the worsening environmental conditions. The 133-year spruce record from 12 samples dated back to the Younger Dryas; the decayed wood appearance indicated a much older age for these microfossils. The cross sections of these specimens had pith and weathered outer rings, suggesting that the trees died before they were buried. Absence of the outermost rings in most samples does not allow us to determine details on the stand termination. However, the trees showed sudden reduction in their growth for approximately the final four years that are preserved, which could be related to a disturbance.

Our study of fossil wood from the Schneider site showed that the macrofossils deposited during the Younger Dryas–Holocene transition have good potential for dendrochronology. The tree rings are well preserved, cross-datable, and suitable for deriving various tree-ring parameters, including tree-ring widths and radiocarbon measurements from a series of cross-dated rings. It is possible to apply these tree rings for absolute and precise dating of the pond deposits through development of a highly resolved radiocarbon record. The ring-growth characteristics and anomalies can be used for ecological studies of vegetation and pond formation. However, the paleoclimatic potential of these tree rings still must be established (fig. 9.7). A dataset of tree-ring width series is needed to estimate climatic signals in the tree-ring width variability. We are hopeful that the 145 specimens that remain to be analyzed will provide the necessary data.

Macrofossil and charcoal analyses

The stratigraphic section was sampled to identify plant macrofossils (fig. 9.8A) and abundance of charcoal (fig. 9.9). As mentioned previously, the entire section was not characterized because the top meter was removed prior to the initial site visit. The combined thickness of the peat and gyttja at the sampled section is approximately 2.2 m. The thickness of these two units varies across the pond, which was likely caused by an uneven pond depth in addition to dewatering and consolidation of sediments following excavation.

Plant macrofossils include aquatic and upland taxa (fig. 9.8A). Aquatic species include pondweed *(Potamogeton* spp.), pond-lily *(Nuphar* sp.), muskgrass *(Chara* sp.), and waternymph *(Najas flexilis).* Upland taxa include tamarack (*Larix laricina)*, black spruce (*Picea mariana),* and white spruce (*P. glauca).* Buckbean *(Menyanthes trifoliata)* is an obligate wetland species. Most wood fragments were too small for identification. Conifer needles were most abundant at the base of the profile, directly above glacial sediment, and have been interpreted as a trash layer resulting from downwasting of stagnant ice (Wright and Stefanova, 2004). The abundance of needles decreases over time. Larger branches and logs were primarily concentrated at the top of the gyttja and clay unit, immediately below the peat moss. Many of these wood



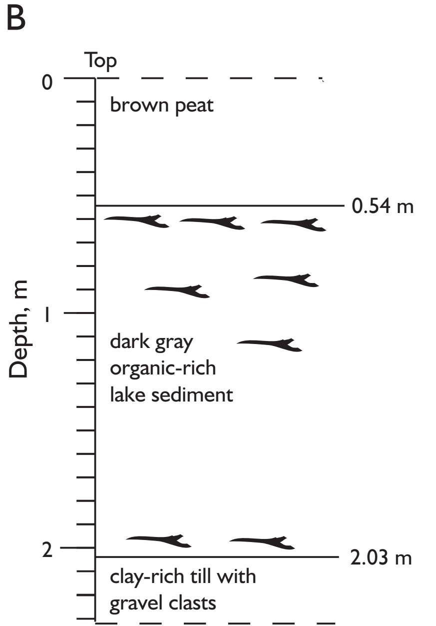
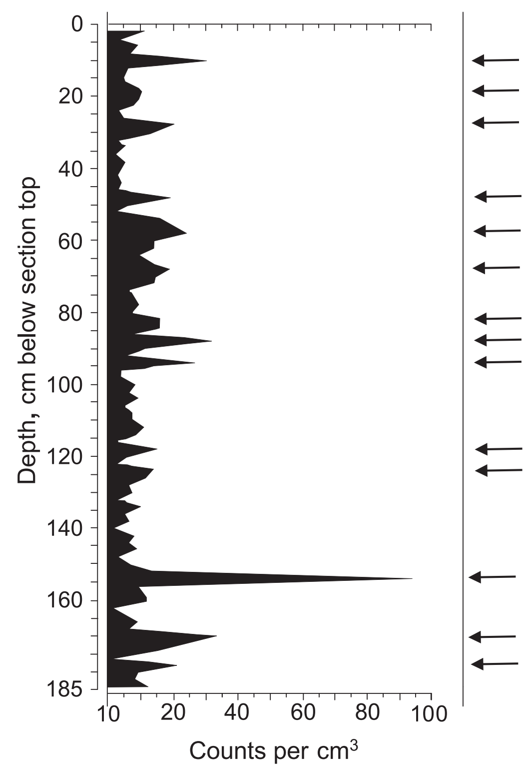


Figure 9.8. Macrofossil diagram **(A)** and stratigraphic section **(B)** for Schneider farm sediment.

Figure 9.9. Charcoal diagram for Schneider farm sediment. Arrows indicate peaks in total charcoal interpreted to represent individual fire events.



fragments displayed beaver-chewed ends.

Preliminary visual analysis of the charcoal record indicated perhaps 14 distinct peaks in charcoal abundances (fig. 9.9), which are interpreted to represent individual fire events. The size of the charcoal fragments studied (>125 μm) suggests that primary transport was wash-in or short-distance atmospheric transport from fires adjacent to the lake or within several kilometers (Whitlock and Larson, 2001). Fire frequency appears to have increased over time.

The decrease in conifer needles and increase in fire frequency are suggestive of a long-term drying trend. The observed hydrological shift from lake to wetland may have been caused by a combination of site-specific factors, including beaver damming and lake infilling, as well as a regional shift towards increased aridity. There is no clear-cut evidence of climatic change associated with the glacial readvance that buried the Two Creeks forest bed.

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Insights into late Pleistocene–early Holocene paleoecology from fossil wood around the Great Lakes region

Irina P. Panyushkina and Steven W. Leavitt

Introduction and background

Early publications about midwestern geology reported the finding of wood in glacial deposits (for example, Goldthwait, 1907; Alden, 1918). Goldthwait (1907) noted preserved wood along the shoreline on the west side of Lake Michigan, including the Two Creeks forest bed, which was overrun by the final advance of the Laurentide Ice Sheet. Wood is found around the Great Lakes in many geological circumstances associated with various modes of preservation, but the quality of the wood is variable from site to site. The deglaciation of the Great Lakes region was marked by rapid colonization of forests on the newly emergent terrestrial landscapes. Preserved remains of such forests provide tree rings for high-resolution climate-related analysis. For example, short-lived glacial readvances inundated some sites with water and sediment, resulting in burial and nearly ideal preservation of forest trees in original growth positions. Such was the case with trunks and branches preserved at the classic Two Creeks site in eastern Wisconsin. Wood as logs or in situ stumps has been preserved in submerged conditions (fig. 1) and also associated with eolian sands, alluvial deposits (fig. 2), lacustrine sediments, and bogs.

We have investigated wood at a number of sites around the Great Lakes (fig. 3; table 1) in an effort to better understand high-resolution environmental variability during the period from about

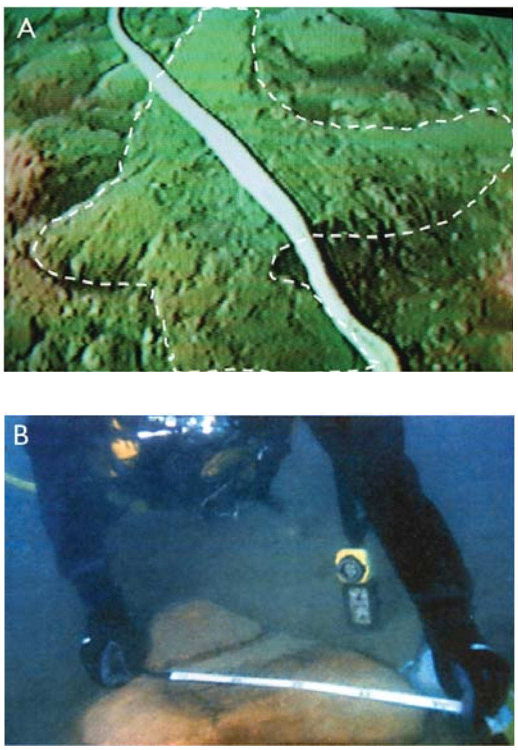


Figure 1. Stumps in submerged forests in Lake Michigan near Jacksonport in Door County, Wisconsin, at depth of approximately 12 m **(A;** dashed lines indicate outline of roots), and in Lake Huron near Sanilac County, Michigan, in approximately 12.5 m of water **(B).** (Photograph A by F. Pranschke; photograph B by L. Clyburn.)

Figure 2. Spruce stump with roots in Gribben Basin site, Michigan; the sand and gravel outwash was associated with a brief advance of the Laurentide Ice Sheet in Lake Superior. (Photograph by P.E. Martin.)

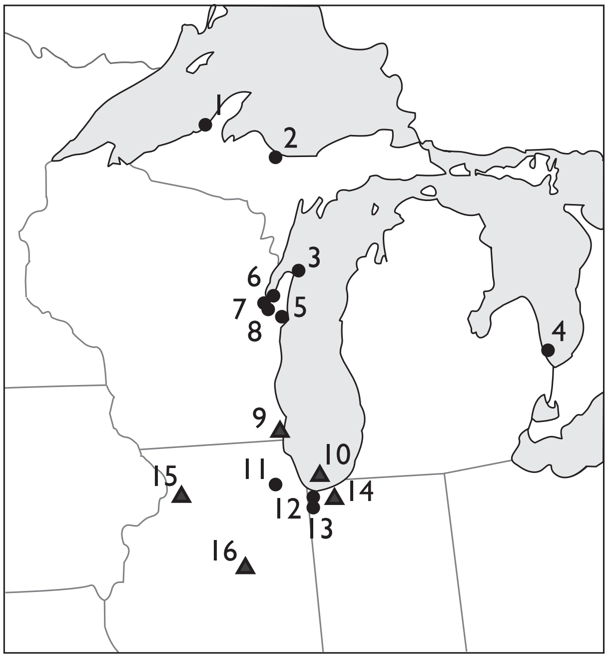


Figure 3. Location of recent and ongoing investigations involving wood between approximately 14,000 and 4,000 cal yr BP. Circles indicate conifers and triangles indicate hardwoods (angiosperms). 1=Elm River, 2=Gribben Basin, 3=Jacksonport, 4=Sanilac, 5=Two Creeks, 6= Green Bay–GB, 7=Green Bay–AH, 8=New Denmark, 9=Southport, 10=Olson, 11=Brewster, 12=Gary Sand Pit, 13=Liverpool East, 14=Brown's Sand Pit, 15=Markman, 16=Lincoln Quarry.

Table 1. Selected sites with ancient wood, in order of increasing age. Number next to site name refers to site in figure 3. *Carya*=hickory; *Gleditsia*=honeylocust; *Fraxinus*=ash; *Larix*=larch; *Morus*=mulberry; *Picea*=spruce; *Quercus*=oak; *Thuja*=white cedar; *Ulmus*=elm.

