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Quaternary Geology of Calumet and Manitowoc Counties, Wisconsin

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Front: Brillion esker, © William Mode
Back: Brillion esker sediment, © William Mode

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Abstract

All of what is now Calumet and Manitowoc Counties was covered by ice during the last great glaciation, known as the late Wisconsin Glaciation, between about 25,000 and 10,000 radiocarbon years (30,000 and 11,000 calendar years) ago.*

The glaciers left behind landforms and sediments that profoundly influence life in these counties today. The shapes of hills and valleys, how much clay is in the soil, and where sand and gravel resources are located are just a few examples of how glaciation has influenced the landscape. The two-county area was probably covered by several earlier glaciations, but the record of these was mostly erased by the most recent glaciation.

Deposits left by this glaciation are classified into three geologic formations. The oldest is the Hayton Formation, which is only present in the subsurface. The next youngest, the generally sandy Holy Hill Formation, is at the surface in the southern part of the counties. Deposits of the Kewaunee Formation are the youngest, and are at the surface throughout the rest of the area. Kewaunee Formation sediments are generally reddish-brown and fairly clayey. The distribution of these deposits is shown in plate 1 at a scale of 1:100,000.

Several sites preserve a record of past environments. About 13,000 radiocarbon years (16,000 calendar years) ago, the glacier had retreated to the north and tundra vegetation covered the landscape. By about 12,000 radiocarbon years (14,000 calendar years) ago, the climate had warmed and spruce trees covered the landscape. The ice sheet then readvanced, burying the trees and forest litter that now make up the Two Creeks Forest Bed, an important geologic feature preserved as an Ice Age National Scientific Reserve site.

**Radiocarbon ages are given both as radiocarbon years and calendar years because both timescales are commonly used for glacial and postglacial events.*

Introduction

The landscape of Manitowoc and Calumet Counties, like much of Wisconsin, was sculpted by millions of years of erosion by rivers and at times by glaciers, then tens to hundreds of feet of sediment was laid down by glaciers and the rivers that flowed from them. The most recent of these vast ice sheets covered this area between about 25,000 and 10,000 radiocarbon years (30,000 and 11,000 calendar years) ago. Since that time, rivers have slowly downcut into this sediment, waves have eroded the Lake Michigan and Lake Winnebago shorelines, and younger organic sediment has filled, or at least partly filled, lake basins, producing wetlands.

In Calumet and Manitowoc Counties, most of the bedrock is covered by 10 to 300 ft (3 to 100 m) of diamicton, gravel, sand, silt, and clay deposited during the Wisconsin Glaciation (the last glacial episode). The nature of these surface sediments greatly affects how we use the land: sand and gravel, clay, and peat are important resources. Buildings and roads require huge amounts of aggregate (sand and gravel) during construction, most of it mined locally. How much precipitation infiltrates into the ground depends on the slope of the ground surface and on the nature of near-surface sediments. Groundwater flow paths are also influenced by surface slopes and the nature of near-surface sediments. Farming practices are greatly influenced by what was left behind by the glacier. Clayey soil derived from the most recent glacial advances is common in all but southwestern Manitowoc and southern Calumet Counties. In that area, soils are sandier and more stony and formed on somewhat older glacial deposits.

The purpose of this report is to provide an interpretation of the distribution and history of these deposits and a map showing their distribution. The distribution of sediments and landforms, and cross sections schematically portraying deposits below the surface are shown on plates 1 and 2. The remainder of this report describes the sediments and landforms shown on the map and outlines the glacial history of the area.

This report should be useful to planning and development agencies in making land-use decisions, and to the non-metallic mining industry to aid in locating potential sources of aggregate. The stratigraphic framework presented provides a basis for future detailed groundwater studies in the area. Finally, we hope that naturalists and others interested in the geologic history of the land on which they live will enjoy reading this report. We describe landform distribution and the nature of the glacial sediments. These sections provide the data necessary to understand the glacial history of the Green Bay and Lake Michigan Lobes, and the history of associated glacial lakes, which are discussed in the third major section of the report.

The science of geology has its own vocabulary and it is difficult to write a detailed report like this without using some of it. We define many geologic terms as they are used in the text and also provide a glossary with brief definitions. Terms that appear in the glossary are shown in bold type the first time they are used in the text. Some detailed explanations, mostly of interest to geologists or others with a deep interest in the topic, are presented in focus boxes. These can be skipped over without any loss of a basic understanding of the geology of the two counties.

Map and cross-section reliability

The reliability of information shown on plates 1 and 2 is variable, depending greatly on where exposures exist. All road cuts, gravel pits, shoreline bluffs, and other exposures were described and sampled. Well construction reports, USDA Soil Survey maps (Otter, 1980), and other resources were also used to compile what is known of surficial deposits in the area. These were compiled on 1:24,000 maps and reduced to 1:100,000 for publication.

Limited resources prevented collecting site-specific information from the subsurface except at a few places, so a great deal of interpretation has gone into the cross sections. In addition, geologic information for the Manitowoc County portion of plate 2 is projected into the line of the cross section from up to 1.6 mi (2.5 km) north and south, so elevations of contacts in the cross sections are schematic. For the Calumet County portion of plate 2, geologic contact elevations are the same as on the well logs used to construct the cross-sections, because the sections are not projected.

Summary of geologic history

The sediments and landforms that are the focus of this report all formed during the **Quaternary Period**, which began about 2.6 million years ago (fig. 1). However, the oldest rocks in the Great Lakes region are Precambrian-age **igneous** and **metamorphic rocks** that formed during mountain-building episodes more than a billion years ago. These rocks form the southern part of the Canadian Shield, a broad area of very old rocks that forms the nucleus of the North American continent. LaBerge (1994) provides a more detailed history of this early part of Wisconsin's geologic history, as do Dott and Attig (2004). Although none of these Precambrian metamorphic and igneous rocks are at the surface in Calumet and Manitowoc Counties, they form the basement on which younger rocks have been deposited. For purposes of groundwater studies, these Canadian Shield rocks are generally considered to have very low permeability, and thus generally do not yield significant amounts of groundwater.

For over a billion years, north-central Wisconsin has been at a higher elevation (called an arch) than areas to the east, west, and south. This arch has determined the distribution of **sedimentary rocks** deposited during the Paleozoic Era (fig. 1) as well as the path of rivers and glaciers in much more recent geologic time. All of the rocks at or near the surface in Calumet and Manitowoc Counties are Paleozoic age (Silurian and Ordovician), as shown in figure 2.

The Paleozoic Era: An overview

What is now Wisconsin was at the edge of the continent 500 million years ago when seas were advancing into this area. All of the rocks that are present at the surface and that extend several hundred meters below Calumet and Manitowoc Counties are sedimentary rocks that were deposited in this extensive sea that repeatedly expanded (transgressed) and contracted (regressed) over a period of more than 100 million years. At times, when rivers carried sandy sediment into the sea in this area, **sand** was deposited. The sand was buried and later cemented into sandstone, as more sediment accumulated on top. Ocean currents carried finer **silt** and **clay** particles into deeper water in what is now Michigan and Illinois. The silt and clay produced shale when buried and consolidated into rock. Deep basins continued to develop in Michigan and Illinois throughout much of the Paleozoic, but what is eastern Wisconsin today remained a shallow-water continental shelf when submerged by the sea. Only occasionally were the seas deep enough here for silt and clay to be deposited. Thus, little shale formed beneath these counties, but its presence in the Lake Winnebago–Green Bay lowland played an important role in determining the shape of the landscape much later, when glaciers covered the area.

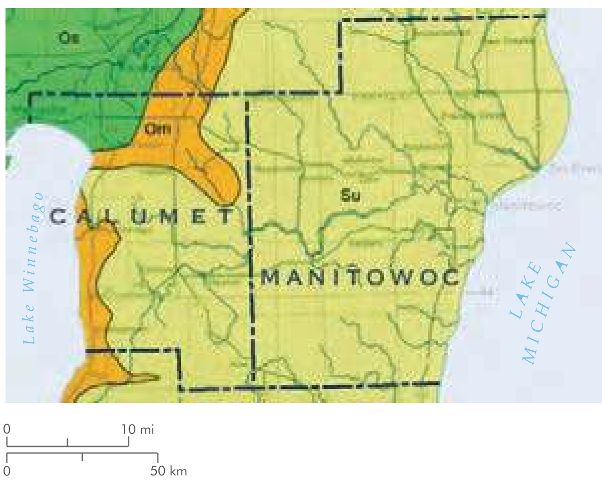
Figure 1. Geologic time scale. Surface bedrock and overlying sediments in Calumet and Manitowoc Counties are shaded gray.

GEOLOGIC-TIME TERMS			
Era	**Period**	**Epoch**	**Years ago**
Cenozoic	Quaternary	Holocene	10,000
		Pleistocene	2,600,000
	Neogene	Pliocene	5,300,000
		Miocene	23,000,000
	Paleogene	Oligocene	
		Eocene	
		Paleocene	66,000,000
Mesozoic	Cretaceous		
	Jurassic		
	Triassic		252,000,000
Paleozoic	Permian		
	Carboniferous	Pennsylvanian	
		Mississippian	
	Devonian		
	Silurian		
	Ordovician		
	Cambrian		541,000,000
Precambrian			

At times during the Paleozoic, when sediment from rivers was diverted to other areas, bits of shell, single-celled organisms, and the skeletons of calcareous algae and coral accumulated on the sea floor and were slowly cemented into limestone ($CaCO_3$, calcium carbonate) and dolomite ($CaMg(CO_3)_2$, calcium-magnesium carbonate). By Silurian time, massive coral reefs had formed (Kluessendorf and Mikulic, 1989).

When the sea level dropped, the land was exposed to erosion. Streams eroded the surface, forming river valleys across the area, removing some of the previously deposited sandstone and dolomite. This erosion surface is called an **unconformity**. When the sea rose again, sediment covered the uneven land surface, producing an uneven contact between the rock types at the unconformity. A major unconformity occurs below the base of the Silurian dolomite, where iron oxide (the mineral hematite), an end product of tropical weathering, is up to a meter thick.

With few exceptions, individual formation and member names of the Paleozoic rocks beneath Calumet and Manitowoc Counties are not discussed in the following text, but we provide an overview of rock unit characteristics. Figure 3 shows a diagrammatic cross section across east-central Wisconsin showing these



Explanation

PALEOZOIC ROCKS

Silurian System

Su — Dolomite, undivided; includes Cayugan, Niagaran and Alexandrian series.

Ordovician System

Om — *Maquoketa Formation.* Shale, dolomitic shale and dolomite; includes overlying Neda Formation (age uncertain) consisting of aolitic iron oxides and shale.

Os — *Sinnipee Group.* Dolomite with some limestone and shale; includes Galena, Decorah and Platteville Formation.

Figure 2. Bedrock geology of Calumet and Manitowoc Counties. *(From Mudrey and others, 1982.)*

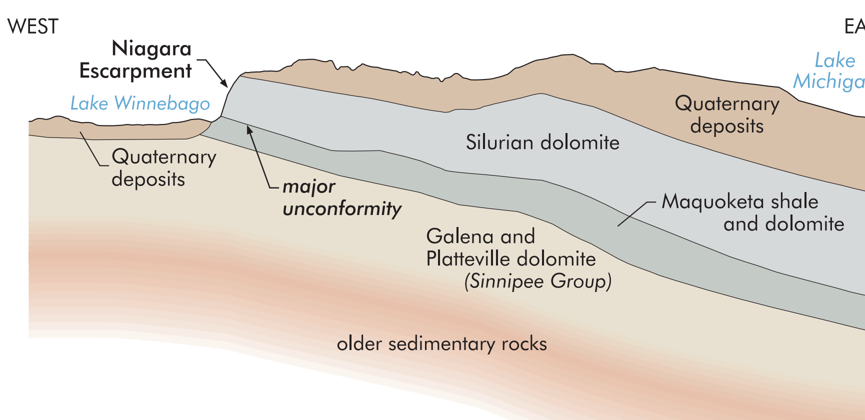


Figure 3. Sketch of geologic units beneath Calumet and Manitowoc Counties. Dip of the rocks is greatly exaggerated and diagram is not to scale. Actual dip is 30 to 40 ft/mi. *(Drafted by Mary Diman.)*

rock units. Note that the rock layers dip to the east, and that the dip is greatly exaggerated. The actual dip of the rocks is 30 to 40 ft/mi (5 to 7 m/km) (Brown, 1986). Bedrock elevation is shown in figure 4.

West of Calumet County the surface rocks are those of the Sinnipee Group, which consists of Platteville dolomite, overlain by generally thin Decorah shale, which in turn is overlain by the Galena dolomite. The Sinnipee Group is overlain by the Maquoketa Formation, which is a unit that contains mostly soft shale, although there are some dolomite layers. The unit was mined for bricks in several places in the Fox River–Lake Winnebago lowland as recently as the 1960s. Most of the shale has been eroded away by preglacial streams and glacier ice. In other areas it is covered by glacial deposits, so its outcrop area is very limited. A thin iron deposit lies at the unconformity on top of the Maquoketa Formation. This iron deposit probably is an old soil formed under tropical weathering conditions during the interval of erosion and weathering at the end of Ordovician time. The iron consists of oolites (tiny spheres) probably produced as the Silurian sea advanced over the landscape, reworking the iron-rich tropical soils.

The next youngest Paleozoic rock, and nearest the surface in Calumet and Manitowoc Counties, is Silurian-age dolomite (figs. 2 and 3). The 410- to 440-million-year-old dolomite is dense, but cracks in the rock produce a modest permeability (Krohelski, 1986) and many domestic wells draw water from this dolomite. Krohelski considers this rock unit and the glacial deposits above to be the surface aquifer. By Silurian time corals began to form major reefs. Brachiopods, other corals, and mollusks are abundant in these rocks, and can be found throughout this area. Reefs are made up of rock that is more resistant than adjacent rocks, and many of the hills in Calumet and Manitowoc Counties stand higher than the surrounding terrain because they are composed of reef dolomite. Many rock quarries, for

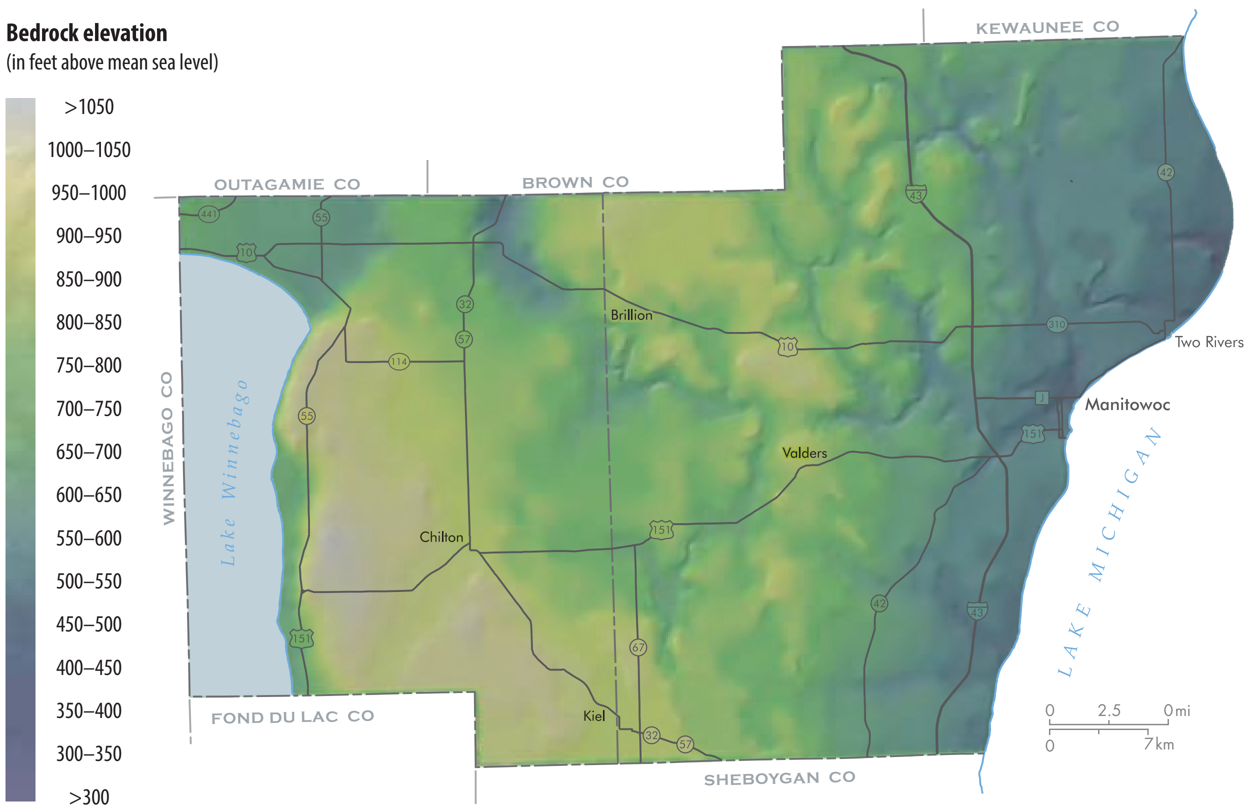


Figure 4. Bedrock elevation map of Calumet and Manitowoc Counties. *(From Wisconsin Geological and Natural History Survey, unpub. data.)*

instance the huge quarries at Valders and Brillion (locations shown in fig. 5), are in coral reef deposits. The Silurian rocks of northeastern Wisconsin have been studied and further classified by Kluessendorf and Mikulic (1989), but are not discussed in detail here. Their major impact on the landscape is discussed in the next section.

Following retreat of the sea in which the Silurian dolomite was deposited, the sea rose again in the Devonian Period. Except along a small sliver of eastern Wisconsin and beneath Lake Michigan, the rocks deposited during the Devonian Period have been eroded away. There may have been other, more-recent incursions of the sea into Wisconsin, but all evidence of these younger events has also been erased from the geologic record. From Devonian time until glaciers first covered the surface, perhaps a million years ago, Wisconsin was a landmass slowly eroding away much like it is today. River valleys were cut, and many of these valleys remain buried below thick glacial deposits. Little is known of the early drainage history of eastern Wisconsin, but Martin (1916) speculated that streams flowed southeast, draining into a river valley (fig. 6). Although most of the evidence is gone, it seems clear that what is now Lake Michigan was a river valley in preglacial time.