Site	Species	General radiocarbon age (^{14}C yr BP)	Approximate calibrated age (cal yr BP)	Information and dating source
12 Gary Sand Pit	*Pinus*	3,100–4,000	3,300–4,500	Leavitt and others (2006)
1 Elm River	*Larix*	5,500–5,800	6,200–6,700	unpublished dates (S. Shetron; Panyushkina and Leavitt)
3 Jacksonport	*Thuja/Picea*	6,500	7,300–7,500	Pranschke and Shabica (1993); Leavitt and others (2006)
4 Sanilac	*Thuja*	6,400–7,100	7,300–8,000	Hunter and others (2006); Leavitt and others (2006)
9 Southport	*Quercus/Carya/Fraxinus*	4,800–7,600	5,500–8,400	Sander (1969); Schneider and others (1977); Leavitt (1989)
10 Olson	*Quercus/Fraxinus*	8,100–8,400	9,000–9,500	Chrzastowski and others (1991); Pranschke and Shabica (1993)
15 Markman	*Ulmus*	9,000	10,000–10,200	Kim (1982); unpublished dates (Panyushkina and Leavitt)
11 Brewster	*Picea/Larix*	9,200–10,900	10,300–12,900	Curry and others (2006)
2 Gribben Basin	*Picea*	9,000–10.300	10,900–12,300	Lowell and others (1999); Pregitzer and others (2000); Leavitt and others (2006)
14 Brown's Sand Pit	*Picea*	10,400–11,800	12,100–13,800	Cole (1987)
13 Liverpool East	*Picea*	9,920–10,420	11,200–12,600	Schneider and Hansel(1990); Morgan and others (1991); Leavitt and others (2006); Panyushkina and others (2005)
5 Two Creeks	*Picea*	11,600–12,000	13,300–13,900	Goldthwait (1907); Wilson (1932, 1936); Thwaites and Bertrand (1957); Broecker and Farrand (1963); Black (1970); Leavitt and Kalin (1992); Kaiser (1994)
8 New Denmark	*Picea*	11,600	13,300–13,500	Moran and others (1988); Leavitt and others (2006)
7 Green Bay–AH	*Picea*	11,100–11,300	12,900–13,200	Leavitt and others (2006)
6 Green Bay–GB	*Picea*	11,900	13,700–13,800	Schweger (1969); Thwaites (1958); Leavitt and others (2006)
16 Lincoln Quarry	*Fraxinus/Morus/Gleditsia/Quercus*	8,500–14,100	9,500–17,000	Panyushkina and others (2004)

Figure 4. Black spruce *(Picea mariana)* cones found by I.P. Panyushkina at the Liverpool East site. (Photograph by S.W. Leavitt.)

14,000 to 4,000 years ago. Through the generous contribution of wood and site information from others, together with new collections at previously identified sites and recently discovered sites, we have amassed a collection of almost 600 wood specimens from more than 20 sites.

Generally, plant macrofossils such as cones and wood can be important resources for interpreting past environments. If identifiable from the macrofossils, tree species can sometimes be diagnostic of climate and environmental conditions (Thompson and others, 1999). For example, the black spruce (*Picea mariana*) cones (fig. 4) discovered at the Liverpool East site (ca. 10,000 ^{14}C yr BP) in northern Indiana are from a species associated with moist lowland conditions (Panyushkina and others, 2005). The white cedar (*Thuja*) and hemlock (*Tsuga*) dominated wood assemblage with coexisting pine (*Pinus*), spruce (*Picea*), and ash (*Fraxinus*) found in the submerged Sanilac site in Lake Huron (approximately 7,000 ^{14}C yr BP) indicates a community classification of "rich conifer swamp," also known as cedar swamp (Hunter and others, 2006).

The tree rings in wood samples can be particularly important in regard to what their ring size and pattern of growth indicate. For example, individual small rings in a sequence can denote a harsh growing environment that might have resulted from cold conditions at the end of the Pleistocene, and the sequence of changes in ring size during a tree's lifetime can indicate evidence of climate or competition effects. At the Two Creeks site, early observations of ring-size changes suggested suppression of growth in the final decades (for example, Wilson, 1932; Kaiser, 1994), which would be consistent with rising water levels as the readvance of the Lake Michigan Lobe of the Laurentide Ice Sheet blocked northern outlets of meltwater discharge.

We have examined the tree rings of a number of sites to establish tree-ring width "chronologies." We based these chronologies on cross-dating (correlating) the patterns of ring width among samples to establish the age relationship of rings. The chronologies of the late Pleistocene–early Holocene sites in the Great Lakes area are considered "floating" because they cannot be connected to modern tree-ring chronologies and are therefore not absolutely dated. In Europe, however, an absolute chronology from oak tree rings has been built back 10,000 years from modern living trees, historical wood from buildings and structures, and wood preserved in geologic deposits (Friedrich and others, 2004). The European chronology has been extended back to the late Younger Dryas (approximately 11,600 years ago) by matching with a floating pine chronology (Becker and Kromer, 1986). The Midwest floating chronologies we derived in the age range of approximately 10,000 to 14,000 years ago are depicted in figure 5. Although floating, it is possible to extract useful environmental information from the rings by means of the ring-width size and variability mentioned above, "event" chronologies based on micro-anatomical features (frost rings, reaction wood, traumatic resin ducts, etc.), isotopic composition, and spectral analysis of ring-width and isotope series.

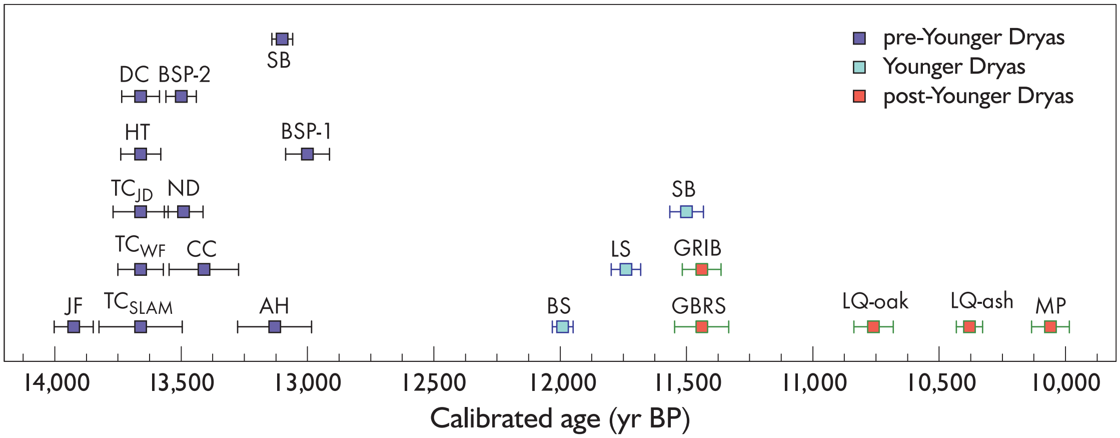


Figure 5. Distribution of the tree-ring chronologies over a 4,000-year interval of the Pleistocene–Holocene transition from Great Lakes sites (Panyushkina and Leavitt, 2006). Tree-ring width chronologies (horizontal bars) are plotted at midpoint of the respective calibrated radiocarbon age. The tree-ring width chronology length is represented by the length of the horizontal bars. The letters designate sites and collections; for example, "TC" represents Two Creeks type locality collections of different researchers (WF=W. Ferguson; JD=J. Dean; SLAM=S. Leavitt and A. McCord), "JF"=Jorgensen Farm, "CC"=Calumet County, "AH"=Green Bay–Amerihost, "ND"=New Denmark, "HT"=heating tunnel (UW–Green Bay), "DC"=Door County, "SB"=Schneider Farm, "BSP"=Brown's Sand Pit, "BS"=Brewster Creek, "LS"=Liverpool East, "GRIB" and "GBRS"=Gribben Basin, "LQ"=Lincoln Quarry, "MP"=Markman Peat. The Liverpool East site may actually plot closer to 12,000 cal yr BP on the basis of radiocarbon "wiggle-matching" of ring sequences to the master radiocarbon calibration curve (Leavitt and others, in press).

Example: Sites of Two Creeks age

Four of the sites in figure 3 (and table 1) are considered to be approximately Two Creeks in age, and represent a small number of discoveries of Two Creeks equivalent wood reported around Wisconsin (four of which appear in figure 6, designated "Two Creeks," "Green Bay–AH," "Green Bay–GB," and "New Denmark"). Some of the reported finds of Two Creeks age material in the region are just single pieces of wood, even "twigs," rather than the assemblages of log samples that would be much more useful for dendrochronological work. However, we found that a lone log discovered in Calumet County, Wisconsin, had 278 rings, the most reported for any sample of Two Creeks age.

The length of the chronology from the Two Creeks type locality is 329 years, based on contributions from 52 trees, and aided greatly by the discovery of a single tree with 233 rings (fig. 7), just one ring short of the log reported by Kaiser (1994) and used in his initial chronology of 252 years developed from only wood in contact with the buried soil. The wood from these sites appears to be spruce, which Kaiser (1994) considered to be black spruce. However, there may well have been white spruce (*P. glauca*) on upland sites and black spruce on lowland sites—Black (1970) suggested that white spruce in the area was generally more abundant here than black spruce.

Nobel Prize recipient Willard Libby (1955) was the first to date wood of the Two Creeks type locality to 10,700 to 12,200 ^{14}C yr BP. Broecker and Farrand (1963), Black and Rubin (1967–68), Suess (1979), Leavitt and Kalin (1992), and Kaiser (1994) have subsequently dated several pieces of wood from the site. The fairly consistent results of Broecker and Farrand (1963), Leavitt and Kalin (1992), and Kaiser (1994) have substantially refined the age estimate

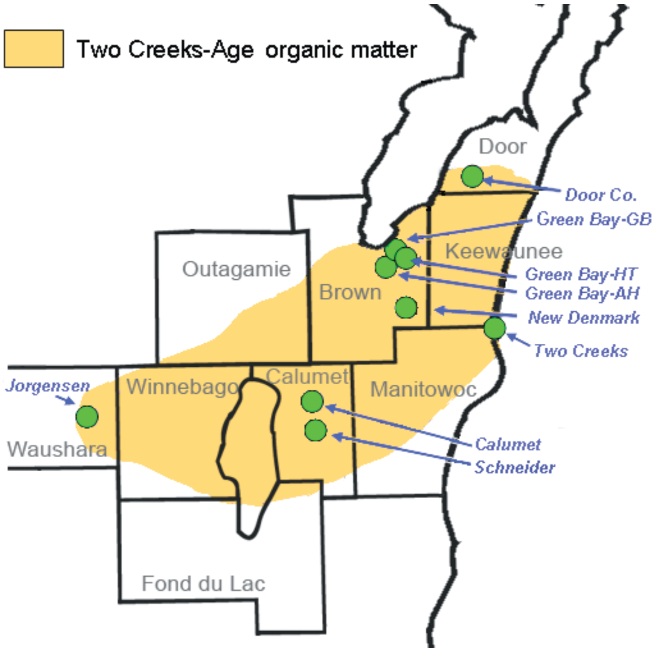


Figure 6. Locations where Two Creeks age-equivalent wood has been found (modified from Black, 1970).

to an average of approximately 11,800 to 11,850 ^{14}C yr BP. The youngest dates of Kaiser (1994) on outside rings tend to fall close to 11,600 ^{14}C yr BP, concordant with what might be expected if the forest bed represented a lifespan of approximately 300 calendar years. Leavitt and others (in press) estimated an outer age in the tree-ring series of approximately 11,600 ^{14}C yr BP, which is equivalent to approximately 13,530 cal yr BP, on the basis of radiocarbon "wiggle"-matching.

Wood being eroded out of the type locality bluff along Lake Michigan has been observed as stumps rooted in the forest bed (fig. 8) and as logs, primarily in the lacustrine sediments immediately above the forest bed, but also in the red till above the lacustrine sediments (Goldthwait, 1907; Wilson, 1932; Black, 1970), with the logs in the till having been transported the farthest. Wilson (1932) suggested that the trees in the lacustrine

Figure 7. Large log (UA-TCIP-12; possibly part of tipped stump) containing 233 rings discovered at the Two Creeks type locality in 2003. (Photograph by I.P. Panyushkina.)

Figure 8. Stump (PKS-TCSL-102) tilting southwest at Two Creeks type locality with roots in the soil horizon and trunk extending up through gray lacustrine sediments. The upper end of trunk is truncated as if sheared off by the advance of the ice sheet past this position. (Photograph by S.W. Leavitt.)

sediments were growing closest to the site and showed a decade or two of growth suppression of the tree rings in response to rising water levels; those in the till did not show the growth suppression, perhaps because they were from upland localities. In most cases, the stratigraphic position of the wood in our collections is unknown, usually because the wood has been eroded out of the bluff. Thus, the 329-year provisional tree-ring chronology undoubtedly contains wood that is likely local as well as wood that has been transported some significant (but unknown) distance north. However, the end of the chronology should represent those trees that had survived the longest (probably local trees) before their ultimate demise from rising water and eventual overriding by the Lake Michigan Lobe.

The Two Creeks ring-width series was standardized into a ring-index series, which removes non-climatic effects, such as long-term growth trends, and produces a mean series index of 1.0 (Fritts, 1976). The tree-ring index variability represents changes of environmental conditions for the period of 329 years at this locality (fig. 9). Only a few positive periods (>1.0) were favorable to tree-ring growth. Pronounced and abrupt negative growth episodes (<1.0) are more frequent in this record. On the basis of the relationship between modern spruce tree rings and mountain glacier mass balance from the Coastal Mountains in British Columbia (Larocque and Smith, 2005), the Two Creeks tree-ring record might be related to dynamics of the Lake Michigan Lobe. The positive tree-ring growth periods could be associated with warmer summer conditions and degradation of the ice sheet. The negative periods of tree rings more likely represent advances of the ice sheet.

We also aligned tree-ring width records of spruce from eight sites of suspected Two Creeks age to the end of growth at the Two Creeks type locality and correlated pointer intervals (distinct sequences of low or high tree-ring values identified visually). It seems that the unique signatures of tree-ring width variability correspond well among these sites (fig. 10), suggesting that the trees responded to the same environmental impacts across the region. Furthermore, wood from some locations showed high spruce growth (mean 0.66–0.46 mm for New Denmark and Green Bay-GB and AH) and higher variance of radial growth (0.27–0.25) in relation to some of the other sites (mean 0.36–0.40 mm and standard deviation 0.22–0.14 for Two Creeks, Green Bay–HT, Door County, and Calumet). There are seven strong and synchronized events of negative impact on tree growth more than 150 years before the Two Creeks trees were buried: The trees

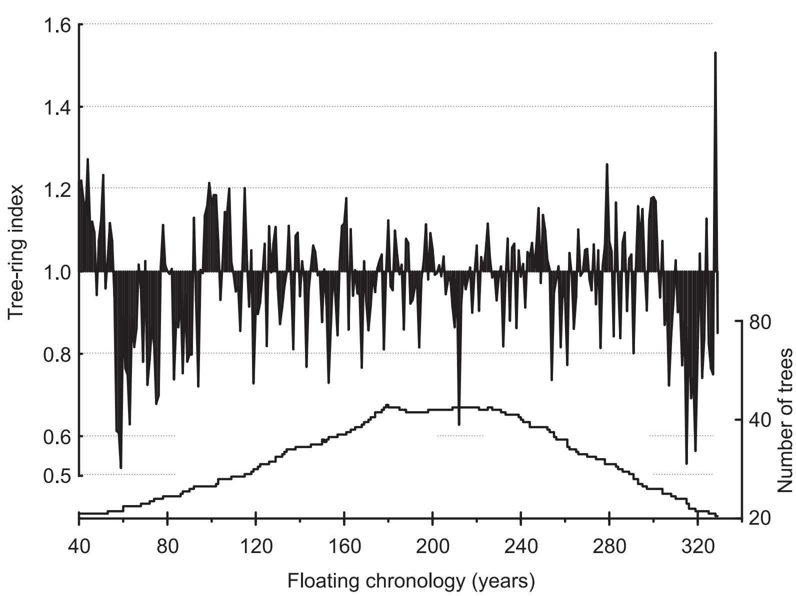


Figure 9. Tree-ring width index chronology of the Two Creek site (upper curve) and number of trees contributing to chronology (lower curve).

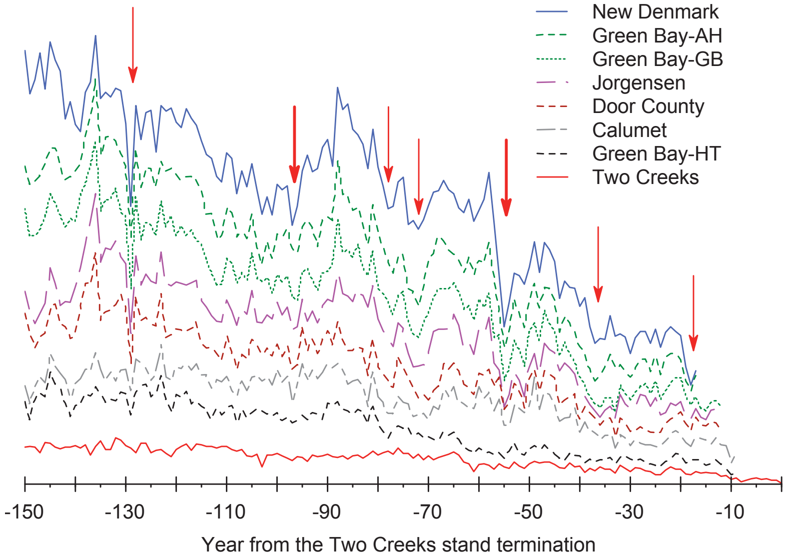


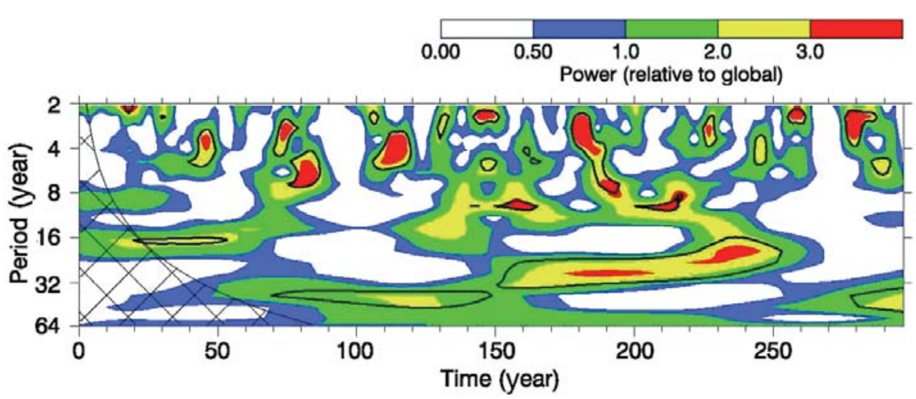
Figure 10. "Stacked plots" of averaged tree-ring width chronologies for locations of Two Creeks age material. Tree-ring width means increase from bottom to top. Caution: The true means and variability are not equally scaled among the sites, so they cannot be compared. The value of these plots is to highlight common response of trees to events and absolute differences in stand termination dates. Red arrows indicate pointer years.

responded strongly to negative impact at 129, 97, 78–77, 72, 55, 37–36 and 19 years before the dieback. These one- and two-year events could be an impact of meltwater and outwash, and there is evidence of traumatic resin ducts in some of the trees following these negative events.

Wilson (1932, 1936) suggested trees found in the red till at the type locality had larger rings than those in or near the forest bed horizon, and therefore represent trees being transported from sites with more favorable growth conditions. We found that average tree-ring widths at the type locality (including logs from all horizons) were actually narrower than those of all the other sites. Furthermore, Wilson (1932, 1936) suggested that some individual small rings of the wood from nearest the forest bed corresponded to wide rings in the wood from the red till above, suggesting that excessive moisture reduced growth close to the shoreline, but enhanced growth in the trees on higher ground from which the wood in the red till was presumably transported. However, our results (fig. 10) suggested occurrences of narrow (and wide) rings are consistent across a large area and over a range of 80 m or more of elevation, so a more universal environmental influence would be required to explain the narrow rings. Perhaps katabatic winds might be a better explanation than meltwater and outwash for this widespread impact.

Termination of all stands occurred within a 17-year period, according to the tree-ring records (fig. 10). It seems the spruce survived longer at the Two Creeks locality in comparison with other sites. However, we do not know how well the breadth of our collection captures true stand ages and termination dates.

Figure 11. The wavelet power spectra for tree rings from the Two Creeks type locality with the first 30 years of record cut off (Panyushkina and Leavitt, 2006). The Morlet wavelet function in Fourier space was applied, and the power has been normalized by the global wavelet spectrum that measures the deviation from the mean spectrum. The cross-hatched region is the cone of influence, where zero padding has reduced the variance. Black contour is the 90 percent significance level, using a red-noise (autoregressive lag 1) background spectrum.