It is likely that the Fox River–Lake Winnebago lowland was also a river valley in preglacial time. The Maquoketa is mostly shale, which, compared to Silurian dolomite, was relatively easily eroded by preglacial streams and by glaciers flowing southwestward, parallel to the axis of the lowland. The present Lake Winnebago, Lower Fox River, and Green Bay are located in the deepest part of this lowland.

This differential erosion has produced the Niagara Escarpment, a steep, west-facing slope in western Calumet County, which bounds an upland area that slopes gently eastward to Lake Michigan. The escarpment rises to about 300 ft (100 m) high and bounds the east side of Lake Winnebago and Green Bay (fig. 7). The escarpment is the surface-water divide between the Fox River and Lake Michigan drainage basins in Calumet County, and forms the backbone of Door County, extending to the northeast across the north side of Lakes Michigan and Huron to Niagara Falls. The Niagara Escarpment had a major control on glacier ice coming into Wisconsin, as discussed in a later section.

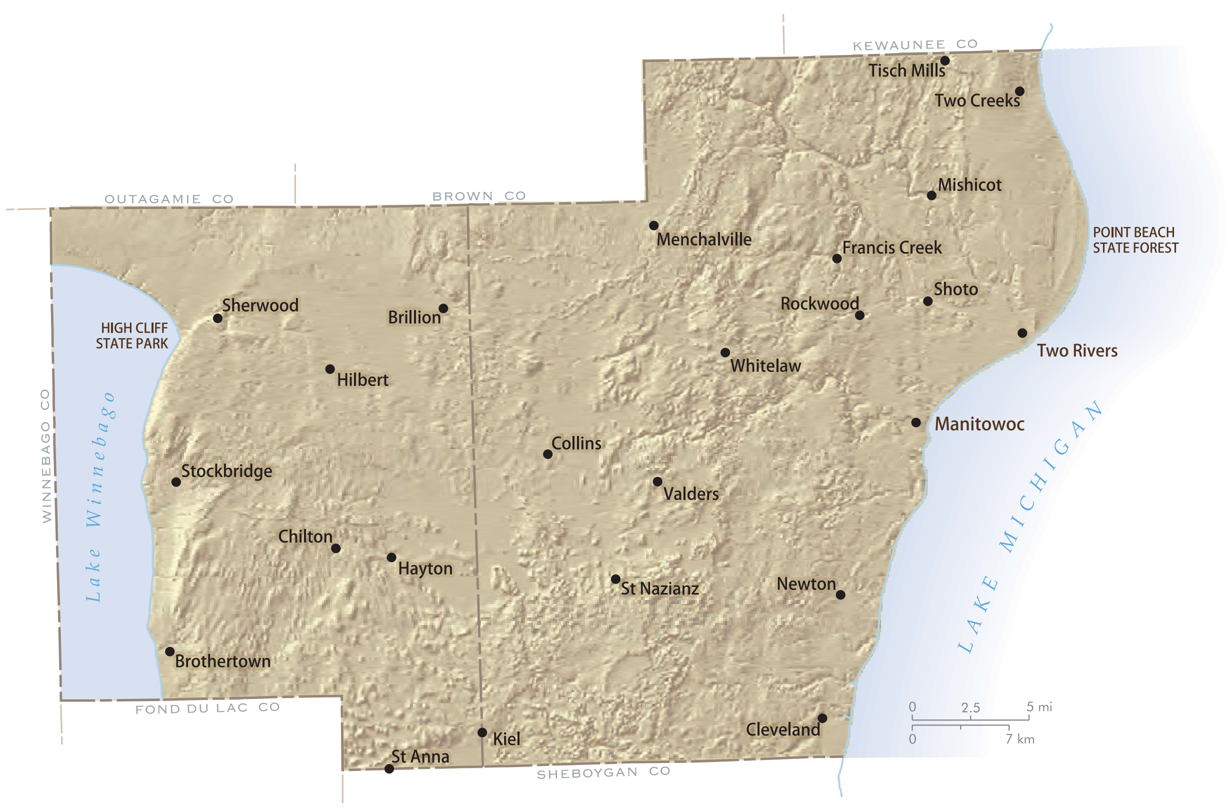


Figure 5. Place names in Calumet and Manitowoc Counties.

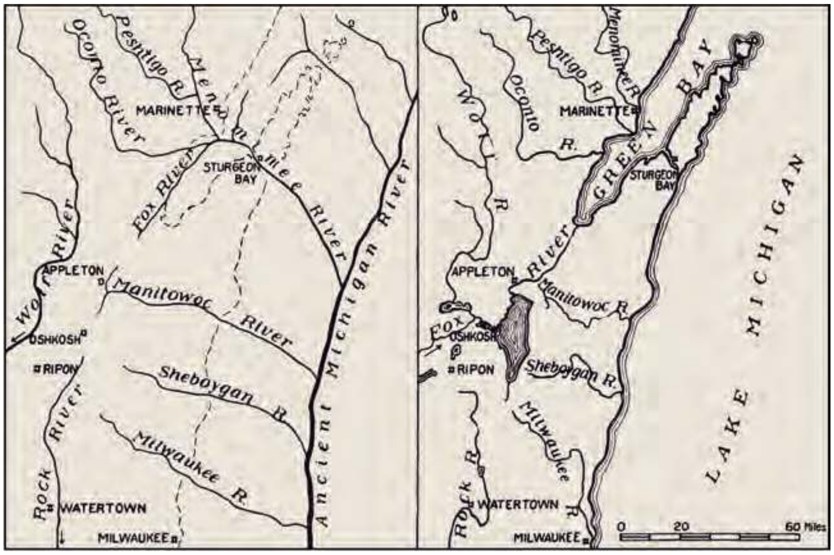


Figure 6. Reconstructed preglacial stream pattern (left) and present-day drainage (right). Although more detail of the topography of the bedrock surface is now known (see fig. 6), we have no better idea of the pattern of pre-glacial drainage. *(From Martin, 1916.)*

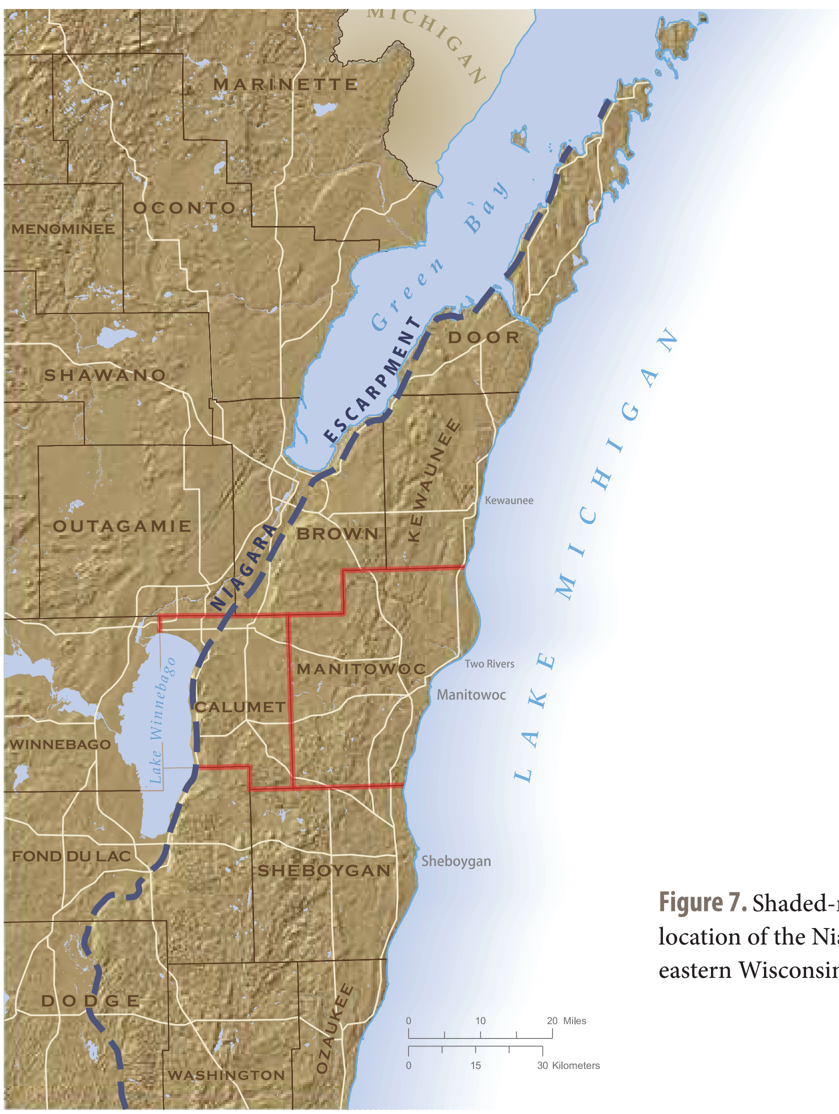


Figure 7. Shaded-relief map showing location of the Niagara Escarpment in eastern Wisconsin.

Overview of Quaternary history

During the Quaternary Period, about the last 2.6 million years, ice sheets likely covered this part of Wisconsin numerous times. However, only glacial deposits from the most recent glaciation are present at the surface in the two-county area. Localized deposits associated with older glaciations are likely present in the subsurface, buried under deposits of the last glaciation (table 1). In particular, there is gray, very compact, silty diamicton (a poorly sorted mixture of boulders, cobbles, sand, silt, and clay) that is present in the subsurface beneath part of Calumet and Manitowoc Counties. This diamicton, belonging to the Hayton Formation, was deposited either during one or more glacial advances prior to the last glaciation, or it may have been deposited early during the advance phase of the last glaciation.

During the last major phase of ice sheet expansion, called the late Wisconsin Glaciation, the southern margin of the Laurentide Ice Sheet had a distinctly lobate shape as it was channeled down the Green Bay and Lake Michigan lowlands (fig. 8). The maximum extent of the Green Bay and Lake Michigan Lobes was probably reached about 23,000 radiocarbon years (30,000 calendar years) ago and covered all of what is now Calumet and Manitowoc Counties with several hundred meters of ice (Socha and others, 1999).

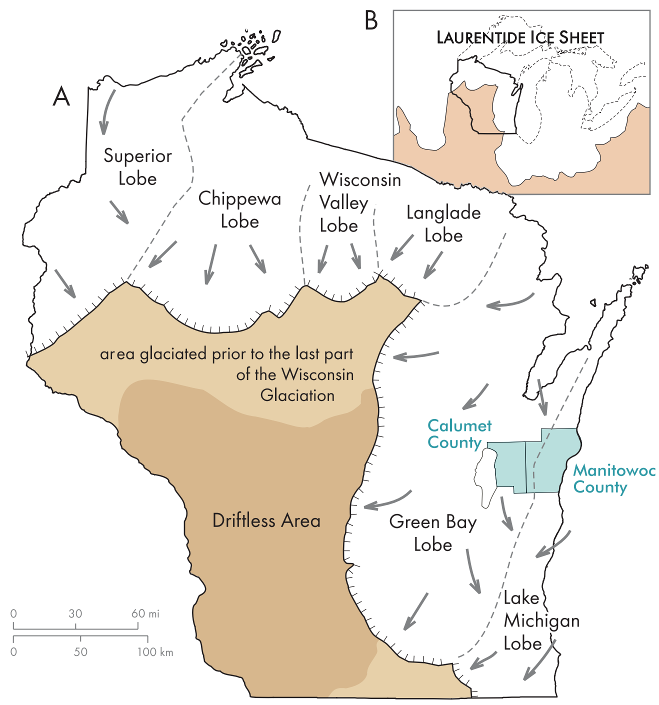


Figure 8. Map showing glacial lobes at maximum extent of the Wisconsin Glaciation. Calumet and Manitowoc Counties are highlighted. Hachures indicate the edge of the ice sheet; arrows indicate direction of ice flow.

Table 1. Quaternary lithostratigraphic units of Calumet and Manitowoc Counties. Units are arranged by age (youngest at the top) and by their association with the Green Bay and Lake Michigan Lobes.

LITHOSTRATIGRAPHIC UNITS	
Green Bay Lobe	**Lake Michigan Lobe**
KEWAUNEE FORMATION	
Glenmore Member	Two Rivers Member
Two Creeks Forest Bed	
Chilton Member	Valders Member
Branch River Member	Ozaukee Member
HOLY HILL FORMATION	
Horicon Member	New Berlin Member
HAYTON FORMATION	
Cato Falls Member	
High Cliff Member	

The ice sheet eroded the landscape, removing most older deposits and the soils that existed on them. It also eroded the bedrock by plucking large blocks, as well as abrading the rock, producing **striations**, scratches that can be seen on many bedrock outcrops (fig. 9).

The retreating ice sheet left behind sediments, including diamicton, sand, gravel, silt, and clay. The retreat wasn't continuous, but was interrupted by readvances of the glacier (fig. 10). In many cases these readvances reworked older deposits and deposited diamicton of different composition. As a result, the diamicton layer deposited by one advance can be distinguished from that deposited by another advance. Generally sandy and stony diamicton of the Holy Hill Formation was deposited by the main late Wisconsin advance. During later readvances, clayey, reddish-brown diamicton of the Kewaunee Formation was deposited, until the glacier finally retreated from this area for the last time about 11,000 radiocarbon years (13,000 calendar years) ago.

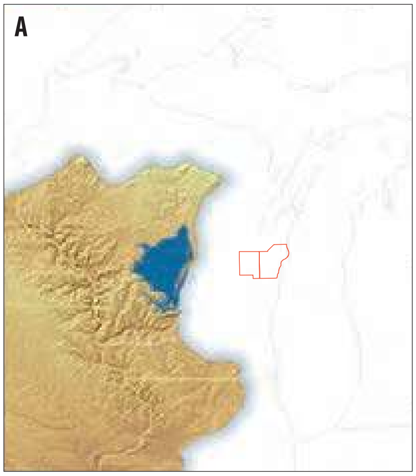
The water level in Lake Michigan fluctuated by over 300 ft (100 m) during the period of ice retreat. In general, the lake was higher during periods of glacial advance and dropped during ice retreat. In the Green Bay lowland, a shallower glacial lake, called glacial Lake Oshkosh, developed in front of the retreating Green Bay Lobe.

Understanding plates 1 and 2

Glaciers are directly or indirectly responsible for modifying much of the landscape and depositing most of the sediments in Calumet and Manitowoc Counties. Landforms and sediments are classified in various ways by geologists, and we combine several types of information in the map units used on plate 1. In order to describe the vertical and horizontal distribution of different layers they are classified as lithostratigraphic units called formations and members, much like the layers of Paleozoic rock discussed earlier. Formations are fairly uniform units of sediment that extend over many square miles and that are distinguishable from each other in the field without laboratory testing (Mickelson and others, 1984; Syverson and others, 2011). Members are subsets of formations and consist of units of diamicton, sand and gravel, and silt and clay that are distributed over more limited areas than formations and that may not be distinguishable without laboratory testing. Generally, the first letter of the map unit on plate 1 (except for a, o, n r, and w, which are not formally named units) designates the formation or member.

Figure 9. Photo of striations on horizontal bedrock surface at Valders Quarry. Glacier moved parallel to long axis of compass. Compass is about 3 inches wide.

Figure 10. Extent of glaciers and large lakes at six times in the past (dates shown are calendar years and are approximate). *(From Mickelson and Attig, 2017.)*



24,000 years ago – Ice at maximum extent
Glacial Lake Wisconsin was dammed along the western edge of the Green Bay Lobe. The Lake Michigan basin was buried below thick glacier ice.

16,000 years ago – First major retreat
The glacier retreated north of the Lake Michigan basin during a period known as the Mackinaw Interstadial. Lake Michigan drained to the east and was likely about 300 ft (100 m) lower than today.

15,250 years ago – Ice readvance
Ice reached its early Port Huron ice-margin position with the Lake Michigan Lobe terminating near Milwaukee and the Green Bay Lobe terminating near the north end of Lake Winnebago. Ozaukee and Branch River tills were deposited at this time. Glacial Lake Oshkosh drained into the Wisconsin River valley and Lake Michigan, which was at the Glenwood level, drained at Chicago.

14,600 years ago – Minor retreat, readvance
After a minor retreat, both lobes readvanced to the late Port Huron ice-margin position. Valders and Chilton tills were deposited at this time. Glacial Lake Oshkosh drained into the Wisconsin River valley and Lake Michigan remained at the Glenwood level and drained through the Chicago outlet.

14,300 years ago – Second major retreat
The glacier again retreated north of the Lake Michigan basin during a period known as the Two Creeks warm time. Glacial Lake Oshkosh had shrunk to the size of present-day Lake Winnebago, Green Bay was dry, and Lake Michigan drained to the east. Lake level was likely about 300 ft (100 m) lower than today.

13,000 years ago – Final advance
The final glacial advance into eastern Wisconsin was the Two Rivers advance, which reached its maximum extent shortly before 13,000 years ago. Glenmore and Two Rivers tills were deposited at this time. Glacial Lake Oshkosh was present in front of the Green Bay Lobe and Lake Michigan was at the Calumet level and drained through the Chicago outlet.

The two formations mapped at the surface in Calumet and Manitowoc Counties are the Holy Hill and Kewaunee Formations (Syverson, 1997; Syverson and others, 2011). The Hayton Formation is not exposed at the surface, but is shown on cross sections. Each formation is subdivided into members. In Calumet and Manitowoc Counties, members of the Holy Hill Formation (H) are not distinguished. However, Branch River (B), Chilton (C), Valders (V), Glenmore (G), and Two Rivers (T) are identified as members of the Kewaunee Formation. The Ozaukee Member is shown in cross sections, but is not exposed at the surface. Throughout this report, we informally use member names to refer to the event that deposited a member and the ice that deposited that unit (for example, the Valders advance and Valders ice) in addition to formally named events or moraines.

The smallest lithostratigraphic unit formally recognized is the bed. The Two Creeks Forest Bed, present in several places in Calumet and Manitowoc Counties, is the only bed of Quaternary age formally defined in Wisconsin.

The type of sediment included in a map unit is indicated by the second letter. We use descriptive and genetic terms to describe deposits. **Diamicton** (d) describes sediment containing a wide range in grain size (clay to boulders) that is commonly interpreted as **till**, a genetic term for sediment deposited directly by the glacier. Because all of the diamicton of the Kewaunee Formation exposed or described from drill holes is interpreted to be till, we often use that term in the text as a short version of "diamicton interpreted to be till." Sand (s) and gravel (g) are descriptive terms that are commonly interpreted as **outwash** deposited by streams flowing away from the glacier, or **esker** sediment deposited by streams flowing through tunnels in the glacier. Sediment composed of finer particles, silt or clay (i), is commonly interpreted as lake sediment. Landforms, such as beach (b) or hummocky end moraine (he), are indicated by the third and fourth letters in the unit names on plate 1.