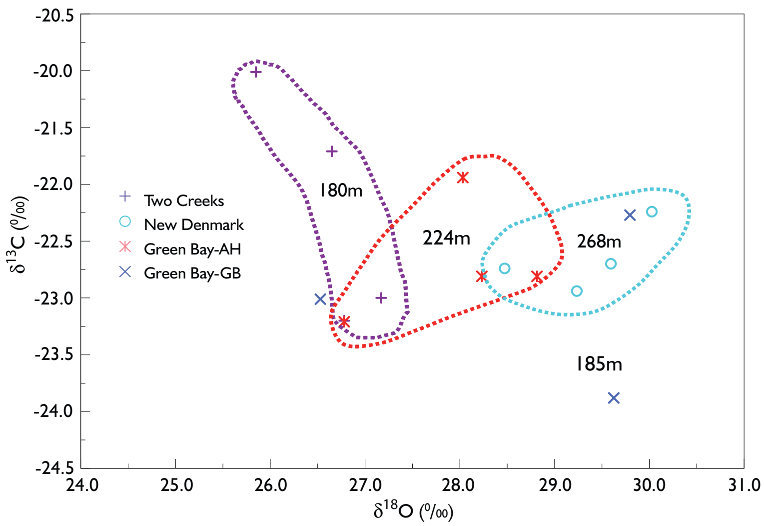


Figure 12. Clusters of points for the Two Creeks, Green Bay–AH, and New Denmark sites in $\delta^{13}C$–$\delta^{18}O$ space. Site elevations above sea level are indicated. The Green Bay–GB site seems to spread across and outside the fields of the other site values. (Modified from Leavitt and others, 2006.)

We analyzed the spectral characteristics of this series (fig. 11) to determine how they compared to modern large-scale climate drivers, such as the ENSO phenomenon that exhibits periodicity of 2.8 to 7.6 years. We detected this periodicity in the index series, at least intermittently, and also observed another strong period of 20 to 30 years. Interestingly, the Two Creeks record showed a 100-year period between years 150 and 250 with high power of the 20 to 30 year periodicity that may be forced by solar magnetic activity (centered at 25–30 years in our case), which weakened toward the end of the record. It seems to correspond to other lines of evidence on the climatic nature of variability seen in the tree-ring records of the Two Creeks age.

The bulk stable isotope composition ($\delta^{13}C$, $\delta^{18}O$, and $\delta^{2}H$) of wood from four of the Two Creeks age sites was also analyzed for several samples per site (Leavitt and others, 2006). A plot of $\delta^{13}C$ and $\delta^{18}O$ indicates that three sites can be distinguished (Two Creeks, New Denmark, and Green Bay–AH) with respect to the fields in which their points lie (fig. 12). Although the $\delta^{13}C$ values seem to largely overlap, the three sites seem to cluster according to modern elevation differences. However, the pattern is the opposite of what would be expected if orographic lifting were decreasing the $\delta^{18}O$ value of water vapor and precipitation eventually incorporated into the tree rings. More likely, temperature is contributing to this phenomenon, related to greater daytime temperatures farther inland (higher elevations) and away from lake cooling during the growing season (Leavitt and others, 2006). The Green Bay–GB site does not follow this pattern, but it was collected from sediments that show evidence of deformation (Schweger, 1969), and therefore the provenience is less certain. This illustrates the great value of having wood from multiple sites of the same age to examine spatial as well as temporal aspects of regional environmental variability.

Conclusions

Wood has been found in geological deposits at a number of sites in the Midwest, and although it is extremely desirable for radiocarbon dating, it has been largely underutilized with respect to environmental reconstructions. In part, this may be a consequence of the uneven quality of wood preservation dependent on conditions of burial, subsurface processes during burial (including hydrology, microbial activity, oxidation potential), and post-depositional effects. Ring-width measurements and chronology development, tabulation of event chronologies, stable isotope analysis, and specialized methods of data and time-series analysis offer great promise for extracting useful and perhaps diagnostic environmental information from this resource.

We have systematically applied these methods to wood from some long-known deposits as well as others that were very recently discovered. Fortunate geological circumstances of wood preservation and chance discovery have resulted in a wide age range of chronologies, but the typical chronology length of approximately 100 to 200 years means that it will be difficult to develop a continuous absolute chronology dating back 10,000+ years, similar to that being developed in Europe. However, good professional contacts will be helpful in continuing acquisition of suitable samples and filling in some of the gaps, and radiocarbon wiggle-matching may resolve the absolute age within a decade or two in the absence of a continuous absolute chronology. The acquisition of contemporaneous wood from widely separated sites offers the opportunity to make further inferences about regional characteristics of high-resolution climate variability.

Acknowledgments

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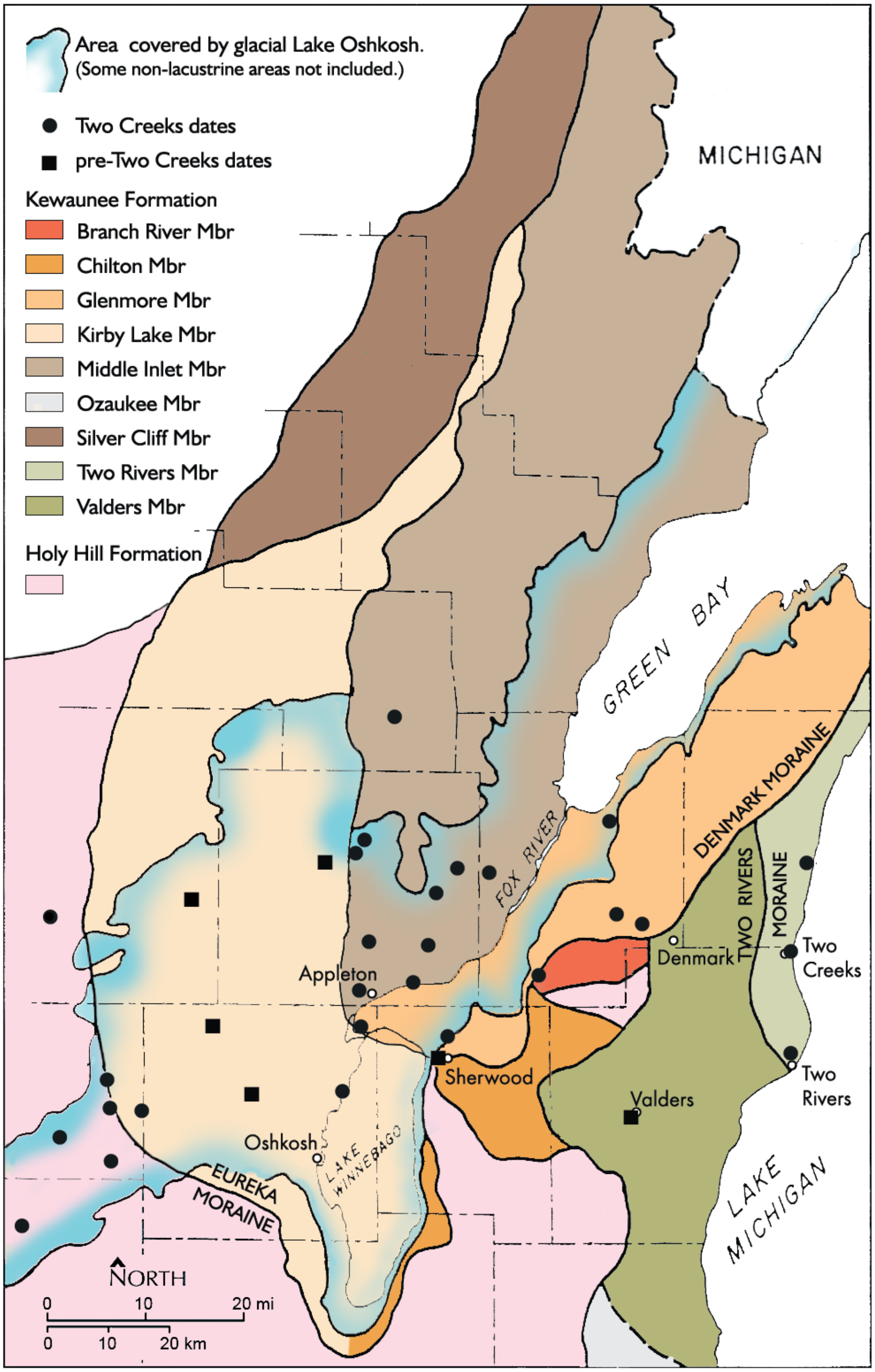

Figure 1. Overview map of stratigraphic units in east-central Wisconsin. Squares indicate locations of samples with pre-Two Creeks radiocarbon dates and circles indicate locations of samples with Two Creeks radiocarbon dates.

Late-glacial ice advances and vegetation changes in east-central Wisconsin

D.M. Mickelson, Thomas S. Hooyer, Betty J. Socha, and Cornelia Winguth

Introduction

The Two Creeks forest bed is arguably the most famous Quaternary deposit in the Great Lakes area. Since its description 100 years ago by Goldthwait (1907), it has been found to be widespread in east-central Wisconsin and at one site in Michigan. It has been the object of many studies; as approaches to research have changed, new studies have been undertaken. In the first use of his method on non-archeological samples, Libby (1952) radiocarbon-dated Two Creeks wood almost 60 years ago. Until that time, people had little idea of when glaciers receded from North America, with the exception of an estimate based on varve correlation. More recently, studies of stable isotopes in Two Creeks tree rings have produced an improved knowledge of late-glacial climate differences over short distances (Leavitt and others, 2006). Our data have allowed us to better understand late-glacial climate in the Great Lakes region and its relationship to climate change elsewhere.

The only organic matter found beneath till in the northern Great Lakes area that is older than the Two Creeks forest bed and younger than the late-glacial maximum has also been discovered in east-central Wisconsin (fig. 1). These organic horizons, found below the Valders and Chilton Members of the Kewaunee Formation (table 1), provide additional insight into paleoenvironmental conditions, deglaciation history, and the history of glacial Lake Oshkosh. They indicate tundra conditions distinctly different from those indicated by the much more widespread and younger Two Creeks forest bed. This change in conditions appears to have taken place quite rapidly at approximately 14,000 cal yr BP (12,400 ^{14}C yr BP)[1], during the Bölling–Alleröd interstadial.

Several recent accelerator mass spectrometry (AMS) dates of this older organic matter lie in two time intervals: 18,069 ±150 to 17,125 ±142 cal yr BP (14,910 ±80 to 13,710 ±50 ^{14}C yr BP) and 16,335 ±630 to 15,846 ±659 cal yr BP (13,380 ±270 and 12,965 ±200 ^{14}C yr BP) and suggest tundra conditions and permafrost (table 1). This is supported by ice-wedge casts, floral remains, and pollen. All are under till of the Kewaunee Formation. No radiocarbon dates between 15,846 ±659 and 13,893 ±152 cal yr BP (12,965 ±200 and 12,110 ±70 ^{14}C yr BP) are known from beneath Kewaunee till, possibly suggesting ice cover between those dates. The ice then receded at least to the northern end of Lake Michigan; this recession corresponds to the time of the Two Creeks forest bed (many dates from approximately 14,000 to 13,000 cal yr BP [12,400 to 11,200 ^{14}C yr BP]). By then, temperature had risen, spruce forest dominated, glacial Lake Oshkosh had drained, and the level of Lake Michigan had fallen below its present level.