Dave Mickelson

Landform distribution and genesis

Because landforms are not randomly distributed in the landscape, it makes sense to identify areas in the two counties that share similar geologic characteristics and history (physiography), and discuss landforms in that context. In this section we briefly define landform terms where they are first used, discuss how they formed (their genesis), provide reference to locations of landforms on plate 1, and illustrate examples on topographic maps. Figure 11 shows the regions covered by each of the landforms.

Fox River lowland

The Fox River lowland (fig. 11, area 1) extends from south of Lake Winnebago through western Calumet County and southeastern Outagamie County, northeastward to Green Bay in Brown County. The lowland slopes to the northeast from an elevation of about 800 ft (244 m) at the north end of Lake Winnebago to about 580 ft (177 m) at the mouth of the Fox River at Green Bay. The area was scoured by glaciers as the ice advanced from the north. As the ice margin retreated, a glacial lake (called glacial Lake Oshkosh) formed in the lowland because drainage out of Green Bay to Lake Michigan was blocked by ice.

The lowland is underlain by silty, sandy, and clayey lake sediment and till. In many places bedrock (mostly Ordovician-age dolomite) is at or near the surface. The major rivers and streams in the lowland, including Duck Creek, Ashwaubenon Creek, and the East River all parallel the Silurian dolomite and other geologic contacts (figs. 2 and 4), indicating the drainage pattern is influenced by the bedrock.

Figure 11. Shaded-relief map of Calumet and Manitowoc Counties showing physiographic regions:

1. Fox River lowland
2. Niagara Escarpment
3. Streamlined landscape (Holy Hill Formation)
4. Kettle Moraine
5. Palimpsest landscape (beneath Chilton and Valders Members)
6. Lake Michigan Lobe moraines
7. Manitowoc River–Twin Rivers lowland and Point Beach
8. Two Rivers moraine

Locally, the lowland is dissected by streams that cut down deeply through the lake sediment and till as lower outlets into Lake Michigan were uncovered by the retreating ice margin. As the ice retreated north of Lake Winnebago, several of these outlets into Lake Michigan were exposed. These were the Manitowoc, the Neshota, the West Twin, the Kewaunee, and the Ahnapee Rivers (Hooyer, 2007). (See focus box 1 for a more detailed history of glacial Lake Oshkosh).

The lowland to the northwest of the Niagara Escarpment in northwestern Calumet County (fig. 11, area 1) is mostly underlain by clayey till that forms a broad low ridge at the north end of Lake Winnebago. The ridge is an end moraine formed by the post–Two Creeks Glenmore advance, and is an extension of the Denmark moraine on the upland to the east. As the ice margin retreated to the north from this end moraine, lake sediment was deposited in front of the ice margin. However, lake sediment is present only locally in the lowland in Calumet County and it is generally thin.

Niagara Escarpment

In southwestern Calumet County the rise from Lake Winnebago to the top of the Niagara Escarpment is a smooth slope, unbroken by cliffs (fig. 11, area 2). From about the point where Highway 151 climbs the escarpment northward there are Silurian dolomite cliffs. Several caves are present along the cliff face and many springs that flow along lower parts of this slope indicate a substantial discharge of groundwater from the Silurian dolomite. There is a break in the cliff near Stockbridge, then a steep cliff continues north to Sherwood, where it again gives way to a more-gentle rise in elevation.

Streamlined landscape (Holy Hill Formation)

The landscape of southern Calumet County (fig. 11, area 3) is dominated by south- and southeast-trending, elongate, streamlined hills called **drumlins**. On plate 1, arrows show

1 FOCUS BOX Fox River lowland and glacial Lake Oshkosh

The preglacial Fox River flowed in a deep bedrock river valley, the Marquette bedrock valley, which trends generally northeast to southwest through Green Lake and Marquette Counties (Stewart, 1976). Advances of the Green Bay Lobe blocked the river with ice and glacial deposits. During the late Wisconsin glacial advances and retreats, glacial lakes, collectively known as glacial Lake Oshkosh, formed in the Fox River lowland. This produced a low, fairly flat lake plain between Lake Winnebago and Green Bay.

Progression of drainage outlets

At glacial maximum, all of the area of glacial Lake Oshkosh was buried under ice. As the ice margin retreated, the first outlet for Lake Oshkosh was the Dekorra outlet northeast of Portage in Columbia County (sec. 16, T13N, R9E) at an elevation of about 790 ft (241 m) (Hooyer, 2007). Water from glacial Lake Oshkosh merged with water in the Wisconsin River and flowed to the southwest to another narrow spillway at an elevation of about 790 ft (241 m) near Dekorra in Columbia County (sec. 12, T11N, R8E).

As the ice retreated north of the Lake Winnebago basin and into the Fox River lowland, successively lower outlets into Lake Michigan were uncovered. The first was into the Manitowoc River in Calumet County, just north and east of Sherwood. The next to be occupied was the Neshota outlet into the West Twin River (elevation 765 ft (233 m)), followed by the Kewaunee River (elevation 685 ft (209 m)) and the Ahnapee River (elevation 635 ft (194 m)) (Hooyer and Mode, 2008).

Challenges tracing history of glacial Lake Oshkosh

The history of glacial Lake Oshkosh and the sequence of events are complicated because new lakes formed with each advance and retreat of the glacier. The record of all but the final series of lakes was mostly destroyed by the latest advances.

Another complication is that the earth’s crust was depressed due to the load of the ice. This tilted northeastern Wisconsin toward the northeast where ice was thickest. The ice advanced several times during the Pleistocene, and crustal rebound may not have been complete before the ice readvanced (Clark and others, 1990, 1994, 2008). Although the pattern of crustal depression and rebound is very complicated and highly speculative, it is known that each time the ice advanced out of the Green Bay lowland, it advanced up the regional slope and water was ponded between the ice front and the uplands. As the ice retreated, the lakes drained via several successively lower outlets. These controlling outlets have been identified by Thwaites and Bertrand (1957), Wielert (1980), Hooyer (2007), and Hooyer and Mode (2008).

the crests of the drumlins and map unit *Hds* is used to delineate the landscape. Drumlins rarely occur alone, and a group of drumlins is called a drumlin field. This drumlin field, called the St. Charles drumlin field, formed at the base of the Green Bay Lobe ice, by a complex set of processes involving erosion and deposition. Thousands of these elongate hills are present southward from here, in Fond du Lac, Dodge, and Jefferson Counties. It seems likely that they formed when the Green Bay Lobe was advancing to, or was at, its maximum position (Colgan and Mickelson, 1997), and that the molded topography is a good representation of the shape of the bottom of the glacier as it moved across this area (fig. 12).

The surface sediment in the hills is primarily diamicton interpreted to be till, but the internal composition of the drumlins varies greatly. Some have a bedrock core, some a gravel core, some are all diamicton. There are relatively few exposures and drill holes, but it appears that the drumlins in Manitowoc and Calumet Counties are either rock cored or primarily diamicton. Many of the St. Charles drumlins are bedrock cored or have a core of very compact Hayton diamicton, with only a thin veneer of Holy Hill deposits on top. Between the drumlins are lowlands, some of which are still peat covered, but many of them have been drained for agriculture.

The South Branch of the Manitowoc River flows northward in a broad valley through the drumlins. The valley has a nearly flat bottom and was a shallow lake when ice was retreating from its maximum extent, and again when it stood at the Chilton ice margin. Sediments deposited near the ice margin occur near the southern boundary of Calumet County in the form of an end moraine of low-relief hummocky topography (*Hdh* on plate 1) with numerous poorly

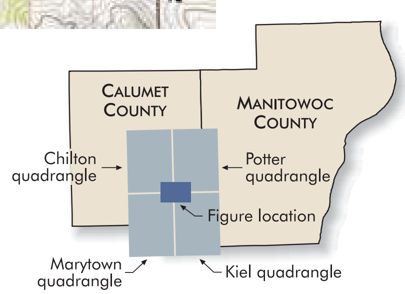


Figure 12. Topographic map of streamlined topography in southern Calumet County, from the U.S. Geological Survey, 7.5-minute, Chilton, Kiel, Marytown, and Potter quadrangles. Note that ice flow was from the northwest. The dolomite hills (left) have very thin till cover; drumlins, shown by arrows, may also have dolomite close to the surface.

drained depressions. **Hummocky topography** is a term used to describe a land surface with irregularly shaped hills and depressions. An **end moraine**, commonly just called a **moraine**, is a ridge composed mostly of diamicton that forms when the glacier edge remains in one place long enough for sediment being brought to the ice margin to accumulate as a ridge, much as a pile accumulates at the end of a conveyor belt. The narrow ridge, which extends from the southwest corner of the county, through St. Anna, and northeastward to just north of Kiel (fig. 13), was named the St. Anna moraine by Alden (1918). A less well-developed moraine is present about 2 mi (3.2 km) north of the St. Anna moraine. It is too small to be mapped except as a symbol on plate 1, and it appears to join the St. Anna moraine about a mile west of Kiel (plate 1).

The Kettle Moraine

The Kettle Moraine (fig. 11, area 4) was deposited between the Green Bay and Lake Michigan Lobes of the Laurentide Ice Sheet during retreat from their maximum positions. This zone of **interlobate** deposits extends from Walworth County in southeastern Wisconsin to northern Manitowoc County, where it is still recognizable, but covered by younger glacial deposits. It is characterized by a hummocky ridge for much of its 125-mi (200 km) extent, and in southern Manitowoc County it consists mostly of high-relief hummocky gravel (*Hgh*). It extends from the Sheboygan County line about 2 mi (3.2 km) east of Kiel northward to just north of St. Nazianz. North of St. Nazianz the Kettle Moraine is present, but it is subdued because it was overridden by a later ice advance. Where it is not covered by younger deposits, numerous small depressions are present in the hummocky topography (fig. 14). Each of these formed when a mass of glacial ice that was buried by gravel later melted out to form a depression called a **kettle**. Surrounding the high-relief hummocky topography are lower-elevation **pitted outwash**

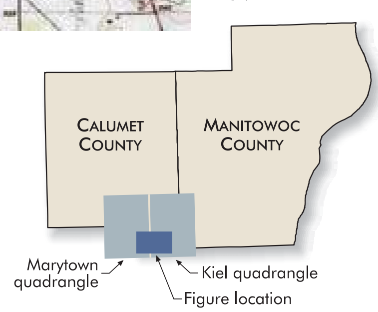


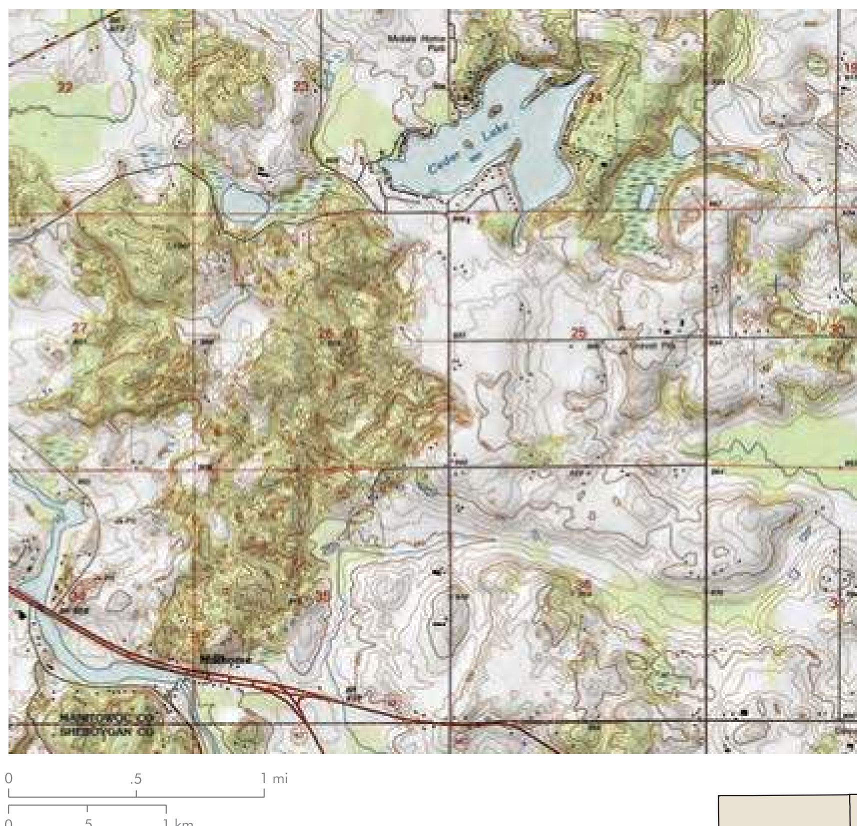
Figure 13. Topographic map showing St. Anna moraine in southern Calumet County, from the U.S. Geological Survey, 7.5-minute Kiel and Marytown quadrangles. Dashed line indicates outermost moraine position.

surfaces (*Hgpp* on plate 1). The "pits" are kettles, and the areas between are the smooth, gently sloping surfaces of the original outwash stream bed.

Chamberlin (1877) named the Kettle Moraine and attributed its formation to stream deposition in a trough between the two lobes during ice retreat. The name *moraine* is a misnomer because the Kettle Moraine is made up almost entirely of sand and gravel while moraines, in a strict sense, are composed mostly of till (Mickelson and others, 1983). However, the name is thoroughly ingrained in the literature and popular usage, thus we continue to use it here as a proper name. Alden (1918) assigned more specific names to landforms in the Kettle Moraine, but maintained the mainly proglacial model of Chamberlin (1878), which subsequent research has supported (Thwaites, 1946; Thwaites and Bertrand, 1957; Black, 1969). Attig (1986) and Carlson and others (2005, 2011) suggested a different genesis for the double-ridged part of the northern Kettle Moraine in Sheboygan County, but the Kettle Moraine in Manitowoc County is mostly a single hummocky ridge.

Palimpsest landscape (beneath the Chilton and Valders Members)

Much of central and northern Calumet and Manitowoc Counties is characterized by relatively thin, reddish-brown diamicton that appears to have been draped over pre-existing topography without modifying it very much (fig. 11, area 5). In this **palimpsest** landscape, the glacial landforms that can be seen now were formed by an earlier glacial event, then overridden. The implication is that the ice of the later advance eroded and transported relatively little sediment, either because it was very short-lived or because conditions



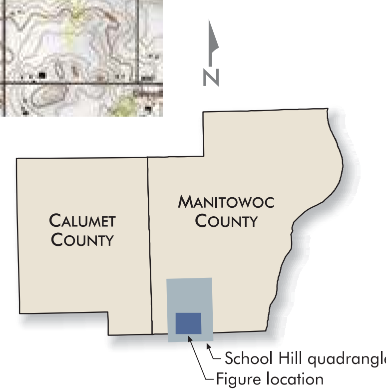


Figure 14. Topographic map showing northern Kettle Moraine and many kettle lakes formed when buried ice blocks melted out. From the U.S. Geological Survey, 7.5-minute School Hill quadrangle.

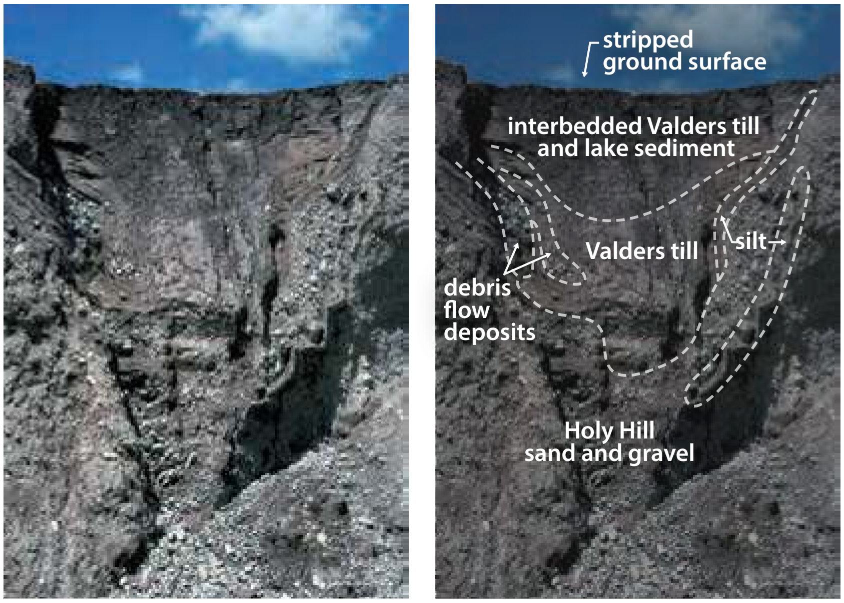


Figure 15. Photo from Fricke gravel pit in eastern Manitowoc County showing a kettle hole in collapsed Holy Hill Formation gravel filled with Valders till and some laminated silty lake sediment. The Valders sediment collapsed into the kettle as the buried ice block melted. The collapsed layering indicates that buried ice remained until after the Valders till was deposited.

at the base of the ice allowed ice to slide over a soft bed without incorporating or depositing much sediment. Much of the diamicton-capped landscape has a gently rolling surface without any particular arrangement or alignment of features. The map units *Vdr* and *Cdr* are used to portray this landscape on plate 1.

Streamlined landscape

As Thwaites and Bertrand (1957) pointed out, drumlins of the earlier advances that are formed of sandy and silty brown and gray tills were hardly modified by the glacier that deposited the Chilton and Valders Members. These palimpsest drumlins are present north of Chilton (*Cds* on plate 1) and in the vicinity of Valders (*Vds and Vdsb* on plate 1). The south-southeast orientation of the drumlins north of Chilton is consistent with the late glacial maximum ice flow toward the interlobate zone now marked by the Kettle Moraine. However, the dominantly north–south orientation of the drumlins north of Valders (plate 1) is not consistent with the orientation and location of the Kettle Moraine, extending from the vicinity of St. Nazianz, where it is buried by Valders diamicton, northward near Whitelaw and farther north to a hummocky gravel deposit in the mile (1.5 km) south of Denmark.