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1 *Radiocarbon dates used in this paper are expressed as calendar dates before present (cal yr BP), plus or minus one standard deviation derived from CalPal calibration. This is followed by the radiocarbon-year equivalent and lab number. For general reference to timing of events, calibrated dates are used with radiocarbon-year dates in parentheses.*

Table 1. Stratigraphic units in the area covered by the Green Bay and Lake Michigan Lobes and the relative stratigraphic position of organic deposits.

Green Bay Lobe West side	Green Bay Lobe East side	Lake Michigan Lobe
	Kewaunee Formation	
Middle Inlet Member	Glenmore Member	Two Rivers Member
	Two Creeks forest bed *(many wood dates averaging 13,500 cal yr BP [11,850 ^{14}C yr BP])*	
Kirby Lake Member	Chilton Member	Valders Member
	Organic horizon *(five dates between 15,846 and 16,335 cal yr BP [12,965 and 13,380 ^{14}C yr BP]; six dates between 17,125 and 18,069 cal yr BP [13,710 and 14,900 ^{14}C yr BP])*	
	older units	

The Two Creeks dates lag behind the dates of peak warming (approximately 14,500 cal yr BP [12,350 ^{14}C yr BP]) in the Greenland Ice Sheet (GRIP) core, but closely match the recession of ice calculated by an ice-flow model (Winguth and others, 2004). The lag may be due to the time it took for ice to recede and for spruce forest to establish, or it may be a true lag in climate adjustment. Presumably in response to Younger Dryas cooling, the Two Rivers advance blocked the northern end of Lake Michigan; as Lake Michigan rose to the Calumet level, approximately 12 m above present lake level, the Two Creeks forest at the type locality was drowned and overrun by ice about 13,000 cal yr BP (11,200 ^{14}C yr BP). In the Oshkosh basin, ice advanced southward to approximately the north end of Lake Winnebago. Water in glacial Lake Oshkosh again rose, flooding spruce trees in much of the basin. Till of the Middle Inlet–Glenmore–Two Rivers advance was then deposited.

History of stratigraphic interpretations of reddish-brown clayey till in east-central Wisconsin

In Ontario, the late-glacial ice advance that deposited two reddish-brown silty till units has long been recognized and called the Port Huron readvance (Dreimanis, 1977). In Michigan, on the east side of Lake Michigan, the Port Huron event was identified as forming two moraines separated by outwash, but composed of similar till (Eschman and Mickelson, 1986). Two Port Huron till units were also identified under Lake Michigan (Lineback and others, 1972). On the basis of radiocarbon dates, the Cheboygan bryophyte bed, which underlies this till, was thought to be older than the Two Creeks forest bed (Farrand and others, 1969). However, the Cheboygan site was later re-dated by Larson and others (1994) and found to be Two Creeks in age. Thus, in Ontario and Michigan and at the bottom of Lake Michigan, a two-phased Port Huron ice advance was recognized by the early 1970s (although the reinterpretation of the Cheboygan bryophyte bed as Two Creeks age may change that interpretation).

In contrast, all the reddish-brown diamicton in east-central Wisconsin, most of which is interpreted to be till, was called the Valders till and, at least locally, the event has been named the Valderan Substage (Willman and Frye, 1970). This advance was thought to postdate the Two Creeks forest bed and to represent a single, major late Wisconsin glacial event in the Great Lakes area (Thwaites and Bertrand, 1957; Black, 1970, 1974).

The Valders till was originally named for the reddish-brown clayey till at the Valders Lime and Stone Quarry, a site at in the city of Valders in central Manitowoc County that has long been visited by geologists (fig. 1; Thwaites and Bertrand, 1957). There, Valders till lies on striated bedrock or older sandy till of the Holy Hill and Hayton Formations. Although it was assumed that all the reddish-brown, clayey till was younger than the Two Creeks forest bed, no site is known outside the Denmark and Two Rivers moraines in Manitowoc and Calumet Counties (fig. 1), where Two Creeks wood is overlain by till. Behind the two moraines, wood is fairly abundant. The post-Two Creeks advance has been recognized in pollen diagrams outside the area of the Kewaunee Formation deposits by an increase in spruce *(Picea)* (West, 1961; Schweger, 1966; Grimm and Maher, 2002), but these sites are not under till (fig. 1).

By the early 1970s, there was a conflict in interpretations that appears to have been largely ignored in the literature at that time. Two Port Huron (immediately pre-Two Creeks) ice advances that deposited reddish-brown, clayey till were recognized under Lake Michigan, in Michigan, and Ontario. In Wisconsin, Black (1970, 1974) argued that there was only one "red till" and that it was post-Two Creeks in age. Evenson (1973) reasoned that the reddish-brown, clayey till in east-central Wisconsin and western Michigan was actually deposited by several glacial advances, some of which predate the Two Creeks forest bed and at least one of which postdates the forest bed. Evenson's arguments were based mainly on geomorphology. In particular, he noted that the Glenwood shoreline of Lake Michigan, which is older than the Two Creeks forest bed (Eschman and Farrand, 1970; Hansel and Mickelson, 1988), was cut into the Valders till surface south of Two Rivers (at Valders Quarry), but not into reddish-brown clayey till north of Two Rivers (fig. 1). Thus, the Valders till, whose type section at Valders Quarry must also be older than the Two Creeks forest bed (Evenson, 1973; Mickelson and Evenson, 1975). Black (1980, 1983), in a series of papers and rebuttals, disagreed with this interpretation, preferring instead the views of earlier workers, who had argued that the till at Valders Quarry was younger than the Two Creeks forest bed at the Two Creeks type locality.

Acomb and others (1982) and McCartney and Mickelson (1982) studied the stratigraphy of the reddish-brown, clayey till units in the Lake Michigan and Green Bay Lobes. They argued that stratigraphic interpretation supported Evenson's ideas about the relative age of the Valders till and the Two Creeks forest bed. That is the present interpretation (table 1), which is similar to that incorporated into the lithostratigraphic classification of Quaternary units in Wisconsin (Mickelson and others, 1984). Although the presumed Valders equivalent in the Green Bay Lobe (Chilton Member) is found beneath the forest bed in several places, the Valders Member appears to be missing at the Two Creeks type locality, where the forest bed overlies lake sediment and the older Ozaukee Member of the Kewaunee Formation. To our knowledge, no site has been documented in which till of the Valders Member can actually be shown to underlie the forest bed, but there is abundant circumstantial evidence that this is the case. Part of the reason for this uncertainty is the nearly identical composition of the units (Acomb and others, 1982).

We now believe that the limits of the Glenmore and Two Rivers Members at the Denmark and Two Rivers moraines (fig. 1), respectively, mark the maximum readvance of post-Two Creeks ice as described by Evenson (1973) near Two Rivers and later by Evenson and Mickelson (1974), Mickelson and Evenson (1975), McCartney and Mickelson (1982) and Acomb and others (1982).

Pre-Two Creeks organic deposits

Sherwood Quarry material

Plant remains found at a site in Calumet County have provided an indication of the environment before the ice of the Green Bay Lobe advanced over the area and deposited reddish-brown clayey Chilton till. However, the till below the organic sediment is not from the advance immediately before the Chilton advance, but the one that deposited the older Cato Falls Member of the Hayton Formation. Locally, concentrations of plant remains, including leaves, stems, mosses, and twigs, are present in gray fine sand and silt underlying the Chilton till. The plant material is radiocarbon dated at 16,434 $\pm$513 cal yr BP (13,370 $\pm$90 ^{14}C yr BP) (Beta–119360). Underlying the gray fine sand and silt is approximately 0.5 m of grayish-brown silty-sand till. (Details of the site geology are given in the stop 8 description, this volume.)

The plants growing on the gray fine sand and silt were overridden by the Chilton ice advance. The radiocarbon date for the plants is an indication of when the ice reached its farthest extent onto the Silurian escarpment. The plant remains found at Sherwood indicate a cold, artic environment—they are arctic taxa, with abundant *Dryas integrifolia* leaves, and many seeds of *Silene acaulis,* much moss, and no evidence of trees (Richard Baker, written communication, 1998). Barrens and cliffs are the general habitat for *D. integrifolia* and *S. acaulis* (Maher and others, 1998). Mosses and *D. integrifolia* are abundant at this site in Sherwood and are common in the Cheybogan bryophyte bed, but are rare at the Valders site (Maher and others, 1998).

Valders Quarry material

Pollen, plant remains, mollusks, and ostracodes were present below Valders till in a glaciolacustrine unit in Valders Quarry (fig. 1), but this site has been mined, and this sediment can no longer be seen. Radiocarbon dates for organic matter collected from the glaciolacustrine unit (fig. 2) ranged from 17,540 ±185 to 15,846 ±659 cal yr BP (14,210 ±90 to 12,965 ±200 ^{14}C yr BP) (Maher and others, 1998). The organic assemblage indicates a cold, tundra-like, open-ground environment (Maher and Mickelson, 1996; Maher and others, 1998). Ostracodes found at the site indicate an open lake environment with changing water depth and proximity to the ice front. Plant macrofossils of more than 20 species that today are seen in tundra and open-forest tundra in northern Canada were found in the glaciolacustrine unit, also indicating a cold, open environment (Maher and others, 1998).

Figure 2. Photograph of organic pond sediment at Valders Lime and Stone Quarry, Valders, Wisconsin.

Summary of age relationships

A cold, apparently tundra environment is indicated by the Valders and Sherwood organic horizons. From the distribution of the relatively few dates, it appears that the sites might represent two ice-free times that were probably separated by an ice advance (table 1), although ice would have been continuously present not far north. It is possible that these advances correlate to the two Port Huron tills (Shorewood and Manitowoc) found under Lake Michigan (Lineback and others, 1972) and to the two Port Huron advances in Michigan and Ontario. By approximately 13,800 cal yr BP (12,000 ^{14}C yr BP), permafrost was gone and spruce trees covered at least part of the landscape, as evidenced by the Two Creeks forest bed.

Two Creeks forest bed

History and distribution

The forest bed was first described by Goldthwait (1907) in sec. 24, T21N, R24E, approximately 3.2 km south of the village of Two Creeks. As far as we know, it has not been exposed there in the past 30 years, except when the Two Creeks Nuclear Power Plant was built, but it is well exposed along the lake shore bluff in sec. 2, T21N, R24E, just south of the Manitowoc–Kewaunee County line (intersection of Highways BB and 42, between Highway 42 and Lake Michigan). This exposure is considered the type section and is now part of the Ice Age National Scientific Reserve. At the type section, the base of the bluff is made up of what we now call Ozaukee Member, which extends about 2 m above beach level near the north end of the section. This till unit drops below lake level to the south. It is overlain by clayey silt and fine sand lake sediment that coarsens upward to the sand and gravel that we have interpreted to be shallow-water and beach sediment. The forest bed lies on this sand and gravel and consists of a few centimeters of organic material that is mostly spruce *(Picea mariana)* needles, twigs, branches, and some rooted stumps (fig. 2). Limited amounts of white spruce, balsam fir, tamarack, and aspen have also been identified (Mode, 1989). This is overlain by 2 to 3 m of silty lake sediment, which in turn is overlain by Two Rivers till. A thin (less than 1 m) sand and gravel unit overlies the till. Throughout the bluff, the layers were deformed by the overriding ice. The till contains numerous pieces of wood picked up as the glacier moved across the dead forest.