Hummocky topography

Within the area covered by the Valders and Chilton Members of the Kewaunee Formation, there are areas of fairly high-relief hummocky topography that seem inconsistent with a glacial advance that did little geologic work and carried little sediment, especially because most of these areas are not related to former ice-margin positions of the Valders or Chilton ice. It is clear that except for the Menchalville moraine (discussed below), these hummocky areas are also palimpsest features. In fact, in several exposures it is clear that glacier ice was buried in sand and gravel during the Holy Hill advance or retreat and remained there when the glacier retreated. The glacier then readvanced to deposit the younger Kewaunee Formation sediments before the buried ice blocks melted out. The gravel pit shown in figure 15 is in an area of high-relief hummocky topography in Newton Township, about 4 mi (6 km) southwest of the city of Manitowoc. Exposures in these gravel pits clearly show that the sand and gravel belongs to the Horicon Member of the Holy Hill Formation, and that ice blocks buried in this gravel melted out after deposition of the Valders diamicton. Evidence of this is the collapse of clayey red-brown diamicton of the Kewaunee Formation into kettles as buried ice blocks melted out (fig. 15).

The hummocky ridge (fig. 11, area 5) extending from St. Nazianz north through Whitelaw contains thick sand and gravel that is discontinuously capped with thin (typically less than 3 to 10 ft (1 to 3 m)) Valders diamicton. If this is the overridden Kettle Moraine, one would expect a more southeasterly trend to the drumlins north of Valders, suggesting that the drumlins formed earlier, during the late-glacial maximum advance. What we, as did Thwaites and Bertrand (1957), interpret to be the buried Kettle Moraine formed during retreat of that ice, but before the Chilton and Valders advances.

Another possibility is that a zone of hummocky sand and gravel overridden by red-brown diamicton that extends northwestward from St. Nazianz (fig. 16) represents a branch of the interlobate zone that was active when the drumlins near Valders formed. This was the interpretation of Alden (1932), but Thwaites and Bertrand (1957) interpreted this hummocky zone as a younger ice-marginal deposit. Our interpretation is that this hummocky zone is the interlobate junction between the Valders and Chilton ice masses and not primarily a palimpsest feature at all. It contains sand and gravel capped with about 6 ft (2 m) of red-brown diamicton just north of Highway 151, and thicker reddish-brown diamicton just north of there.

The low ridge that extends from the hummocky topography shown in figure 16 northward and then northeastward through Maple Grove and Menchalville appears to be mostly Valders diamicton (fig. 17). We interpret this ridge south of Maple Grove as an interlobate moraine where the Lake Michigan Lobe deposited Valders diamicton in contact with the Green Bay Lobe, which deposited Chilton diamicton. This was called the Menchalville moraine by Black (1980). Northeast of Maple Grove the Valders ice margin was partially in a shallow lake (fig. 17; plate 1). In the figure, note that the Branch River becomes much less winding downstream (south) of the Menchalville moraine compared to the very sinuous stream as it flows across the low-sloping former lake bed.

Outwash plains

Sand and gravel deposits are also included in the Valders and Chilton Members, and several map units of gravel and sand in outwash plains are designated on plate 1. The sediments in all the units are moderately well-sorted, well-**stratified**, gravel and sand deposited by glacial

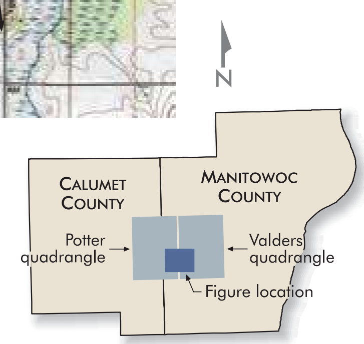
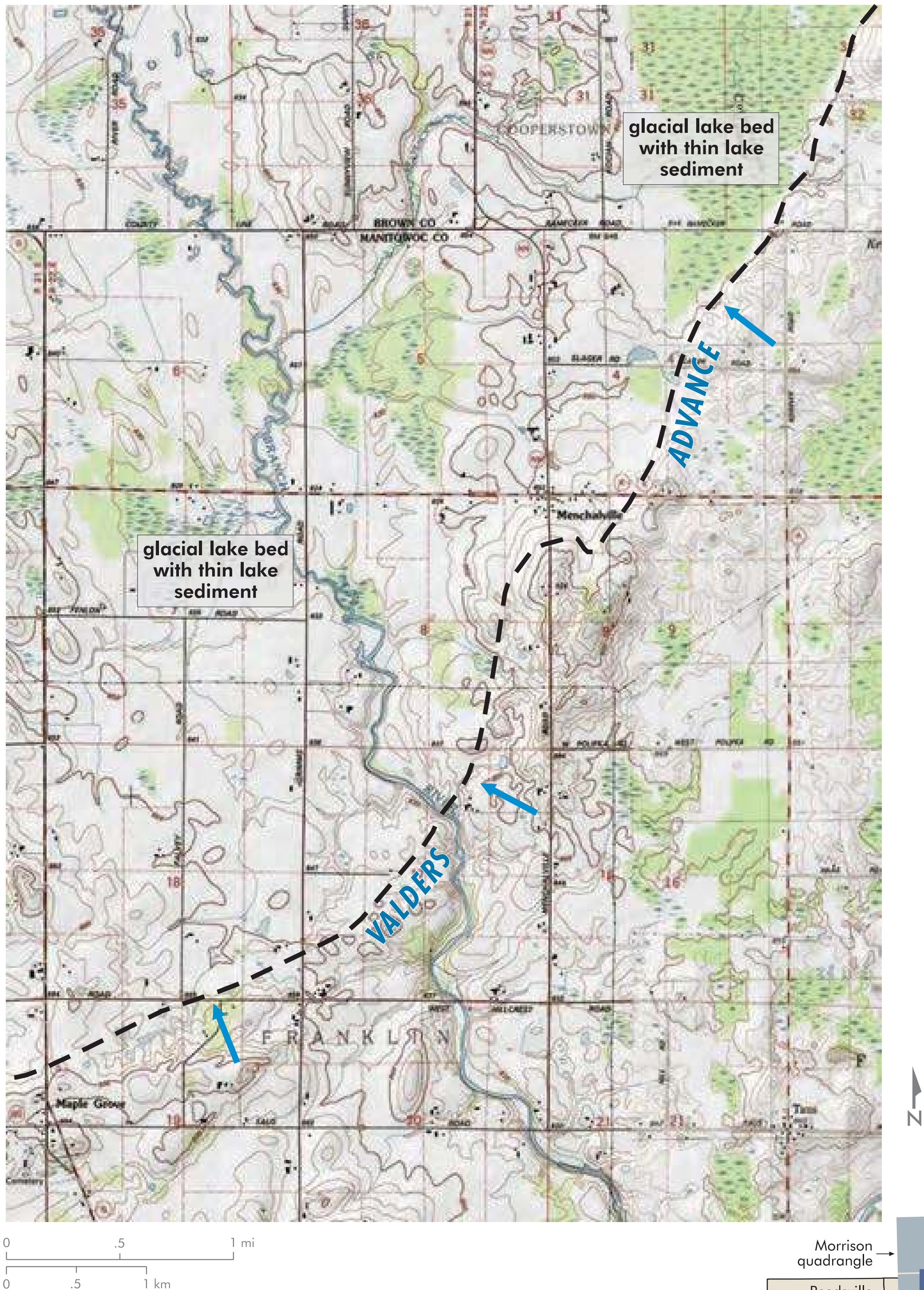


Figure 16. Topographic map of hummocky sand and gravel in the interlobate Menchalville moraine in southwestern Manitowoc County. The ice advanced from the north (blue arrows on left) to the east–west dashed line, depositing Chilton diamicton. At the same time, ice advanced from the northeast to deposit the Valders diamicton (blue arrows on right). Black arrows are drumlins formed during the advance that deposited the older Holy Hill Formation. From the U.S. Geological Survey, 7.5-minute, Potter and Valders quadrangles.



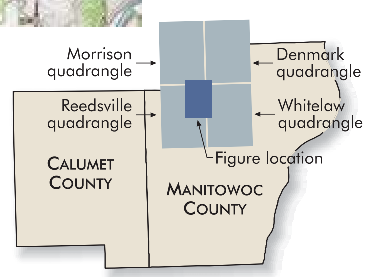


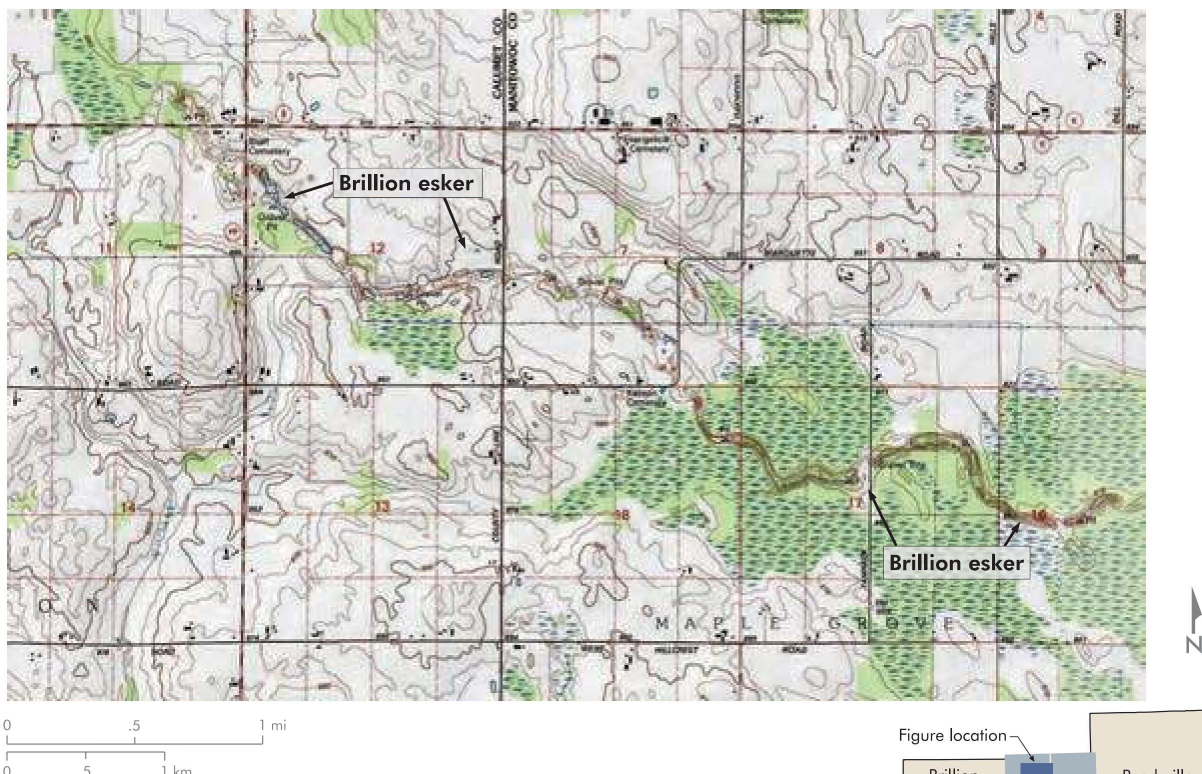
Figure 17. Topographic map of bed of a glacial lake that was dammed when ice was at the Menchalville end moraine. Blue arrows show ice-flow direction and the dashed line marks the outer edge of Valders advance to the Menchalville moraine. From the U.S. Geological Survey, 7.5-minute, Denmark, Morrison, Reedsville, and Whitelaw quadrangles.

streams. Unit *Vgp* has less than 20 percent of original floodplain interrupted by depressions formed by melting ice blocks, *Vgpp* has between 20 and 80 percent of original floodplain surface interrupted by kettles, and *Vgh* has more than 80 percent of the original floodplain surface interrupted by depressions formed by melting ice blocks. *Vgt* is former floodplain surface preserved above the modern floodplain. All units were deposited by **braided streams** flowing away from an ice margin or in valleys running parallel to a glacier margin.

Eskers

Eskers are ridges of sand or sand and gravel deposited by streams flowing through tunnels at or very near the base of the glacier. Thwaites and Bertrand (1957) recognized three eskers in the two-county area related to the Kewaunee Formation advances. They interpreted them all as palimpsest features. A small (less than 10 ft (3 m) high), discontinuous esker is present in the valley of Francis Creek, north and west of the Village of Francis Creek (sec. 15, 16, 20, 21 and 26, T20N, R23E). Much of the gravel has been mined away. While there is no outcrop evidence, it appears that this esker may not be palimpsest. Based on its position in the modern creek valley, it may in fact have formed under Valders ice. Another short esker (about ½ mi (1 km) long) occurs north of Mishicot in sec. 23, T21N, R23E. This may also have been formed during the Valders advance. A few other small eskers were not mentioned by Thwaites and Bertrand (1957). One occurs about ½ mi (1 km) west of Tisch Mills (sec. 1 and 12, T21N, R23E) and in very northern Manitowoc County (sec. 1, T21N, R24E). This appears to be a Valders advance esker as well. Several short gravel ridges that appear to be palimpsest eskers occur near the outer edge of the Valders advance. These are almost certainly Holy Hill Formation eskers, in part overridden by Valders ice (sec. 2, 3, 10, and 13, T18N, R22E).

One of the largest and best-defined eskers in northeastern Wisconsin is the Brillion esker, north of the city of Brillion in Calumet County. It extends about 8 mi (12 km) in a more-or-less east–west direction and in places is over 65 ft (20 m) high (plate 1). Although most of the esker that is easily accessible to roads has been mined away, relatively undisturbed



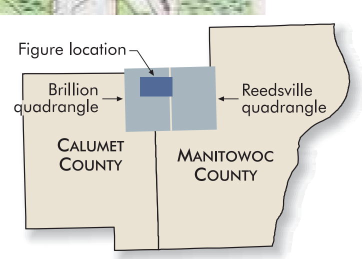


Figure 18. Topographic map of the Brillion esker. Parts of the esker have been mined and are now present as trenches. From the U.S. Geological Survey, 7.5-minute, Brillion and Reedsville quadrangles.

Figure 19. Photo of reddish-brown Chilton diamicton sandwiched between layers of gravel at edge of Brillion esker. If the esker is older than the Chilton readvance, then material above the diamicton must have slumped off the side of the esker, perhaps just after retreat of the Chilton ice.

sections still remain (fig. 18). The esker is underlain by bedrock or sandy diamicton. Outcrops show that it is composed primarily of sand and gravel (predominantly well-rounded dolomite). Because the sedimentology of the esker closely resembles Holy Hill Formation fluvial deposits and interbedded red clayey sediment is found only in the uppermost parts of the esker (fig. 19), it is likely that the esker is palimpsest. If the esker is older than the Valders readvance, then material on top of reddish-brown diamicton must have slumped off the side of the esker, perhaps just after retreat of the Valders ice. Our present interpretation is that the esker formed during a late Wisconsin ice advance that deposited Holy Hill Formation till. The pre–Two Creeks Chilton advance extended into this area and deposited red clayey diamicton over the esker. The esker is located near the margin of the Chilton ice, where typically the red diamicton is a thin drape over existing sandy till landforms.

Suttner (1963) studied the Brillion esker when more of it was still present. He described the ridge as consisting of steeply dipping, discontinuous beds of coarse cobbles and boulders, deposited by gravity, and horizontal beds of coarse and fine sediment deposited by fluvial action and reflecting a fluctuating supply of water. Suttner noted that a portion of the esker formed within the glacier because the base of the esker which is on bedrock rises from an elevation of 835 ft (255 m) to 860 ft (262 m) at a location about 2 mi (3.2 km) downstream. He also concluded that part of the esker formed on the land surface in the open air.

(continues on page 24)

2 FOCUS BOX Manitowoc River and glacial Lake Oshkosh

Note: *This discussion uses present-day elevations, not elevations at the time the glacier edge was in this area.*

The large broad flat valley in which the South Branch Manitowoc River in southwestern Calumet County now flows may be the remnant of a preglacial river channel. At half a mile (1 km) wide, the valley is much larger than would have been cut by the present stream (fig. 20). The channel may have been a north–south trending tributary to the preglacial Sheboygan River, which flowed off the Niagara Escarpment eastward to Lake Michigan (fig. 6). The South Branch Manitowoc River now heads in northern Fond du Lac County and flows northward through western Calumet County, and then northeastward to its confluence with the North Branch Manitowoc River near the Calumet-Manitowoc County border. Most of the north–south trending reach of the river is now at an elevation of about 900 ft (275 m) with a broad wetland present adjacent to the river. The river flows with a low gradient to the northeast to the city of Chilton where it is dammed at an elevation of about 880 ft (268 m).

The South Branch Manitowoc River received meltwater during the Chilton ice advance and a lake was likely present in its river valley and adjacent areas until retreat of the ice in the Fox River lowland allowed drainage of water through the end moraine to the west and into glacial Lake Oshkosh in the Lake Winnebago basin. The morainal deposits on the escarpment are

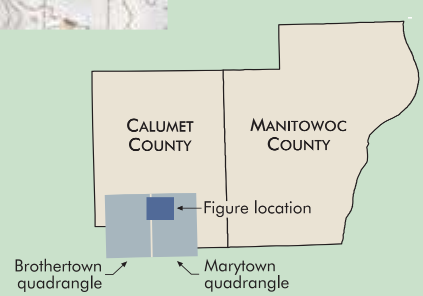
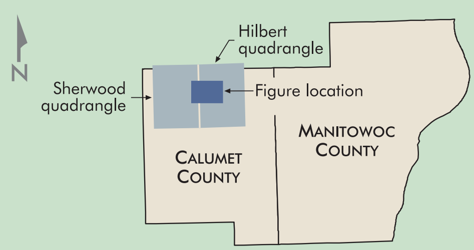


Figure 20. The broad bottom of the Manitowoc River valley, an underfit stream (parts of Brothertown and Marytown U.S. Geological Survey 7.5-minute quadrangles). The area southeast of the river valley has dolomite bedrock very close to the surface. Arrows mark only some of the Green Bay Lobe drumlins formed during the Holy Hill advance and show ice flow direction.

2 FOCUS BOX Manitowoc River and glacial Lake Oshkosh, *continued*

dissected, possibly by meltwater and the lake water draining from the proglacial lake. Flow was off the escarpment and into glacial Lake Oshkosh. Pipe Creek, near the village of Pipe in northern Fond du Lac County, now occupies this channel. Drainage off the escarpment as the ice retreated likely occurred during earlier glacial events also. For example, the terrace composed of predominantly Holy Hill Formation sand and gravel at Stockbridge probably formed when the Holy Hill ice margin retreated to the north and west, off the escarpment, and meltwater and ice-dammed lake water drained into glacial Lake Oshkosh in the Lake Winnebago basin. Supporting this hypothesis, steep gravel beds dip to the west and were formed as debris tumbled down the front slope of a delta. Beds that formed on top of a delta, where sediment was laid down in horizontal layers, are found at an elevation of about 830 ft (253 m) and are overlain by reddish-brown diamicton. Thwaites and Bertrand (1957) describe similar terrace deposits along the Niagara Escarpment in Brown County from DePere south to Greenleaf. There the top of the sand and gravel in the terrace is at an elevation of about 826 ft (252 m). **Cross-bedding** dips to the south and southwest. Reddish clayey till overlies the fluvial sediment. Thwaites and Bertrand (1957) suggest that the sand and gravel was deposited between the rock cliff and the retreating Holy Hill ice front. The outlet was to early Lake Oshkosh, which at that time drained to the Wisconsin River. The North Branch Manitowoc River in northeastern Calumet County (fig. 21) is near an

Figure 21. Topographic map (parts of Hilbert and Sherwood 7.5-minute quadrangles) showing the three shallow channels (arrows) through which water from glacial Lake Oshkosh flowed into the Manitowoc River outlet.