Two Creeks wood is also present in many places along, and a few miles behind, the Two Rivers and Denmark moraines in and west of Denmark, and in the Fox River valley between Green Bay and Appleton (fig. 1). A bryophyte bed in the northern part of the southern peninsula of Michigan near Cheboygan, once thought to be older than Two Creeks, is now thought to be the same age (Larson and others, 1994).

Libby (1952) used the solid carbon method to initially date Two Creeks wood. Many samples have been dated subsequently (for example, Broecker and Farrand, 1963; Black and Rubin, 1967–68; Kaiser, 1994; Leavitt and others, 2006), and the radiocarbon dates center around 13,500 cal yr BP (11,850 ^{14}C yr BP). The Two Creeks ages are nearly all between approximately 14,000 and 13,000 cal yr BP (12,400 and 11,200 ^{14}C yr BP).

Overview of vegetation and other studies

When the ice that deposited the Valders till receded from the location of the type section, Lake Michigan was at the Glenwood level, approximately 18 m above the present level of Lake Michigan. Silt and clay were deposited in the deep water until the glacier receded far enough north to allow lake level to drop. When the water level dropped to below the present level, a spruce forest began to grow at the site and over much of the landscape that is now east-central Wisconsin. On the basis of revegetation histories in southeast Alaska (Lawrence, 1958), it may have been 50 or 100 years before spruce trees became dominant. The soil that developed during this time was described by Lee and Horn (1972) as a thin organic horizon over a gleyed B horizon.

From tree-ring studies at the type section, Kaiser (1994) estimated that the forest grew for at least 252 years and noted that the oldest tree reported had 234 annual growth rings. An almost 1,000-year record of tree rings (from about 14,000 to 13,000 cal yr BP (12,400 to 11,200 ^{14}C yr BP) has been assembled by Leavitt and others (2006) and Panyushkina and Leavitt (this volume). At the type section and at many sites in the Fox River lowland, the trees were killed by rising lake level (Wilson, 1932, 1936). At least two sites between Green Bay and Denmark were exposed during construction of Interstate 43, where there was no lake sediment above the forest bed, and the trees must have been pushed over directly by the advancing ice. In many of the logs, tree rings thin to the outer bark, indicating that the trees were stressed by some combination of rising water and cold winds from the ice sheet. Where not completely isolated from the force of the overriding ice by lake sediment, tree trunks are commonly oriented toward the southwest, indicating the ice-flow direction (Black, 1970). The fossil pollen in the forest bed is relatively poorly preserved, and the population is not diverse (Wiese, 1979). In addition to spruce wood and needles, the forest bed contains aquatic and air-breathing snails, but they do not aid in constraining climatic conditions at the time very well. Morgan and Morgan (1979) and Garry and others (1990) interpreted beetle remains from the site as indicating a more boreal climate than that of today, much like that of the modern boreal forest north of Lake Superior. Note the contrast in the fossil record (and therefore climate) between this site and the somewhat older sites at Valders and Sherwood.

Figure 3. Photograph of rooted stump from Two Creeks forest bed type locality.

Approximately 13,800 cal yr BP (12,000 ^{14}C yr BP), glacial ice advanced into the north end of the Michigan basin, damming the northern outlet and forcing lake level to rise. At the type section, the forest bed is covered with lake sediment deposited as lake level rose to the Calumet level. Figure 3 shows a rooted stump surrounded by this lake sediment. Driftwood is abundant in the sediment as well as in the Two Rivers till above. Lake Michigan Lobe ice then overrode the site, advancing as far as present-day Two Rivers (Schneider and Hansel, 1990), and the Green Bay Lobe advanced southward to the Denmark moraine and into the Fox River basin to the present location of Appleton.

Two new Two Creeks forest bed sites were found during recent mapping in Calumet County (fig. 1). One site is located north of Hilbert in the SW¼NE¼ sec. 19, T20N, R19E. Two boreholes were drilled in the broad, low end moraine that marks the farthest advance of the Glenmore ice into the Brillion basin. The horizon containing the wood is overlain by Glenmore till. The wood was radiocarbon dated at 13,456 ±140 cal yr BP (11,690 ±70 ^{14}C yr BP) (Beta–10481). Spruce wood, interlayered with red clayey till containing dispersed organics and a gleyed (green gray 5GY 5/1) soil horizon, was also encountered in an adjacent borehole. The wood in this horizon appears to be tree trunks or large branches. The wood appears fresh, with yellow color and distinct small growth rings. The wood is radiocarbon dated at 13,893 ±152 cal yr BP (12,110 ±70 ^{14}C yr BP) (Beta–11558).

Another Two Creeks site was found north of Brillion, in southern Brown County, in SE¼SE¼ sec. 16, T21N, R20E. A borehole was drilled in the sharp-crested, narrow Denmark moraine (fig. 1), which marks the farthest advance of the Glenmore ice onto the Silurian escarpment. Material from the boring was dated at 13,101±135 cal yr BP (11,210 ±100 ^{14}C yr BP) (Beta–177408) Underlying the paleosol horizon containing wood is an organic mat, approximately 6 cm thick, consisting predominantly of mosses. The moss horizon is radiocarbon dated at 13,736 ±159 cal yr BP (11,820 ±100 ^{14}C yr BP) (Beta–177409).

Six new Two Creeks sites have recently been discovered in the Oshkosh basin south of the Denmark moraine. At all six locations, wood was discovered buried beneath lake sediment and was radiocarbon dated between 13,850 ±148 and 12,891±53 cal yr BP (12,070 ±60 and 11,020 ±40 ^{14}C yr BP). The ages indicate that the forest bed in the basin most likely drowned due to a rising lake level. At two sites the land elevation is high enough that the lake must have drained southward out of the Dekorra outlet. This indicates that the ice margin must have covered all four eastern outlets to the Michigan basin, including the Manitowoc outlet. Field mapping in Calumet County confirmed this interpretation (Mickelson and Socha, in press). It appears, however, that the Manitowoc outlet was high enough that it was not reoccupied as a glacial Lake Oshkosh outlet during recession of this ice.

Chronology and paleoglaciology of late-glacial ice advance and retreat

Figure 4 shows our interpretation of ice-margin positions and lake phases during the advances immediately before and after Two Creeks. Rapid ice-margin oscillations (advance and recession rates of approximately 0.5 km per year) are indicated by the ice-margin positions between 16,400 and 13,000 cal yr BP (13,350 and 11,200 ^{14}C yr BP). Wet-bed conditions (rapid sliding and bed deformation) may have contributed to these rapid oscillations or surge advances may have been unstable (Colgan, 1999). This period of margin oscillations coincides with the period of rapid climate change indicated in the Greenland ice cores.

The pre-Two Creeks (Kirby Lake–Chilton–Valders Members) and the post-Two Creeks (Middle Inlet–Glenmore–Two Rivers) advances had distinctly different areal patterns that suggest different ice dynamics (figs. 1, 4, and 5). The pre-Two Creeks advance extended about 50 km farther south and southeast than did the post-Two Creeks advance. Yet farther north, the pre-Two Creeks advance was thinner than the post-Two Creeks advance. Near Denmark (fig. 1), for instance, the Chilton margin was somewhere north of the Glenmore margin. The Glenmore advance extended farther and to higher elevations near Denmark, and a similar relationship is obvious on the west side of the Green Bay Lobe (fig. 1). This indicates that the pre-Two Creeks (Kirby Lake–Chilton–Valders) advance had a gentler ice-surface slope than the post-Two Creeks (Middle Inlet–Glenmore–Two Rivers) advance. Socha and others (1999)

Figure 4. Maps showing extent of glacial ice, glacial Lake Oshkosh, and its outlets and the shore of Lake Michigan at various times during deglaciation. Dates are approximations.

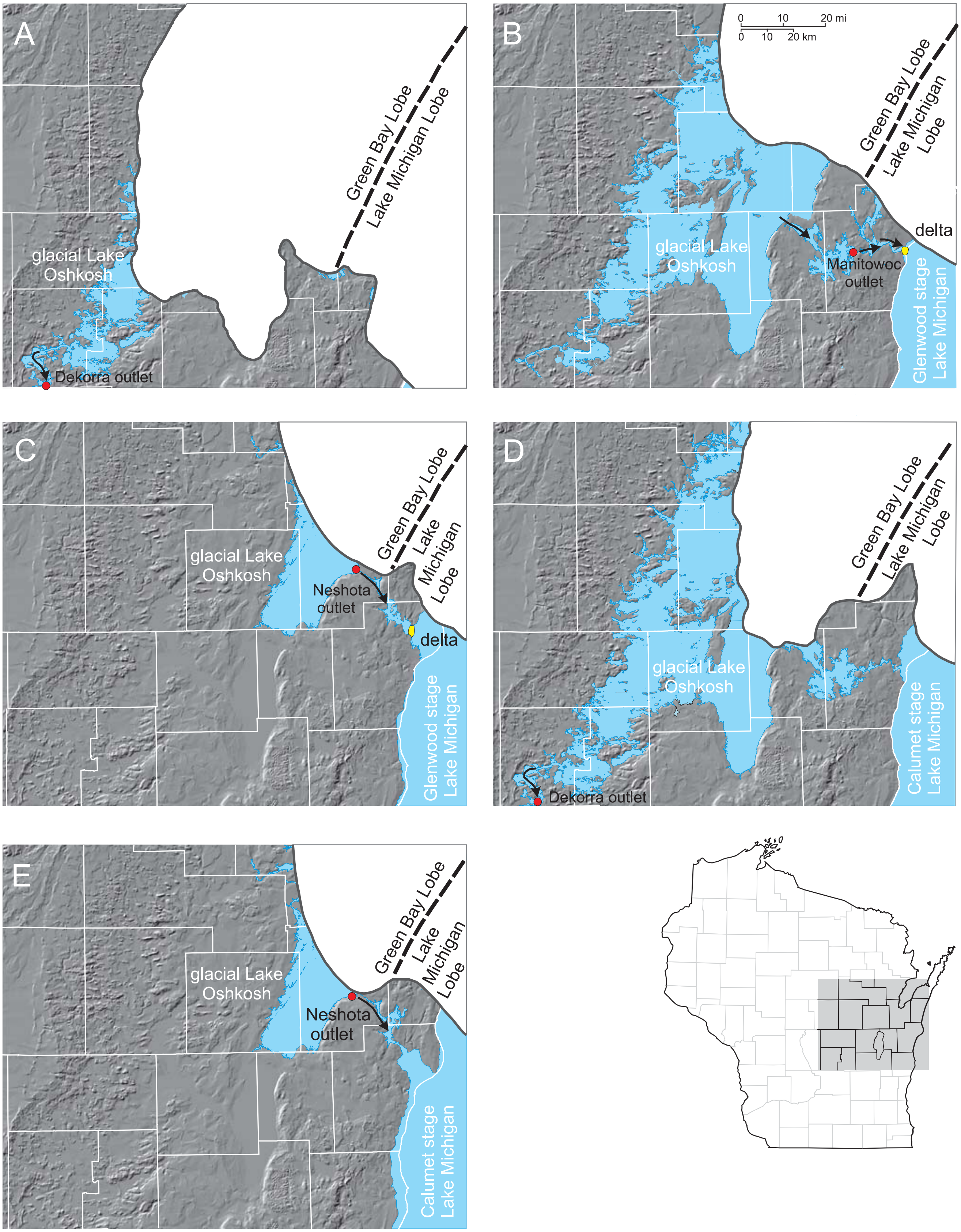
A. At approximately 15,900 cal yr BP (13,000 ^{14}C yr BP), ice reached its maximum readvance position and deposited the Kirby Lake, Chilton, and Valders tills. The west east edge of the Green Bay Lobe formed an interlobate deposit with the Lake Michigan Lobe of the Valders advance. Glacial Lake Oshkosh used the Dekorra outlet to the Wisconsin River. Lake Michigan is at the Glenwood stage.

B. At approximately 15,400 cal yr BP (12,800 ^{14}C yr BP), ice began to recede. Glacial Lake Oshkosh dropped to the Manitowoc outlet and drained through the Manitowoc River to Lake Michigan, where it built a delta at the Glenwood level.

C. At approximately 15,200 cal yr BP (12,600 ^{14}C yr BP), ice continued to recede. Glacial Lake Oshkosh dropped to the Neshota outlet and built a delta (shown in yellow) near Mishicot at the Glenwood level. Ice then receded out of the Michigan basin and lake level dropped well below present level. Except for Lake Winnebago, which likely had a rock threshold, the Oshkosh basin was dry land as was the area that is now Green Bay. The Two Creeks forest began to grow.

D. At approximately 13,600 cal yr BP (11,850 ^{14}C yr BP), ice advanced into the Oshkosh basin depositing the Middle Inlet, Glenmore, and Two Rivers tills. Glacial Lake Oshkosh reformed and rose to the Dekorra outlet, and perhaps at the same time or after a short distance of retreat, the Neshota outlet. The Two Creeks forest was flooded, then covered by ice.

E. At approximately 13,200 cal yr BP (11,400 ^{14}C yr BP), ice receded to open the Neshota outlet. Delta was built into Lake Michigan at the Calumet level. The Manitowoc outlet was not used during this advance and recession cycle.


A
Green Bay Lobe
Lake Michigan Lobe
glacial Lake Oshkosh
Dekorra outlet
B
0 10 20 mi
0 10 20 km
Green Bay Lobe
Lake Michigan Lobe
delta
glacial Lake Oshkosh
Manitowoc outlet
Glenwood stage Lake Michigan
C
Green Bay Lobe
Lake Michigan Lobe
glacial Lake Oshkosh
Neshota outlet
delta
Glenwood stage Lake Michigan
D
Green Bay Lobe
Lake Michigan Lobe
glacial Lake Oshkosh
Calumet stage Lake Michigan
Dekorra outlet
E
Green Bay Lobe
Lake Michigan Lobe
glacial Lake Oshkosh
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Calumet stage Lake Michigan

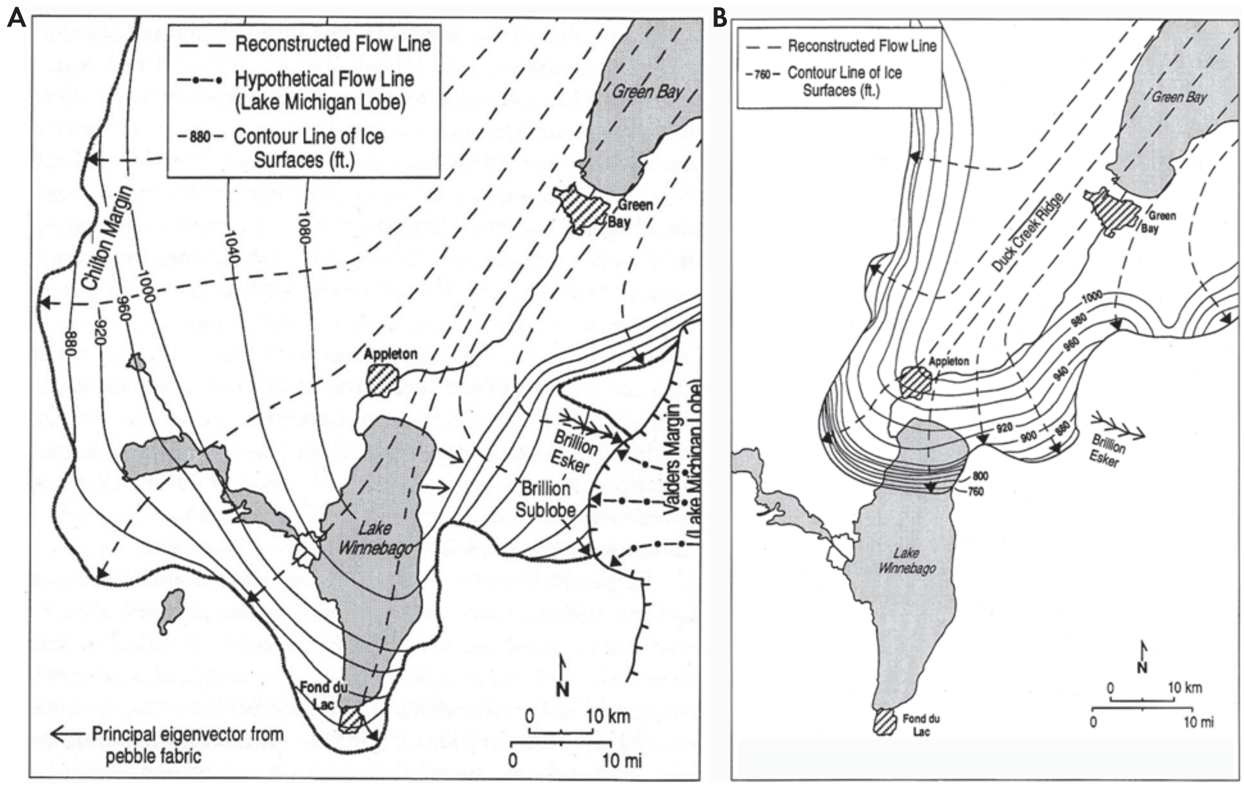

Figure 5. Reconstructed ice surfaces at the maximum extent of the advance that deposited the Kirby Lake, Chilton, and Valders tills **(A)** and Middle Inlet, Glenmore, and Two Rivers tills **(B).** (From Socha and others, 1999.)

reconstructed the ice-surface morphology for these two advances (fig. 5) to estimate ice thickness and basal driving stresses and to evaluate the role of low driving stress in the dynamic behavior of the ice margin. Their reconstruction agrees with the relative ice-surface slopes suggested by the mapped extent of the tills—that is, that the advance that deposited the pre-Two Creeks (Kirby Lake–Chilton–Valders) advance had a gentler ice-surface slope than the post-Two Creeks (Middle Inlet–Glenmore–Two Rivers) advance. They also concluded that surfaces of both ice advances had gentler slopes and lower driving stresses than the Green Bay Lobe advance that deposited Holy Hill Formation (Colgan, 1999).

Comparison with ice-core records

Radiocarbon dates from east-central Wisconsin were calibrated using two popular calibration programs, CALIB (Reimer and others, 2004) and CALPAL (2005), but we used only CALPAL dates to construct figures 5 and 6. Figure 5 was constructed by software called ISOPLOT (Ludwig, 2004), a program commonly used by geochronologists. It shows the probability of the CALPAL calibrated dates reported in this paper being the true age. Times of low probability of being a true age (in other words, a time least likely for the organic material to be growing) may be times of ice cover. Times of high probability (the time the organic material was most likely growing) are interpreted as times when the area was ice free. The Two Creeks dates show a high probability of ice-free time between 14,000 and 13,000 cal yr BP (12,400 and 11,200 ^{14}C yr BP). There are few older dates, but the data are suggestive of an ice-free time about

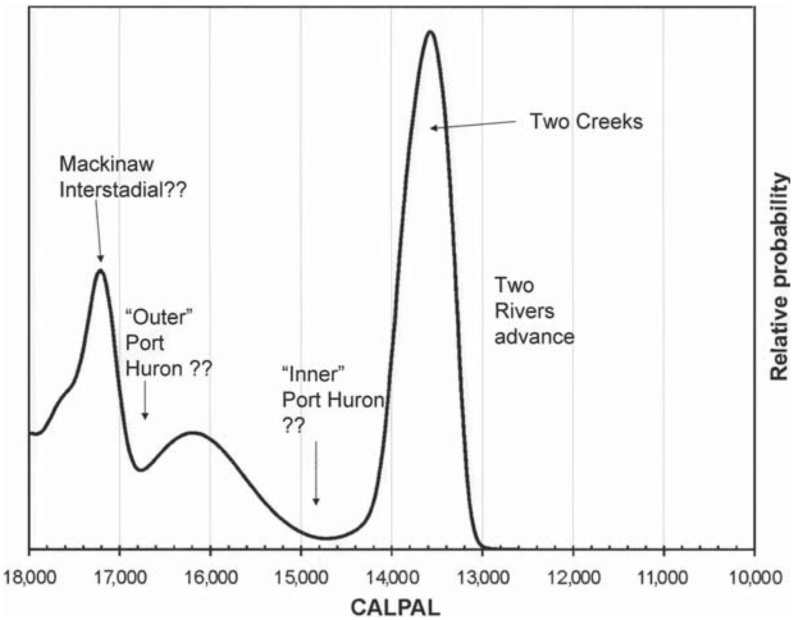

Figure 6. Plot from ISOPLOT program of the probability distribution of radiocarbon dates from east-central Wisconsin. (Not all Two Creeks dates were used because using the large number of Two Creeks dates and the small number of pre-Two Creeks dates would make the diagram impossibly tall or the older dates inconsequential.)

17,200 cal yr BP (13,800 ^{14}C yr BP) and again about 16,300 cal yr BP (13,300 ^{14}C yr BP). It is possible that the two times of more probable ice cover (least likely age for organic material) at about 16,800 and 15,000 cal yr BP (13,600 and 12,600 ^{14}C yr BP) correlate with the inner and outer Port Huron advances recognized in Michigan and the late-glacial till stratigraphy of the Green Bay and Lake Michigan Lobes. Because of the small number of dates and the lack of complete stratigraphic context, this is only speculative.