2 FOCUS BOX Manitowoc River and glacial Lake Oshkosh, *continued*

area where the bedrock surface is only at an elevation of about 500 ft (152 m). This low area in the bedrock surface may be where a preglacial river (a tributary to the ancestral Fox River) cut back into the escarpment and flowed northwest to the Fox River. This channel with flow to the northwest could have been active prior to and during the red till glaciations as more than 200 ft (60 m) of clayey red sediment fills in the channel (plate 2, D–D′). More details of the deglaciation history of this area are presented in the glacial history section.

Glacial Lake Oshkosh formed again during the Chilton advance and the Manitowoc River was once again an outlet for the lake. However, if the Chilton ice margin retreated north of the escarpment at Sherwood to open the Manitowoc River basin before the Valders ice margin retreated to the east of Cato Falls, lower in the Manitowoc River valley, drainage for a short time would have been to the west, into Lake Oshkosh. North of Sherwood the Niagara Escarpment is eroded back to the east and the uppermost bedrock is Maquoketa shale (fig. 2). Well logs indicate that north of Sherwood the bedrock is at an elevation of about only 750 ft (229 m) (plate 2, D–D′). Water from east of the escarpment could have drained to the west through this low area, and into the Lake Oshkosh basin and then to the southwest via the Fox River to the Wisconsin River near Portage (elevation 790 ft (241 m)). Evidence for these pre-Glenmore events have mostly been erased by the Glenmore advance.

As Glenmore ice began to retreat from its maximum position, the area of what is now Calumet County was likely isostatically depressed (because of the weight of the ice) about 160 ft (50 m) relative to the Portage area (Hooyer and Mode, 2008) so it seems likely that water drained from glacial Lake Oshkosh eastward down the Manitowoc River as soon as that path was ice free. These outlets are shown in figure 21.

Lake Michigan Lobe moraines south and west of the Twin Rivers lowland

Near the western extent of the Valders ice in southern Manitowoc County, the glacier moved over deposits of the Holy Hill Formation, and the red-brown diamicton is only a few meters thick except in a small end moraine at its outer limit (plate 1). Closer to Lake Michigan, however, the ice advanced over clayey diamicton and lake sediment of the Ozaukee Member (fig. 11, area 6). The moraine of the Ozaukee advance was completely covered by the Valders advance. Thus, from about the location of Highway 42 eastward toward Lake Michigan there is a thick sequence of alternating clayey diamicton and lake sediment, and the bedrock surface is generally more than 100 ft (30 m) below the surface (plate 2). The Valders advance appears to have done little to modify this landscape.

What appears to be a broad moraine extending north from the south county line between Highway 42 and Interstate 43 is mapped as *Vdh*, because we interpret this to be a buried moraine of the Ozaukee advance and not a moraine of the Valders advance.

The Ozaukee and Valders diamicton units are difficult to distinguish in many drill holes. However, based on the presence of sand and gravel beneath relatively thin Valders diamicton in exposures north of Cleveland and along the Lake Michigan shoreline, we interpret most of the thick, clayey diamicton in drill holes in this area to be Ozaukee diamicton. Thus, all of this landscape west and south of the Twin Rivers lowland we interpret to be palimpsest.

Manitowoc River–Twin Rivers lowland and Point Beach

At present, water from Lake Michigan passes through Lake Huron and into the St. Mary's River at Port Huron, Michigan. Whenever glacier ice blocked the north end of Lake Michigan, water level rose within the basin and flowed to the Mississippi through an outlet at the south end of the lake (see focus box 3, "Former Levels of Lake Michigan"). Beaches recording these higher lake levels are present in Manitowoc County inland from the present shoreline. Map unit *Vslp* (plate 1) and area 7 in figure 11 show the area covered by the lake at its maximum extent in eastern Manitowoc County. The Glenwood shoreline had little beach development, but much of the ground surface throughout the area has a thin cap of wave-washed sand and fine gravel. In some places, sand may be thick where streams entered the lake or where currents allowed the buildup of sand. There was relatively little deposition at the mouth of the Manitowoc River when water was at the Glenwood level during retreat of the Valders ice. A wave-washed diamicton surface extends through the city of Manitowoc, beneath the airport, and northward just east of Rockwood (plate 1). Most of this fairly flat surface at about 650 ft (198 m) appears to be diamicton and not a **delta** fanning out into the lake.

Glenwood-level features are only present as far north as the Manitowoc and Twin Rivers lowland, indicating that they formed before the Two Rivers advance (Evenson, 1973; Mickelson and Evenson, 1975). In a gravel pit on the north side of Two Rivers, at the front of the Two Rivers moraine, diamicton of the Two Rivers advance overlies nearshore sand at an elevation of 620 ft (189 m) (NW,

3 FOCUS BOX Former levels of Lake Michigan

It has been known since the late 1800s that water levels in the Great Lakes have been higher in the past than they are now, leaving behind a series of beaches farther inland. Goldthwait (1906, 1907) mapped the extent of these beaches along the Lake Michigan shoreline.

Numerous papers have been published that attempt to define the age and significance of these beaches, and there is still disagreement about the details of the history of lake level fluctuations (Clark and others, 1990). A major review was done by Hough (1958), but there are now many more radiocarbon dates that document the story. More recent overviews of Lake Michigan history since deglaciation have been covered by Hansel and others (1985), Hansel and Mickelson (1988), Colman and others (1994, 1995), Larsen (1994), and Larson and Schaetzl (2001).

We see beaches well above modern lake level for several reasons: isostatic depression and subsequent rebound, higher outlets, and more water being supplied to the lake from precipitation and glacial meltwater than today.

Glacio-isostatic depression and rebound

The sheer weight or load of the ice sheet compresses the earth's crust in a process known as glacio-**isostatic depression**. Given sufficient time, the ice sheet can depress the land surface a distance equal to about 30 percent of the ice thickness. Thus, during and shortly after the last glaciation, the earth's surface in what is now Manitowoc and Calumet Counties was substantially lower in elevation—by perhaps more than 350 ft (107 m)—than it is today.

As the ice thins and retreats, the crust lifts back up in a process known as **isostatic rebound**. The crust continues to rise at an ever-slower rate for thousands of years after the ice is gone, until the earth's surface returns to about the elevation it had before glaciation. Rebound explains why beaches that formed when the ice sheet was just retreating have higher elevations now than they were when they first formed. In Door County, shorelines have been uplifted as much as 75 ft (25 m) relative to beaches of the same age in southern Manitowoc County (Thwaites and Bertrand, 1957). The land at the north end of what is now Lake Michigan was depressed even more because the ice was thicker there. Clark and others (2008) provide recent analysis of postglacial uplift in eastern Wisconsin.

Changing outlets

The outlet at the south end of Lake Huron is considered to be the "controlling outlet" for Lake Michigan today, and it has been for about the last 5,000 years. In the last 150 years, water level in Lake Michigan has fluctuated a little over 6 ft (2 m) because of changes in precipitation and evaporation. Over the last 4,000 years, fluctuation has been 6 to 13 ft (2 to 4 m) (Thompson and others, 1991).

Whenever the northern end of Lake Michigan was covered by glacier ice, the Lake Huron outlet was blocked and the lake rose to drain down the Illinois River at Chicago (Hansel and Mickelson, 1988; Schneider and Hansel, 1990). Three

3 FOCUS BOX Former levels of Lake Michigan, *continued*

higher levels of Lake Michigan are recognized: Glenwood at 640 ft (195 m), Calumet at 620 ft (189 m) and Toleston at 605 ft (185 m). These elevations are measured at the Chicago outlet, and minor differential rebound causes the shorelines in Manitowoc County to be about 10 ft (3 m) higher. Present mean lake level is about 580 ft (176 m). The same Chicago outlet was used when the lake was at the successively lower Glenwood, Calumet, and Toleston levels. The reason for this is disputed. The traditional explanation is that downcutting of the outlet took place either at the end of the Valders advance (Bretz, 1951) or at the beginning of the Two Rivers advance (Kehew, 1993) and again after the Two Rivers advance. An alternative hypothesis proposed by Hansel and Mickelson (1988) is that the same outlet was occupied each time with little downcutting, but large differences in the volume of water coming into the Lake Michigan basin caused the differences in lake level.

After deglaciation

Whenever glacier ice retreated north of the northern end of Lake Michigan, water level dropped. Low, isostatically depressed outlets that progressively opened as ice retreated carried water into the eastern Great Lakes or across southern Ontario to the St. Lawrence River (Hough, 1958; Hansel and others, 1985). At this time, Lake Algonquin developed, and what are now Lakes Michigan, Huron, and Superior were at the same level (Hough, 1958). There is no evidence for this lake level in Manitowoc County and beaches of this age may be below present lake level (Larsen, 1987) in Manitowoc County or they may coincide with the Toleston level and cannot be identified for that reason. Because the land under these northern outlets was much depressed, the level of Lake Michigan fell to levels more than 300 ft (100 m) below its present level (Colman and others, 1995). This event is called the Chippewa lowstand. A period lasting several thousand years followed where the water level was very low, and streams in this area must have downcut substantially below present lake level in their lower reaches. This would have produced two lakes in the present Lake Michigan basin with a river connecting them.

Lake level rises again

As northern outlets rose due to isostatic rebound, the level of Lake Michigan rose as well. By about 5,000 years ago the lake had risen to the Toleston level at 605 ft (184.5 m) and water again drained through the Chicago outlet. This highstand of the lake is called the Nipissing phase of Lake Michigan, and beaches and dunes from it abound along the Manitowoc County shoreline. Because the Nipissing phase lasted longer and is more recent, beach features are better developed than landforms at the Glenwood or Calumet levels.

sec. 31, T20N, R25E). The sand there contains Two Creeks–age wood (Schneider, 1990). Many of the sand deposits in the Two Rivers lowland were deposited into the Calumet level of Lake Michigan, and Two Creeks–age wood has been found in several excavations including one near the Roncalli High School in the city of Manitowoc (SW, sec. 9, T19N, R24E) and in lake sediment along the low eroding bluff on the Lake Michigan shore (NE, sec. 16, T19N, R24E). Note that neither of these sites has till above Two Creeks wood. As they are outside the Two Rivers moraine, it would not be expected. After retreat of ice from the Two Rivers moraine, the level of Lake Michigan remained at the Calumet level until the north end of Lake Michigan was deglaciated, allowing the level to drop substantially.

Rivers flowing into Lake Michigan were deeply downcut when the lake level dropped, then slowly filled their valleys whenever the level rose again. Often the downcutting events left terraces, which are abandoned floodplains, or wave-cut bluffs in the larger river valleys. The highest terrace mapped in figure 22, unit *T1*, appears to have been the floodplain occupied by the stream when Lake Michigan was at the Glenwood level. Unit *T2* formed when the lake was at the Calumet level, and unit *T3* formed when the lake was at the Toleston level. Unit *T3* is a mid-Holocene (Nipissing) shoreline and associated stream terrace at an elevation of about 620 ft (189 m). Unit *Vst* indicates terraces above modern streams where individual terraces cannot be associated with a particular lake level or where they are too small to map.

The sediment in map unit *Vspp* (Valders sand and some gravel in outwash and lacustrine plains) is moderately well-sorted, well-stratified sand and sandy gravel deposited by glacial

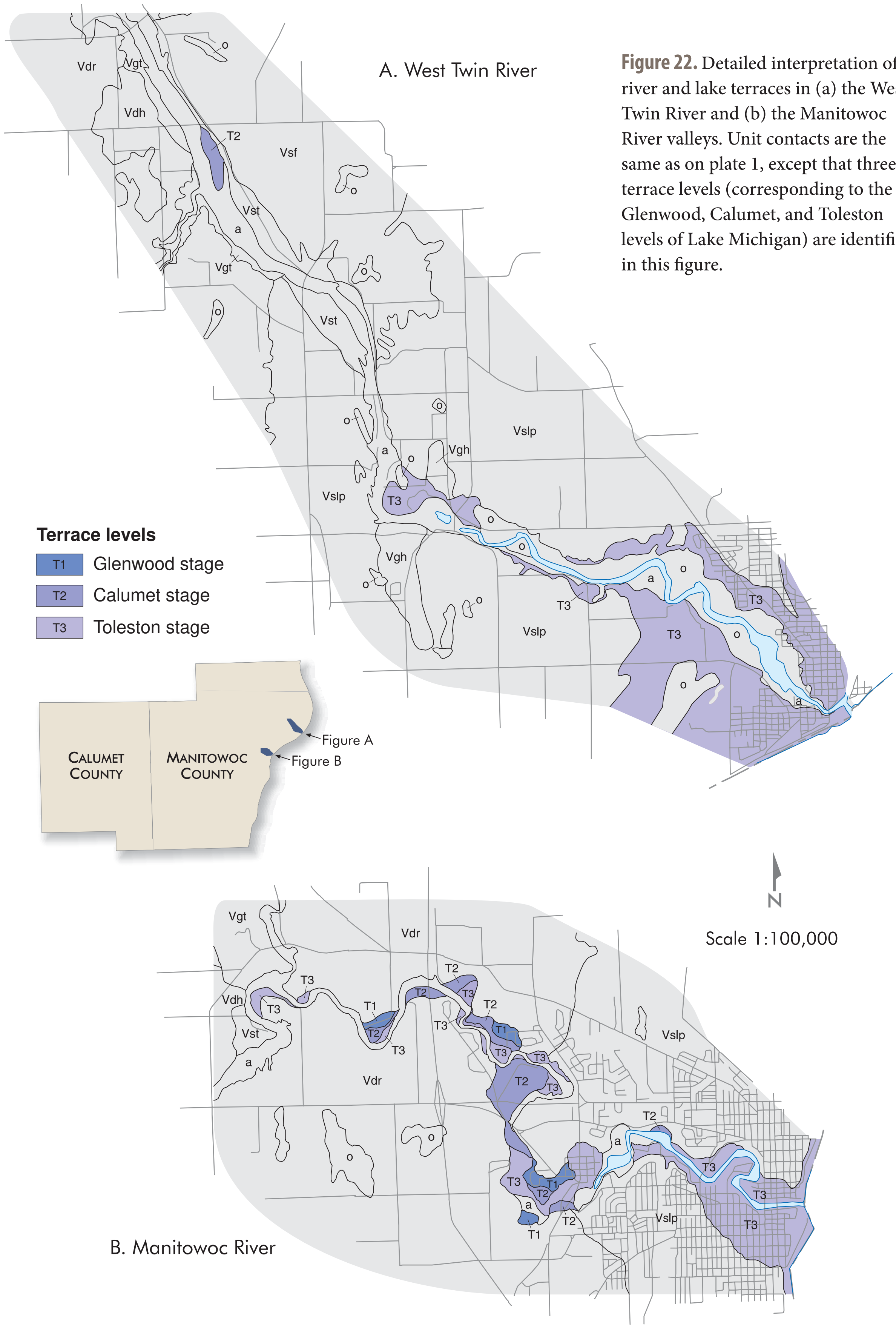
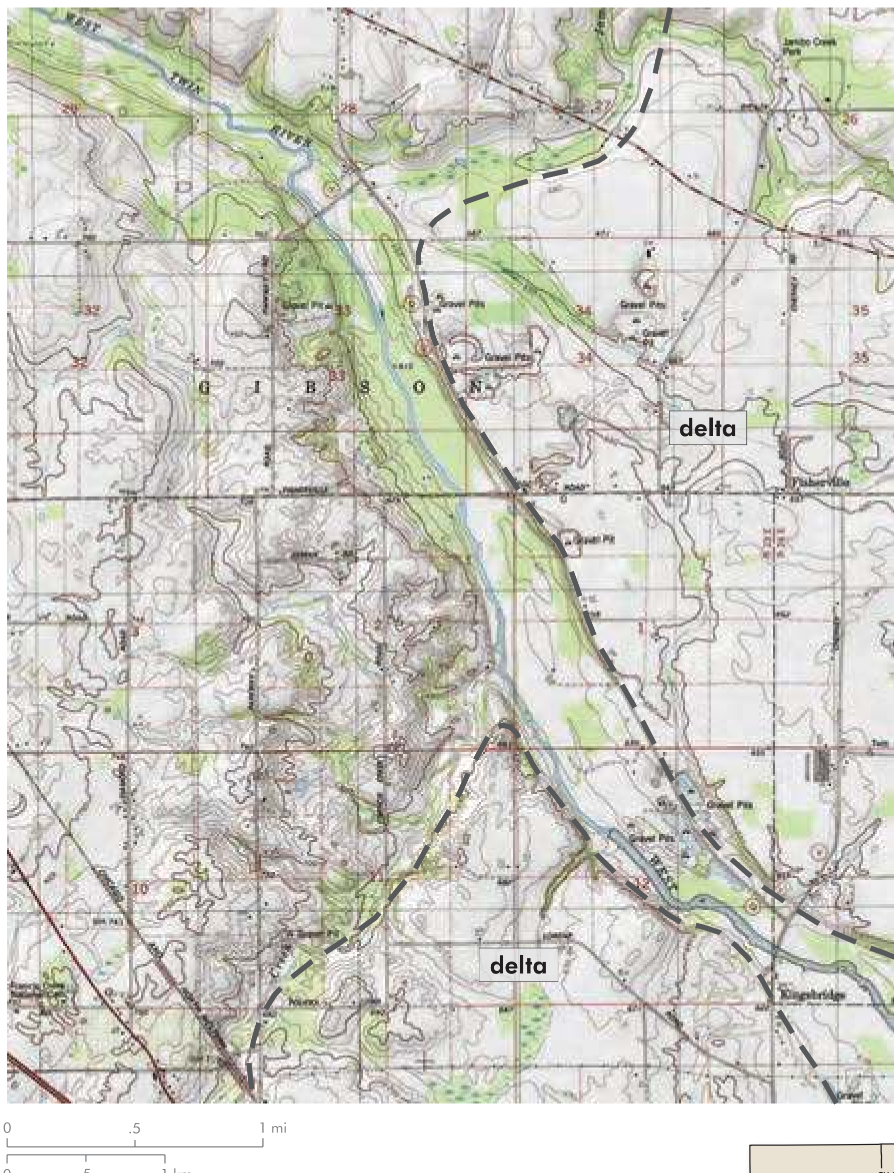


Figure 22. Detailed interpretation of river and lake terraces in (a) the West Twin River and (b) the Manitowoc River valleys. Unit contacts are the same as on plate 1, except that three terrace levels (corresponding to the Glenwood, Calumet, and Toleston levels of Lake Michigan) are identified in this figure.

streams or by waves and currents in glacial lakes. This unit has between 20 and 80 percent of original floodplain interrupted by depressions formed by melting ice blocks.

Unit *Vsf* (plate 1) was deposited in a fan or delta surface sloping gently away from the former ice margin. The area west of Mishicot in northeastern Manitowoc County (sec. 1, T20N, R23E and adjacent sections) is a delta deposited at the mouth of the West Twin River when Lake Michigan was at the Glenwood level (fig. 23). The lowland of the East and West Twin Rivers was an embayment of Lake Michigan when the lake was at the Glenwood level and again several times later. Thwaites and Bertrand (1957) thought that the lowland was part of a separate lake, called Lake Shoto, because of the belief that there was only one "red till" advance and that, therefore, till on both sides of the lowland were of the same age. We now believe that the lowland was connected to Lake Michigan as Valders ice retreated and probably as Valders ice and Two Rivers ice advanced as well (Evenson, 1973; Mickelson and Evenson, 1975). Beach deposits at the Glenwood level occur in several places in the basin. For



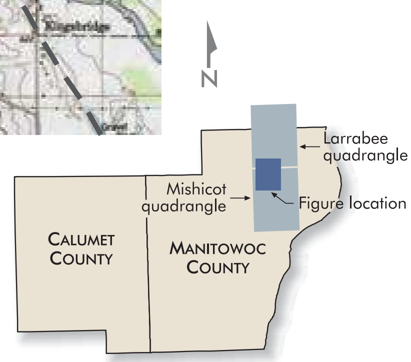


Figure 23. Topographic map of delta at the mouth of the West Twin River built during the Glenwood stage of Lake Michigan. From the U.S. Geological Survey, 7.5-minute Larrabee and Mishicot quadrangles.

example, a gravel deposit about ½ mi (1 km) south of Shoto (SW, sec. 29, T20N, R23E) on the West Twin River contains very well-sorted, openwork gravel typical of beach deposits at an elevation of 640 to 650 ft (196 to 198 m) above sea level.

Fischer Creek Conservation Area on the shoreline between Cleveland and Northeim (sec. 14, T17N, R23E) has good examples of dunes derived from deposits of higher lake levels (fig. 24). The surface was covered with shallow water at the Glenwood and Calumet levels, and some dune formation may have begun as water subsided from that surface. When lake level dropped, wind reworked the sand into dunes, and there was probably continuing movement of the sand whenever the vegetation on the surface was disturbed. The dunes have a northwest–southeast orientation, presumably from winds blowing from the southeast because there appears to be no large source of sand to the northwest. Several weakly developed soils are buried in the dune sand indicating several periods of stability alternating with dune reactivation. No samples have been found for radiocarbon dating, but it seems likely that dune formation was active as water rose to the Toleston level during the Nipissing phase. The shoreline at Fischer Creek Conservation Area has an eroding bluff. This shore has likely been eroding since lake level rose to its present level about 5,500 radiocarbon years (6,200 calendar years) ago. Thus, there is no way of knowing for sure what deposits have been removed by erosion. The shoreline during the Nipissing phase may have been about a mile (1.6 kilometers) farther east at that time, but all evidence has been eroded.

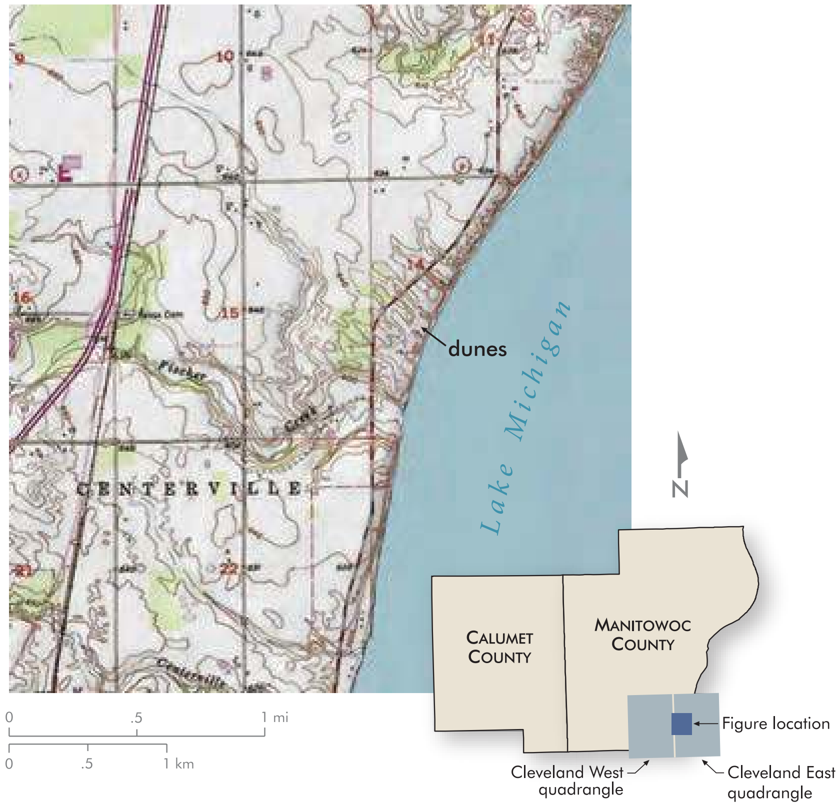


Figure 24. Topographic map of dunes at Fischer Creek Conservation Area (parts of the U.S. Geological Survey, 7.5-minute Clevelend East and Cleveland West quadrangles).

In and near the Two Rivers lowland, deposition has been more continuous. Dott and Mickelson (1995) have studied in detail the preserved record of lake level rise to the Nipissing highstand and the formation of beach ridges during subsequent lake level drop. Parallel ridges and intervening swales can be found in Point Beach State Forest (fig. 25). The road into the park crosses these ridges and swales. Figure 26 shows a cross section from the location of the shoreline at the Nipissing highstand eastward to the present beach at the Rawley Point Lighthouse.

Radiocarbon dates of 5,970±80 and 5,740±120 years ago (6,680 and 6,536 calendar years, respectively) on wood in lake sediment above Two Rivers diamicton record the rise of the lake from the Chippewa lowstand to the Nipissing highstand (see focus box 3, "Former Levels of Lake Michigan"). At that time a spit, a point of land made up of wave-deposited sand, built out into the lake and organic sediment began to collect in the lagoon behind it (now Molash Swamp). Sand dunes later formed on the spit, further raising its elevation and protecting the lagoon behind. By 4,750±90 radiocarbon years ago (5,350 calendar years) peat was accumulating in the swamp (fig. 26). By this time the active shoreline was on the east side of the spit. As lake level fell from the Nipissing highstand, a series of beaches formed as the shoreline advanced lakeward. During low lake levels, sand was blown up onto the older, higher beaches that were farther inland,

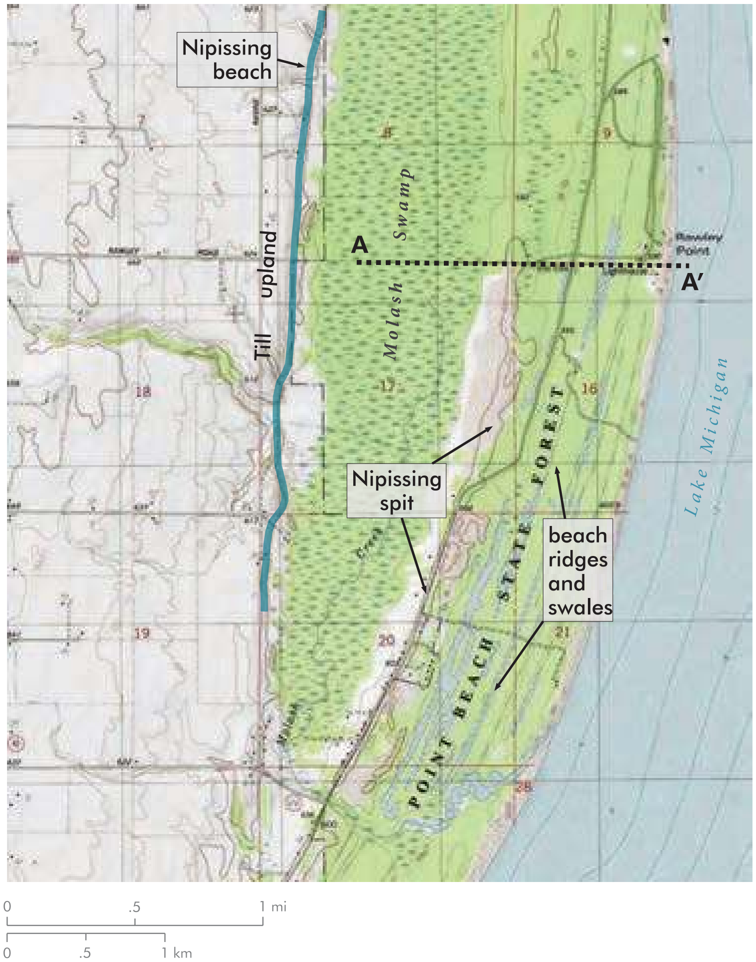
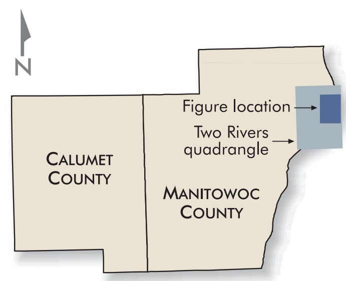


Figure 25. Topographic map of Point Beach State Park (part of Two Rivers 7.5-minute quadrangle). The eastern edge of till upland (blue line) marks the shoreline during the Nipissing highstand of the lake. At that time the spit was built by migrating sand and the Molash Swamp was a lagoon that slowly filled with organic deposits. Dotted line (A–A') shows location of cross section in figure 26.



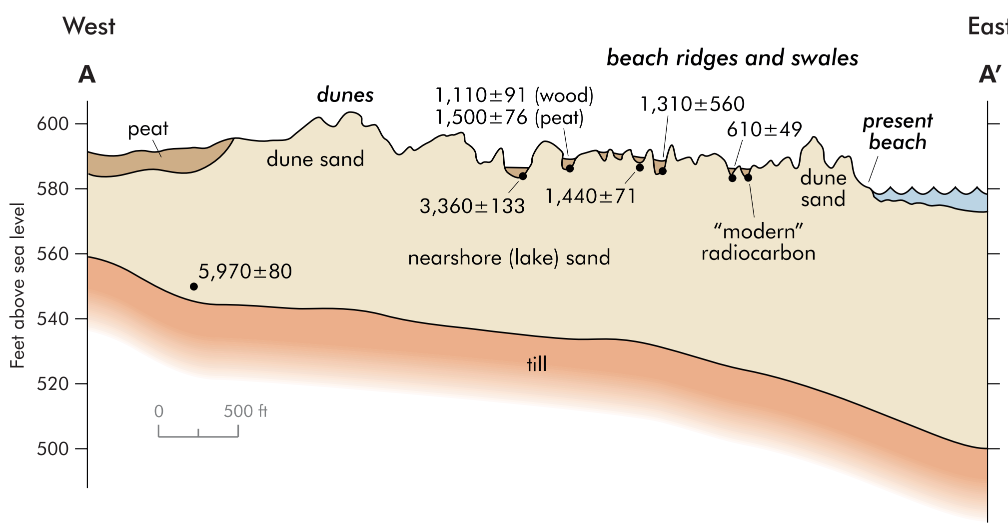


Figure 26. East–west cross section from Molash Swamp to Lake Michigan shore at Point Beach State Park. Location of cross section shown in figure 25. Radiocarbon dates are in calendar years. (From Dott and Mickelson, 1995.)

creating dunes that covered up the beach ridges. Subsequent water level rise eroded away some of these beach ridges, but some are preserved.

Radiocarbon dates from narrow lagoons between the beach ridges (figs. 26 and 27) illustrate the eastward migration of the shoreline. The radiocarbon ages and the elevation of the base of the beach ridges (Dott and Mickelson, 1995) also illustrate the fluctuation in lake levels due to precipitation and evaporation changes superimposed on a progressive drop due to downcutting of the controlling outlet at Port Huron.

A similar set of beach ridges is present west of Two Rivers at the Woodland Dunes Nature Preserve (fig. 28). The oldest dates on the Nipissing rise there are 6,100±90 radiocarbon years (7,500 calendar years) ago on driftwood and 5,640±90 radiocarbon years (6,405 calendar years) ago on peat (fig. 29).

Two Rivers moraine

The advance of the Lake Michigan Lobe following growth of the Two Creeks forest about 12,000 radiocarbon (14,000 calendar) years ago ended at a well-developed end moraine (fig. 11, area 8). This end moraine is located beneath the water tower in the city of Two Rivers. The moraine extends northward into Kewaunee County (Clayton, 2013) where it meets the similar age Denmark moraine of the Green Bay Lobe. The ridge, mapped as *Tdh* and *Tdhe* on plate 1, appears to contain true end moraine segments that mark ice-margin positions. Diamicton is substantially thicker here than in areas to the west. Low-relief hummocky topography, shown in figure 30, is fairly common in clayey till in low broad-crested moraines around the central and southern part of Lake Michigan. It was produced by the slow meltout of glacier ice that had a thin cover of clay-rich sediment on it. This **supraglacial sediment** accumulated as debris-rich ice melted out.

The hummocks are aligned, probably because the glacier was riding up on slow-moving ice near the margin as it slowed down where it came up out of the lake basin. As the glacier advanced, sediment sheared slightly upward into the front edge of the ice along preferred paths (shear planes). This, in turn, resulted in debris accumulating in bands on the ice surface. The bands had varying thicknesses and were roughly parallel to the ice margin. The slight differences in thickness were retained when the overlying ice melted, depositing the debris and creating a hummocky (hilly) landscape. The irregular ridges are around 3 ft (1 m) high and are separated by shallow depressions (fig. 30). The topography here has much lower relief than the high-relief hummocky topography of the Kettle Moraine discussed earlier.

Figure 27. Photo of swale (center) and beach ridges (left and right) taken from entrance road to Point Beach State Park.

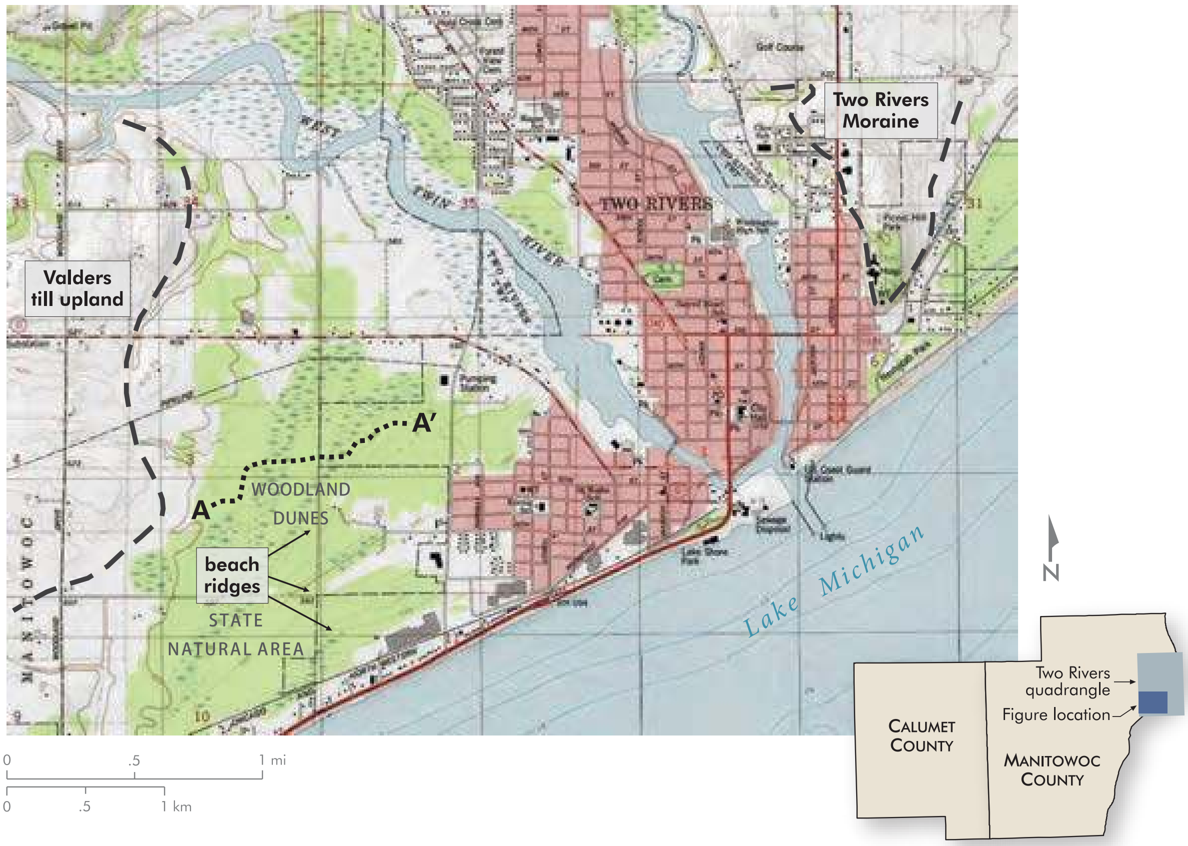


Figure 28. Topographic map of Nipissing stage dunes and beach ridges west of Two Rivers. Southern end of Two Rivers moraine is shown within dashed line in upper right. Area east of till upland is a mix of beach ridges and dunes with wetlands between. Dotted line (A–A') shows location of cross section in figure 29. From the U.S. Geological Survey, 7.5-minute Two Rivers quadrangle.