We have, however, enough dates for the Two Creeks forest bed to compare our interpretation of climate with the climate conditions at the same time in ice-core records from Greenland. Figure 7A shows a plot of the GRIP $\delta^{18}O$ record. The colored vertical bars indicate the probability distributions of ice-free times shown in figure 5. Note that not all of the Two Creeks dates were used, so the actual placement of the vertical bar may be slightly biased by the selection of dates, but all dates suggest an ice-free time from about 14,000 (12,400 ^{14}C yr BP) to about 13,000 cal yr BP (11,200 ^{14}C yr BP). This period covers part of the Bölling–Alleröd warm time, but lags the peak warmth by about 500 years. Note in figure 7A that Younger Dryas cooling in the GRIP core started almost exactly when the Two Creeks forest bed was finally buried (about 12,800 cal yr BP). Winguth and others (2004) tuned the GRIP record with a Missouri speleothem record (Dorale and others, 1998) as input for an ice-flow model. The mean annual temperature at the location of Madison generated by the model is shown in figure 7B. This curve shows a warm time that almost exactly coincides with the ice-free Two Creeks time in Wisconsin. Winguth and others (2004) also modeled the movement of the margin of the Green Bay Lobe, and figure 6C shows rapid recession during that time.

Conclusions

The time between 18,000 and 13,000 cal yr BP (14,900 and 11,200 ^{14}C yr BP) was a time of rapid environmental change in east-central Wisconsin. On the basis of only a few radiocarbon dates, it appears that ice advances at approximately 17,200 and 16,300 cal yr BP (13,800 and 13,300 ^{14}C yr BP) may represent two pre-Two Creeks ice advances that are recorded in the stratigraphic record of the Green Bay Lobe, the western Lake Michigan Lobe, the bottom of Lake Michigan, and the Inner and Outer Port Huron moraines of Michigan and southern Ontario. Until more organic deposits from this time are found and dated, this idea is speculative. Ice-free times between glacial events had cold conditions with tundra vegetation. The re-

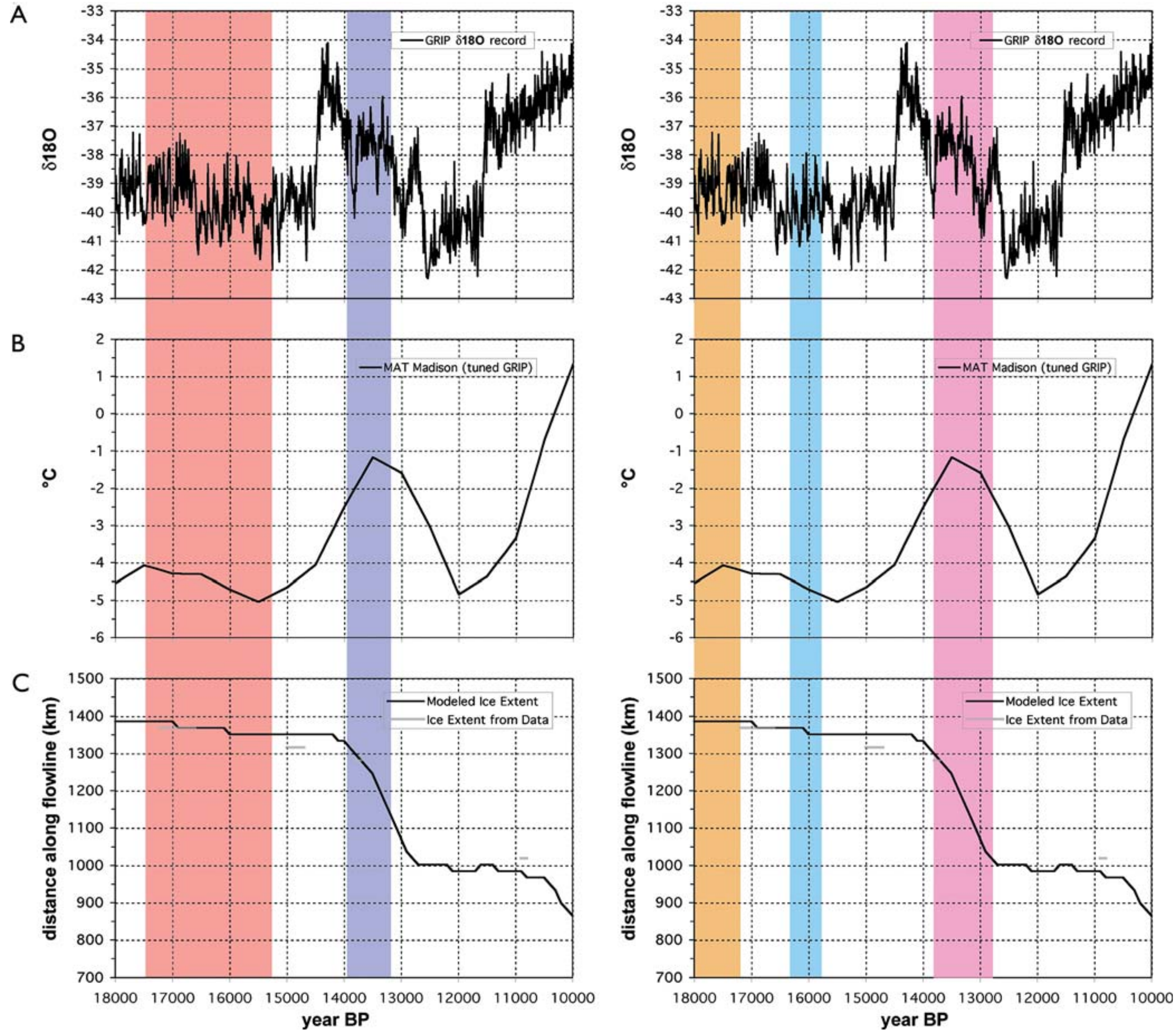

Figure 7. Plot of calendar years on the horizontal axis and **(A)** the GRIP $\delta^{18}O$ record, **(B)** the modeled temperature at Madison, Wisconsin, using a "tuned" GRIP record (from Winguth and others, 2004), and **(C)** modeled ice extent of the Green Bay Lobe (from Winguth and others, 2004). Vertical bands through the figure are times of likely ice-free times in east-central Wisconsin based on ISOPLOT analysis. All dates calibrated with CalPal.

cord in Wisconsin does not relate in a convincing way to the ice-core record (fig. 6A). This may be because we have so few dates in that age range.

The Two Creeks forest bed appears to correlate with the Bölling–Alleröd warm time, although the CalPal calibration dated the bed approximately 500 years after the peak warmth. This might be expected if ice had to recede before the forest could grow. The Two Creeks dates seem to correspond better with the temperature curve generated by Winguth and others (2004). The post-Two Creeks ice advance, which deposited the Middle Inlet–Glenmore–Two Rivers Members, appears to be a reaction to the cooling that led to the Younger Dryas cooling seen in the ice core.

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Meetings of the Midwest Friends of the Pleistocene

1	1950	Eastern Wisconsin	
2	1951	Southeastern Minnesota	H.E. Wright, Jr. and R.V. Ruhe
3	1952	Western Illinois and eastern Iowa	P.R. Shaffer and W.H. Scholtes
(4)	1953	Northeastern Wisconsin	F.T. Thwaites
(5)	1954	Central Minnesota	H.E. Wright, Jr., and A.F. Schneider
6	1955	Southwestern Iowa	R.V. Ruhe
(7)	1956	Northwestern lower Michigan	J.H. Zumberge and W.N. Melhorn
8	1957	South-central Indiana	W.D. Thornbury and W.J. Wayne
9	1958	Eastern North Dakota	W.M. Laird and others
10	1959	Western Wisconsin	R.F. Black
11	1960	Eastern South Dakota	A.G. Agnew and others
12	1961	Eastern Alberta	C.P. Gravenor and others
13	1962	Eastern Ohio	R.P. Goldthwait
14	1963	Western Illinois	J.C. Frye and H.B. Willman
15	1964	Eastern Minnesota	H.E. Wright, Jr. and E.J. Cushing
16	1965	Northeastern Iowa	R.V. Ruhe and others
17	1966	Eastern Nebraska	E.C. Reed and others
18	1967	South-central North Dakota	Lee Clayton and T.F. Freers
19	1969	Cyprus Hills, Saskatchewan and Alberta	W.O. Kupsch
20	1971	Kansas and Missouri Border	C.K. Bayne and others
21	1972	East-central Illinois	W.H. Johnson, L.R. Follmer, and others
22	1973	West-central Michigan and east-central Wisconsin	E.B. Evenson and others
23	1975	Western Missouri	W.H. Allen and others
24	1976	Meade County, Kansas	C.K. Bayne and others
25	1978	Southwestern Indiana	R.V. Ruhe and C.G. Olson
26	1979	Central Illinois	L.R. Follmer, E.D. McKay, and others
27	1980	Yarmouth, Iowa	G.R. Hallberg and others
28	1981	Northeastern lower Michigan	W.A. Burgis and D.F. Eschman
29	1982	Driftless Area, Wisconsin	J.C. Knox and others
30	1983	Wabash Valley, Indiana	N.K. Bleuer and others
31	1984	West-central Wisconsin	R.W. Baker
32	1985	North-central Illinois	R.C. Berg and others
33	1986	Northeastern Kansas	W.C. Johnson and others
34	1987	North-central Ohio	S.M. Totten and J.P. Szabo
35	1988	Southwestern Michigan	G.J. Larson and G.W. Monaghan
36	1989	Northeastern South Dakota	J.P. Gilbertson
37	1990	Southwestern Iowa	E.A. Bettis III and others
38	1991	Mississippi Valley, Missouri and Illinois	E.R. Hajic, W.H. Johnson, and others
39	1992	Northeastern Minnesota	J.D. Lehr and H.C. Hobbs
40	1993	Door Peninsula, Wisconsin	A.F. Schneider and others
41	1994	Eastern Ohio and western Indiana	T.V. Lowell and C.S. Brockman
42	1995	Southern Illinois and southeast Missouri	S.P. Esling and M.D. Blum
43	1996	Eastern North Dakota and northwestern Minnesota	K.I. Harris and others
44	1998	North-central Wisconsin	J.W. Attig and others
45	1999	North-central Indiana and south-central Michigan	S.E. Brown, T.G. Fisher, and others
46	2000	Southeastern Nebraska and northeastern Kansas	R.D. Mandel and E.A. Bettis III
47	2001	Northwestern Ontario and northeastern Minnesota	B.A.M. Phillips and others
48	2002	East-central Upper Michigan	W.L. Loope and J.B. Anderton
49	2003	Southwestern Michigan	B.D. Stone, K.A. Kincare, and others
50	2004	Central Minnesota	A.R. Knaeble, G.N. Meyer, and others
51	2005	North-central Illinois	E.D. McKay, R.C. Berg, and others
52	2006	Northwest-central North Dakota	L.A. Manz and others
53	2007	East-central Wisconsin	T.S. Hooyer and others

No meetings were held in 1968, 1970, 1974, 1977, and 1997. Meeting numbers in parentheses have been listed previously as "U" or unnumbered. The 1952 meeting that is commonly included in the list of Midwest FOP meetings as southwestern Ohio was actually an Eastern FOP meeting in central Ohio, to which Midwest Friends were invited by Dick Goldthwait the previous week in western Illinois.