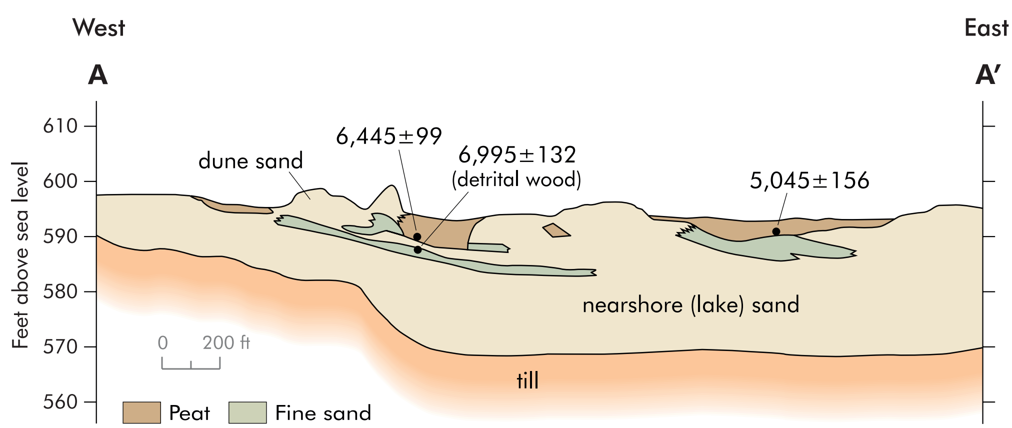


Figure 29. Cross section of Nipissing stage deposits at Woodland Dunes Nature Preserve. Location of cross section shown in figure 28. Radiocarbon dates are in calendar years. (From Dott and Mickelson, 1995.)

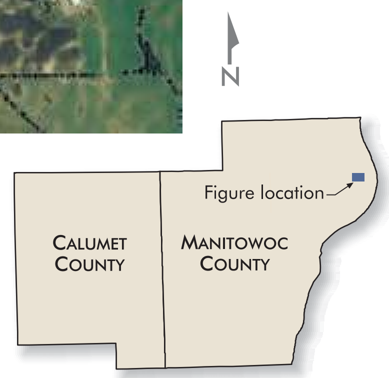


Figure 30. Air photo of low-relief hummocky topography of the Two Rivers moraine. Intersection in left center of photo is Rawley Road and U.S. Highway 42 (T21N, R24E, parts of sec. 7, 12, 13, 18).

Sediments and lithostratigraphy

Sediment deposited by the Green Bay and Lake Michigan Lobes covers all of Calumet and Manitowoc Counties (plate 1). The deposits are classified into lithostratigraphic units called formations, which are further subdivided into members. The members are composed of diamicton (a sediment containing a wide range of grain sizes, from clay to boulders) and commonly interpreted as till (sediment deposited directly by the glacier), sand and gravel (commonly interpreted as outwash deposited by streams flowing away from the glacier), and silty clay (interpreted as lake sediment). The lithostratigraphic units identified in the field area include members of the Hayton, Holy Hill, and Kewaunee Formations (Mickelson and Syverson, 1997; Syverson and others, 2011). Stratigraphic interpretations are based on exposures in the bluffs along Lake Michigan, exposures in quarries and aggregate pits, from boreholes drilled between 1996 and 1998, and from well construction reports kept at the Wisconsin Geological and Natural History Survey.

The following section provides a description of the sediments in the stratigraphic units from oldest to youngest. Members or formations in the area covered by the Green Bay Lobe are described first, followed by a description of the time-equivalent member in the area covered by the Lake Michigan Lobe. The stratigraphy is outlined in table 1. The results of laboratory analysis of diamicton samples from each stratigraphic unit in Calumet and Manitowoc Counties are summarized in table 2 and the grain-size distribution for each unit is shown in figure 31. All of the reported sample results are for diamicton samples interpreted to be till. Other sample results are reported in the Quaternary sample database, TillPro, available from the Wisconsin Geological and Natural History Survey, but are not included in the mean values in table 2.

One important reason for understanding the nature and distribution of glacial deposits is that till units, at least, have relatively constant properties over areas the size of a county. Thus, engineering properties can be estimated (although this cannot substitute for onsite investigation) for the till throughout this area based on past testing. These properties are described in focus box 4, "Engineering Properties of Till Units."

Hayton Formation

There are no radiocarbon dates that determine the time of the glacial advance that deposited the Hayton Formation (Syverson and others, 2011), but it must have been before about 20,000 radiocarbon (24,000 calendar) years ago. Diamicton of the Hayton Formation is present in the subsurface in much of Calumet and Manitowoc Counties. The diamicton generally directly overlies bedrock, is typically extremely dense, and ranges in thickness from a few inches to about 60 ft (20 m) at the type section near Valders in Manitowoc County. The till is light brownish-gray (10YR 6/2) or pale brown (10YR 6/3) in color, using **Munsell color notation**. Till samples have an average magnetic susceptibility of 2.46 x 10^{-3} MKS units (table 2). They are generally gravelly, silty, sandy **loam**, and the 16 samples analyzed have a mean sand:silt:clay ratio of 41:46:13 percent. The silt-size fraction of the matrix contains about 64 percent carbonate. Dolomite is the most common rock type of pebbles and cobbles, but igneous and metamorphic rocks are conspicuous. Using the Unified Soil Classification System (USCS), the till is generally classified as silty sand (SM).

Holy Hill Formation

The Holy Hill Formation was deposited during the late Wisconsin Glaciation, beginning perhaps more than 26,000 radiocarbon (30,000 calendar) years ago and continuing to around 13,000 radiocarbon (16,000 calendar) years ago (Mickelson and Syverson, 1997). The New Berlin Member of the Holy Hill Formation was deposited by ice of the Lake

Dave Mickelson

Table 2. Summary of magnetic susceptibility, grain size, and carbonate content of till in each stratigraphic unit.

	Magn. susc. (x 10^{-3})	Sand (%)	Silt (%)	Clay (%)	Cal-cite (%)	Dolo-mite (%)	Car-bonate (%)
KEWAUNEE FORMATION							
GLENMORE MEMBER							
Calumet							
Mean	1.18	10	41	49	3	27	30
Max	1.70	21	51	59	4	60	62
Min	0.48	4	34	33	2	22	21
Count	15	19	19	19	15	15	15
Manitowoc							
Mean	0.98	15	49	36	4	24	28
Max	1.2	23	53	39	5	27	30
Min	0.7	13	47	28	3	21	26
Count	5	5	5	5	5	5	5
Calumet, Manitowoc							
Mean	1.13	11	43	46	3	27	30
Max	1.70	23	53	59	5	60	62
Min	0.48	4	34	28	2	21	21
Count	20	24	24	24	20	20	20
CHILTON MEMBER							
Calumet							
Mean	1.60	18	46	36	4	46	50
Max	4.98	36	63	61	6	75	80
Min	0.50	3	35	15	2	22	24
Count	51	52	52	52	34	34	34
Manitowoc							
Mean	1.84	18	54	29	—	—	—
Max	2.23	44	73	39	—	—	—
Min	7.19	5	17	21	—	—	—
Count	29	34	30	30	—	—	—
Calumet, Manitowoc							
Mean	1.69	18	49	33	4	46	50
Max	4.98	44	73	61	6	75	80
Min	0.50	3	17	15	2	22	24
Count	80	86	82	82	34	34	34
BRANCH RIVER MEMBER							
Calumet							
Mean	2.31	39	44	17	2	49	51
Max	3.26	46	63	31	2	52	54
Min	1.62	22	24	6	1	43	45
Count	9	9	9	9	3	3	3
Manitowoc							
Mean	2.03	33	49	18	2	30	31
Max	2.90	52	66	26	4	62	65
Min	1.50	18	36	5	0	1	1
Count	14	17	17	17	14	14	14
Calumet, Manitowoc							
Mean	2.14	35	47	18	2	33	35
Max	3.26	52	66	31	4	62	65
Min	1.50	18	24	5	0	1	1
Count	23	26	26	26	17	17	17
KEWAUNEE FORMATION *(cont.)*							
TWO RIVERS MEMBER							
Manitowoc							
Mean	1.18	26	42	33	4	42	46
Max	1.30	40	56	42	8	76	84
Min	0.80	17	38	21	3	27	31
Count	9	10	10	10	9	9	9
VALDERS MEMBER							
Manitowoc							
Mean	1.71	23	48	29	3	39	42
Max	12.22	46	75	60	35	60	64
Min	0.70	6	28	5	0	1	2
Count	125	152	149	149	88	88	88
OZAUKEE MEMBER							
Manitowoc							
Mean	1.35	9	52	40	3	53	56
Max	1.77	17	81	57	7	112	118
Min	1.00	2	34	10	2	34	37
Count	37	40	40	40	6	6	6
HOLY HILL FORMATION							
Manitowoc							
Mean	2.19	38	45	17	3	54	57
Max	3.30	57	63	31	4	78	83
Min	0.60	25	28	6	2	39	42
Count	37	44	44	44	26	26	26
Calumet							
Mean	1.94	48	40	12	3	76	80
Max	4.82	58	47	24	4	83	86
Min	0.78	40	18	3	3	64	68
Count	14	14	14	14	10	10	10
Calumet, Manitowoc							
Mean	2.12	40	44	16	2	46	48
Max	4.82	58	63	31	4	83	86
Min	0.60	25	18	3	0	0	1
Count	51	58	58	58	48	48	48
HAYTON FORMATION							
Calumet							
Mean	2.46	41	46	13	4	60	64
Max	3.38	66	53	20	13	85	89
Min	1.20	32	34	0	3	48	53
Count	16	16	16	16	13	13	13

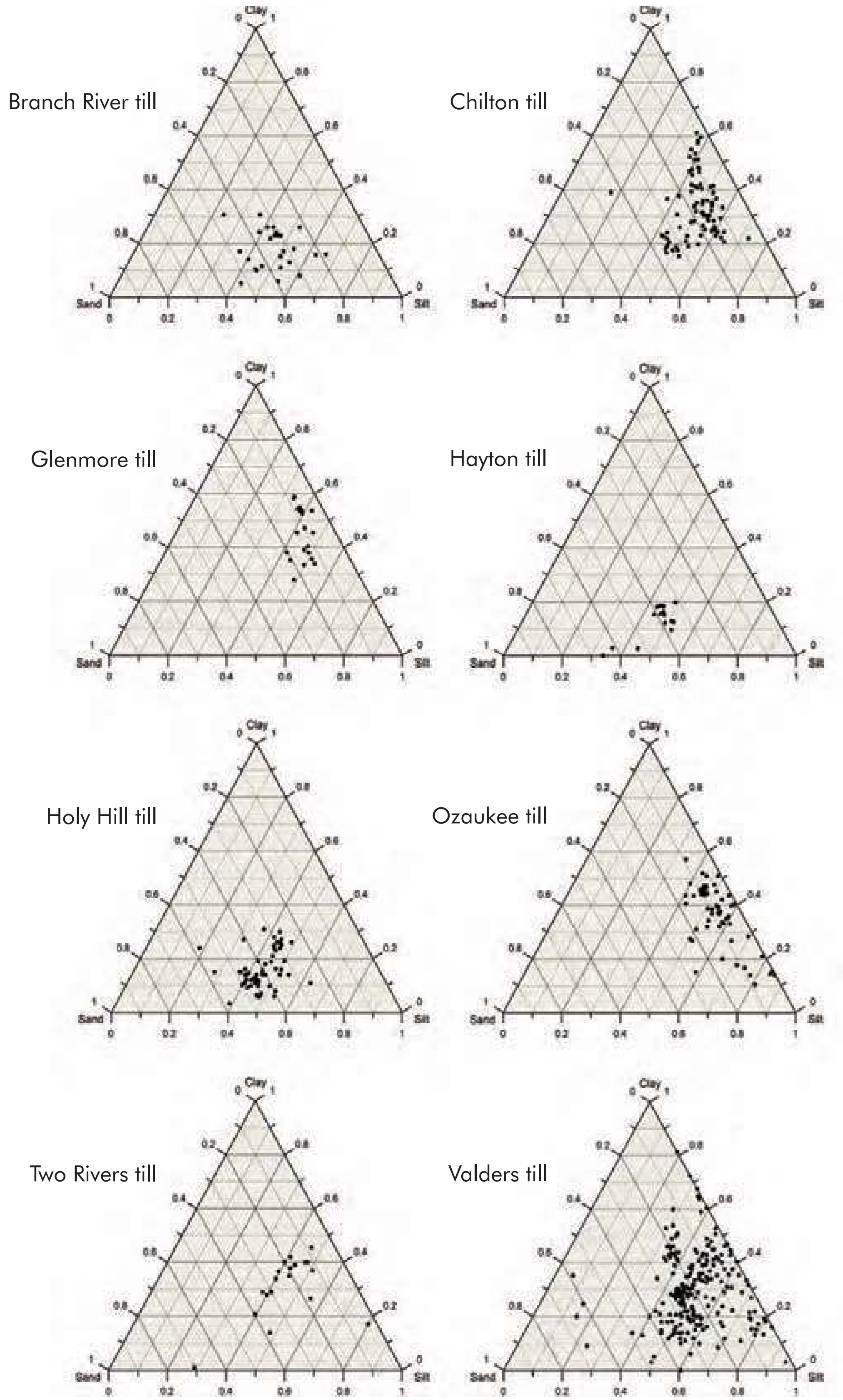


Figure 31. Triangular diagrams showing percent sand, silt, and clay of till in each stratigraphic unit.

4 FOCUS BOX Engineering properties of till units

Although no measurements of engineering properties except for grain-size distribution were made for this study, we here summarize what has been published in several compilations (Rodenbeck and others, 1987; Rodenbeck, 1988; Simpkins and others, 1990; Edil and Mickelson, 1995). A summary of grain-size distribution is presented in the main text along with a discussion of stratigraphic units.

One reason for mapping glacial deposits is to allow engineers, planners, and others a way of predicting properties before actually taking samples at a site. However, the properties given here should not be used as a substitute for site-specific measurements of engineering properties. Only till units are discussed because they tend to be reasonably similar over broad areas or change properties in a predictable manner from place to place. Water-deposited sediment is much more variable and difficult to predict over broad areas because grain-size distribution and other properties depend, to a great extent, on the velocity of water in which they were deposited.

Not all stratigraphic units have been tested, so those are not shown in the tables. In general, Kewaunee Formation till has a hydraulic conductivity that is two to three orders of magnitude lower than the conductivity of Holy Hill Formation till. The exception is when field-measured conductivity tests are used. These often include the effect of fractures, increasing the hydraulic conductivity substantially (Valders Member in table 3). This effect also produces lower hydraulic conductivities in laboratory tests than field tests. Mostly because of the larger amount of sand in the Holy Hill Formation, dry unit weight, hydraulic conductivity, and friction angle are higher than in the Kewaunee Formation. Generally, Holy Hill Formation till has zero cohesion, whereas all of the Kewaunee Formation tills that have been tested have cohesion.

There is a relatively small range in friction angle among the Kewaunee units. They fall between 29 and 31 degrees. Friction angle of the New Berlin Member of the Holy Hill Formation in eastern Wisconsin is about 35 degrees (table 3). Overconsolidation appears higher in some samples of the Ozaukee and Valders tills north of Sheboygan than in the Ozaukee till south of Sheboygan (table 4). One sample of Two Rivers till has undergone even more overconsolidation. There are several possible explanations to account for the difference. One reason may be that, based on independent evidence discussed elsewhere in this report, there was permafrost (tundra vegetation) during deposition of the Ozaukee and Valders tills, but not during deposition of the Two Rivers till. If ice remained in the pore spaces (interstitial ice) until the weight of the ice was removed, as it would in the former case, then overconsolidation due to weight of the ice would be minimal (Acomb and others, 1979) compared to

Table 3. Engineering properties of till units in Manitowoc County. (From Edil and Mickelson, 1995.)

Till units	Hydraulic conductivity (# samples) Laboratory	Hydraulic conductivity (# samples) Field	Dry unit weight (kN/m³)	Water content (%)	Friction angle (degrees)	Cohesion (kPa)
KEWAUNEE FORMATION						
Two Rivers	40×10^{-9} (11)	—	19.0	16	30	11
Valders	400×10^{-9} (19)	3.25×10^{-5} (12)	17.7	17	29	28
Ozaukee N[a]	50×10^{-9} (27)	5.0×10^{-7} (9)	18.6	17	31	24
Ozaukee S	—	1.6×10^{-10} (1)	17.9	18	30	7
HOLY HILL FORMATION						
New Berlin	—	2×10^{-6} (20)	19.9	8	35	0

[a] "Ozaukee N" refers to data from what was formerly called Haven till. (See focus box 5, "Unresolved Stratigraphic Questions.")

Michigan Lobe, and the Horicon Member of the Holy Hill Formation was deposited by ice of the Green Bay Lobe. These are not distinguished on plate 1 because there appears to be no New Berlin Member diamicton at the surface in the two-county area. There may be sand and gravel of the New Berlin Member at the surface. There is New Berlin Member till in the subsurface, but this is shown as undifferentiated Holy Hill diamicton on plate 2 and in table 2.

The Horicon Member is present at the surface in southern Calumet County and southwestern Manitowoc County (plate 1). Because the Holy Hill Formation units are indistinguishable in these two counties, we present data from Holy Hill Formation sediments interpreted to be till from each county (table 2). All of the Holy Hill till in map units *Hdr*, *Hds*, and *Hdh* (rolling, streamlined, and hummocky topography, respectively) is pale brown (10YR 6/3) to light reddish-brown (2.5YR 7/4) in color and belongs to the Horicon Member. Till samples in both counties have a combined average magnetic susceptibility of 2.12 x 10^{-3} MKS units. The till is generally crudely stratified or unstratified, gravelly, clayey, silty sand. Classified using the USCS, the diamicton is generally silty sand (SM). The 14 till samples from Calumet County analyzed for grain-size distribution have a mean sand:silt:clay ratio of 48:40:12, and the 44 samples from Manitowoc County have a mean sand:silt:clay ratio of 38:45:17. The silt-size fraction of the matrix contains about 80 percent carbonate in Calumet County and about 60 percent in Manitowoc County. The reason for this difference isn't clear, because both lobes flowed across carbonate rock for much of their flowlines. 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 diamicton of the Hayton Formation.

We've identified five stream-deposited sediments associated with the Holy Hill Formation. The sediment in these units was deposited by melt-

4 FOCUS BOX Engineering properties of till units, *continued*

samples in which interstitial ice had melted and water could drain. A more-likely explanation is that the high numbers result from dewatering that occurred when the water table was lower following deglaciation. These possibilities are discussed more completely in Edil and Mickelson (1995).

Table 4. Compressibility and preconsolidation values for Kewaunee tills in eastern Wisconsin (not just Calumet and Manitowoc Counties). (From Edil and Mickelson, 1995.)

Till units	Depth of sample (m)	Initial void ratio	Compression index	Effective overburden stress (kPa)	Pre-consolidation stress (kPa)	Over-consolidation ratio
KEWAUNEE FORMATION						
Two Rivers	1.5	0.50	0.20	30	931	31
	4.8	0.49	0.15	109	518	5
Valders	2.4	0.38	0.10	58	518	9
	6.9	0.54	0.20	156	614	4
Ozaukee N[a]	6.5	0.48	0.17	87	835	10
	6.6	0.44	0.12	132	518	4
	8.4	0.49	0.14	182	672	4
	9.2	0.48	0.11	118	413	4
Ozaukee S	6.0	0.54	0.20	130	422	4
	12.3	0.54	0.13	156	413	3
	15.3	0.51	0.20	330	634	2
	15.3	0.48	0.17	330	672	2

[a] "Ozaukee N" refers to data from what was formerly called Haven till. (See focus box 5, "Unresolved Stratigraphic Questions.")

water from the Green Bay Lobe and perhaps from the Lake Michigan Lobe. Without independent source information they cannot be distinguished. The glaciofluvial units *Hgh* (hummocky gravel and sand), and *Hsh* (hummocky sand and gravel) were deposited on and beneath glacier ice by meltwater streams near the ice margin. The sediment later collapsed as underlying ice melted to produce hummocky topography. The sediments are poorly to moderately well sorted, and poorly to moderately well stratified. The glaciofluvial unit *Hsp* (sand and gravel in outwash plains) was deposited by braided streams in front of the glacier margin. The sediments are moderately well sorted, and well stratified. Map unit *Hgpp* (gravel and sand in pitted plains) has more than 20 percent collapsed (pitted) surface, yet exhibits some of the uncollapsed former stream bed. Map unit *Hst* (sand and gravel in outwash terraces) is similar to *Hsp* except that the stream surface was abandoned as the stream downcut its bed, forming a terrace that is now above the floodplain of a modern stream. Map unit *Hslp* (sand in lake plains) is moderately well sorted and is found in low places in the landscape, where lakes were once dammed by retreating glaciers.

Kewaunee Formation

The members of the Kewaunee Formation present at the surface in Manitowoc County include the Branch River, Chilton, and Glenmore Members in the area covered by the Green Bay Lobe, and the Valders and Two Rivers Members in the Lake Michigan Lobe. The Ozaukee Member is present in the subsurface only, and its westward extent is not well defined. The Valders Member covers most of the surface of Manitowoc County. More than half of Calumet County has diamicton of the Chilton and Glenmore Members at the surface. All of these members contain reddish-brown to light-brown clay-rich diamicton, most of which is interpreted to be till.

There are still some unanswered questions about the distribution of stratigraphic units. This is particularly true of the Kewaunee Formation units because of the similarity of the members. These issues are discussed in focus box 5, “Unresolved Stratigraphic Questions.”

Ozaukee Member (Lake Michigan Lobe)

Acomb and others (1982) and many compilations (Mickelson and others, 1984; Attig and others, 2011) since have recognized the Haven Member as the unit below the Valders Member in northern Sheboygan and Manitowoc Counties. This has recently been called into question (see focus box 5, “Unresolved Stratigraphic Questions”). Therefore, we use the term Ozaukee for the stratigraphic unit below the Valders Member in this report even though literature for the past 20 years has used the term Haven.

In Manitowoc County, 40 till samples of the Ozaukee Member have a mean sand:silt:clay ratio of 9:52:40 (table 2). The samples have a Munsell color of light brown (7.5YR 6/3). Till samples have an average magnetic susceptibility of 1.35×10^{-3} MKS units. The Ozaukee Member is well exposed in bluffs along Lake Michigan in southern Manitowoc County (plate 2). It is present in the subsurface beneath Valders Member deposits generally east of Highway 42. It is also easily identified in the bluff exposures based on stratigraphic position relative to the thin Valders Member deposits that lie above it (Acomb, 1978; Acomb and others, 1982).

Branch River Member (Green Bay Lobe)

The diamicton included in map units *Bdh* (Branch River diamicton in areas of hummocky topography) and *Bdr* (Branch River diamicton in rolling topography) is primarily clayey silt. In the two counties, 26 samples interpreted to be Branch River till have a mean sand:silt:clay ratio of 35:47:18 and 23 samples have a magnetic susceptibility of 2.14×10^{-3} MKS (table 2). Till samples are most commonly reddish-brown (5YR 5/4). The diamicton in *Bdh* is mostly basal till. The diamicton in *Bdh* and *Bdr* is crudely stratified or unstratified, and is generally compact and uniform. The upper few feet of diamicton in map unit *Bdh* may have sand lenses and other discontinuities. The upper 10 ft (3 m) of both *Bdh* and *Bdr* contain fractures. Unit *Bdhe* is mapped in hummocky end moraine. The sediment in map unit *Bgf* (Branch River gravel and sand in alluvial fans) is moderately well-sorted, well-stratified gravel and sand deposited by braided streams in front of the end moraine. The sediment in map unit *Bslp* (Branch River silt and sand in lacustrine plains) is moderately well-sorted silt, sand, and clay that was deposited in standing water in low areas dammed by the ice margin.

Chilton Member (Green Bay Lobe)

The diamicton included in Chilton map units *Cdr* (diamicton in areas of rolling topography), *Cdh* (diamicton in hummocky areas), *Cdhe* (diamicton in end moraines), and *Cds* (diamicton in streamlined areas) is silty clay or clayey silt. In some places the diamicton contains bedded or dispersed organic material including twigs and mosses. The diamicton is most commonly reddish-brown (5YR 5/4) and is similar in color to the Glenmore diamicton. Samples of the Chilton diamicton from the two counties have a mean magnetic susceptibility of

5 FOCUS BOX Unresolved stratigraphic questions in the Kewaunee Formation

The overlap in properties of the diamicton of different members of the Kewaunee Formation creates problems for developing stratigraphic correlations and explanations of glacial history. All of the stratigraphic units above the Holy Hill Formation in east-central and northeastern Wisconsin were formally defined as members of the Kewaunee Formation in Mickelson and others (1984). These were based on sampling done in the late 1970s and early 1980s and summarized in Acomb (1978), Acomb and others (1982), and McCartney and Mickelson (1982).

Haven Member now called Ozaukee Member

Three members of the Kewaunee Formation were recognized in the area covered by the Lake Michigan Lobe from Milwaukee to Kewaunee County. These distinctions were based on grain-size distribution of samples interpreted to be till collected along the Lake Michigan bluff shoreline. Unfortunately, the bluff exposures are not continuous, and there is a gap of about 20 mi (32 km) between northern Ozaukee County and the city of Sheboygan with no bluff exposures. North of Sheboygan there are two Kewaunee Formation till units exposed in the bluff (Principato, 1999), south of the gap there is only one.

For the 1984 definitions it was decided that rather than apply the name Ozaukee to units north and south of the gap, the lower till north of the city of Sheboygan should be called the Haven till. Since that time, mapping in Sheboygan County (Carlson, 2002; Carlson and others, 2011) appears to demonstrate that the Haven and Ozaukee units are actually the same, and that there is a progressive change in till properties from north to south. Based on that argument, what was called Haven Member in earlier reports and publications is here called Ozaukee Member.

This interpretation is consistent with the interpretation of only two pre–Two Creeks Kewaunee Formation units defined in the area glaciated by the Green Bay Lobe (McCartney and Mickelson, 1982) and the interpretation of Lake Michigan Lobe stratigraphy under Lake Michigan. The two members found under Lake Michigan, Shorewood and Manitowoc, defined by Lineback and others (1972, 1974) presumably correlate with Ozaukee and Valders units found on land.

Differing interpretations of Chilton and Branch River Members

Another question surrounds the interpretation of the Chilton and Branch River Members in very northwestern Manitowoc County and southern Brown County. McCartney and Mickelson (1982) and Need (1985) interpret that Branch River till is present at the surface in sec. 5, 6, 7, and 18 of T21N, R22E south of Pelishek Corners and that it continues into Brown County to the west; Socha and others (1999) interpret that area as Chilton till. The till there is sandier than typical Chilton till, but Socha and others (1999) argue that this is because the ice had crossed more upland and therefore incorporated more sand-size particles. Whether there are two different stratigraphic units or a gradational change in a single unit has not been resolved.

Correlation with western Green Bay Lobe

A compositional change also occurs from the east to west sides of the Green Bay Lobe, although these facies differences are more likely due to local differences in substrate. The Kirby Lake till on the west side of the lobe is correlative with the Chilton till on the east side of the lobe, but is more sandy (36 percent versus 17 percent sand in the matrix), and the Middle Inlet till on the west side of the lobe is correlative with the Glenmore till on the east side of the lobe, but is significantly sandier (61 percent versus 15 percent sand in the matrix) (McCartney and Mickelson, 1982).

The Branch River–Chilton grain-size differences may be also be due to local substrate differences with little or no change in location through time. In the center of the lobe the ice was picking up and carrying fine-grained lake sediment, while at the margins the ice encountered sandy till and bedrock.

In summary, it is unresolved as to whether the grain-size differences are due to variations as a result of source location, changes of the source with time, or if the Branch River and Chilton tills represent distinctly different advances.

1.69 x 10^{-3} MKS units. The diamicton is generally compact and uniform, except in the upper few meters where sand lenses and other discontinuities may occur. Classified using the USCS, the diamicton is generally lean clay (CL), but some samples are fat clay (CH). The 30 samples of this till from Manitowoc County have a mean sand:silt:clay ratio of 18:54:29, and from Calumet County, 52 samples have a mean sand:silt:clay ratio of 18:46:36. The diamicton in map units *Cdr* and *Cds* are generally interpreted as basal till; map unit *Cdh* is mostly till, but commonly has a less-than-1 m thick layer of mudflow deposits in the upper portion of the sediment. The hummocks in *Cdh* and the streamlined forms in *Cds* are inherited from the surface form of older glacial deposits. *Cdp* (diamicton in plains) is mostly a wave-washed till surface. Map unit *Cslp* (clay, silt, and sand in lake plains) occupies low places in the landscape that were dammed by retreating glacier ice.

Valders Member (Lake Michigan Lobe)

The diamicton included in Valders map units *Vdh* (diamicton in areas of hummocky topography), *Vds* (diamicton in areas of palimpsest streamlined topography), and *Vdr* (diamicton in rolling topography) is primarily silt loam to silty clay loam. The 149 till samples, all collected in Manitowoc County, have a mean sand:silt:clay ratio of 23:48:29 percent. The samples are similar in color to the Ozaukee diamicton, most commonly light brown (7.5YR 6/4), but the Valders till is commonly redder in outcrop than the Ozaukee. This redder color is probably due to increased oxidation of the normally light-brown diamicton. The contrast in color is most evident where both the Ozaukee and Valders diamictons are present in bluff exposures, and outwash or lake sediment separates the two diamicton units. Samples of the Valders till have a mean magnetic susceptibility of 1.7x10^{-3} MKS units.

The diamicton in *Vdh* is mostly basal till with commonly a less-than-3 ft (1 m) thick layer of mudflow deposits. The diamicton is crudely stratified or unstratified, and is generally compact and uniform, except in the upper few feet, where sand lenses and other discontinuities are found. Most of the diamicton contains fractures. The landscape where map unit *Vdhe* (hummocky end moraine) is found is in part true end moraine and possibly also partly palimpsest. The landscape of *Vds* and *Vdr* is inherited from the underlying topography. Units *Vdsb* and *Vdrb* indicate shallow bedrock located mostly within 10 ft (3 m) of the surface. There are also sand and gravel deposits included in the Valders Member on plate 1. *Vgp* is outwash gravel in uncollapsed plains, *Vgpp* and *Vspp* are pitted outwash, *Vgh* is hummocky gravel and *Vgt* and *Vst*, and *Vsf* are outwash sand and gravel in terraces and deltas.

The sediment in map unit *Vslp* (Valders postglacial wave-cut terraces and associated river terraces along the shore of Lake Michigan) is sand and gravel that ranges in thickness from less than 1 ft (0.3 m) to about 10 ft (3 m) and overlies older diamicton. The sand and gravel is generally well sorted, and in many places contains imbricated pebbles. The sand is interpreted as water-deposited beach and offshore sediment, and represents higher (Glenwood and Calumet) phases of Lake Michigan. Former ice-marginal lake basins have a layer of sand and silt that is 6.5- to 65-ft (2 to 20 m) thick and often covered by peat.

The age of the Valders Member has been debated because of its importance as a geologic unit related to the Two Creeks Forest Bed. A summary of these discussions is given in focus box 6, "History of Stratigraphic Interpretations of Reddish Clayey Till in Northeastern Wisconsin."

Two Creeks Forest Bed

The Two Creeks Forest Bed was described and named by Goldthwait (1907). It is widespread in east-central Wisconsin, and numerous ages average about 11,850 radiocarbon (13,660 calendar) years. It lies above Ozaukee till and lake sediment, and below lake sediment and the Two Rivers till at the type section. It is described in more detail in focus box 7, "More about the Two Creeks Forest Bed."

Glenmore Member (Green Bay Lobe)

The Glenmore Member is the youngest member of the Kewaunee Formation in the Green Bay Lobe and overlies the Two Creeks Forest Bed. The diamicton included in Glenmore map units *Gdr* (diamicton in areas of rolling topography) and *Gdm* (diamicton in terminal or recessional moraines) is silty clay. Some of the diamicton includes dispersed spruce wood and mosses of the Two Creeks Forest Bed. Spruce wood and mosses are also found at the base of the unit. Till samples are most commonly reddish-brown (5YR 5/4) and are similar in color to the Chilton till. Samples of the Glenmore till from Calumet County have a mean magnetic susceptibility of 1.2 x10^{-3} MKS units, and from Manitowoc County, 1.0 x10^{-3} MKS units. The diamicton is generally compact and uniform, except in the upper few feet where sand lenses and other discontinuities occur. Classified using the USCS, the diamicton is generally lean clay (CL) but some samples are fat clay (CH). The 19 samples from Calumet County analyzed for grain-size distribution have a mean sand:silt:clay ratio of 10:41:49, and the 5 samples from Manitowoc County average 15:49:36.

(continues on page 46)

6 FOCUS BOX

History of stratigraphic interpretations of reddish clayey till in northeastern Wisconsin

Until the early 1970s, the reddish-brown diamicton that covers most of eastern Wisconsin was thought to be younger than the Two Creeks Forest Bed and represent one major late Wisconsin glacial event in the Great Lakes area. The diamicton, most of which is interpreted to be till, was called the Valders till and, at least locally, the time of presumed readvance had substage significance (Valderan Substage) (Willman and Frye, 1970).

The Valders till was named for the reddish-brown clayey till at the Valders Quarry, a site long visited by geologists (Thwaites and Bertrand, 1957). Here Valders till lies on older sandy till of the Holy Hill Formation or on striated bedrock.

Although it was assumed that all of the reddish clayey till was younger than the Two Creeks Forest Bed, there is no known site outside the Denmark and Two Rivers moraines in Manitowoc and Calumet Counties where Two Creeks wood is found overlain by till, nor are pieces of Two Creeks wood found in reddish-brown clayey till. Behind the Denmark and Two Rivers moraines, wood is fairly abundant in and beneath red clayey diamicton. A post–Two Creeks cooling is recognized in pollen diagrams outside the area of the Kewaunee Formation deposits (Schweger, 1966).

Area interpretation

In 1973 Evenson proposed that the red till in eastern Wisconsin and western Michigan was actually deposited by several glacial advances, some of which predate the Two Creeks Forest Bed and others that postdate the forest bed. Evenson's original arguments were based mainly on geomorphology. In particular, he noted that the Glenwood shoreline of Lake Michigan was cut into the Valders till surface south of Two Rivers (including the till at Valders Quarry), but not into red tills north of Two Rivers. Based on radiocarbon age determinations, the Glenwood shoreline is older than the Two Creeks Forest Bed (Eschman and Farrand, 1970; Hansel and Mickelson, 1988). Thus, the Valders till, whose type section is at the Valders Quarry (fig. 4), 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 researchers who had assumed that the till at Valders Quarry was equivalent in age to that above the Two Creeks Forest Bed at the Two Creeks type locality.

Regional evidence

Acomb and others (1982) and McCartney and Mickelson (1982) studied the stratigraphy of the red 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 in spite of the fact that the authors were unable to find any site where the Valders till was directly underlying the Two Creeks Forest Bed. Although the presumed equivalent in the Green Bay Lobe (Chilton Member) is found beneath the forest bed, the Valders Member is missing at the Two Creeks type locality, where the forest bed overlies lake sediment and the older Haven (now Ozaukee) Member. To our knowledge there is still no documented site where till of the Valders Member can actually be shown to underlie the forest bed.

Thus, we now believe that the limits of the Two Rivers and Glenmore Members (the Denmark and Two Rivers moraines) 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). The advance is significantly less extensive and less important than was thought in the 1960s, and we no longer use the term Valderan Substage.

7 FOCUS BOX More about the Two Creeks Forest Bed

The Two Creeks Forest Bed in northeastern Manitowoc County is among the most famous and most studied geological localities in the Great Lakes area. Wood found between layers of glacial till provided the first geologic sample to be dated with the radiocarbon method by Libby in 1949. Until that time, people had little idea of when glaciers retreated from North America. Many samples from the site have since been dated (Broecker and Farrand, 1963; Black and Rubin, 1967–68; Kaiser, 1994) and most of the uncalibrated radiocarbon dates on wood are around 12,050 to 11,850 radiocarbon (about 14,100 to 13,800 calendar) years old. More recently, dating of snails found below the wood layer extends the age of the forest bed back to about 14,500 calendar years (Rech and others, 2011).

Located about 2 mi south of the Village of Two Creeks (sec. 24, T21N, R24E), the Two Creeks Forest Bed was first described by Goldthwait (1907). As far as we know, there has been no exposure of it there in the past 30 years, but it is well exposed along the lake shore bluff in sec. 2, T21N, R24E, just south of the county line (intersection of Highways BB and 42, between Highway 42 and Lake Michigan). This site is considered the type section and is now part of the Ice Age National Scientific Reserve.

Wood of the same age, known as Two Creeks wood, is also present at and just behind the Two Rivers and Denmark moraines near Denmark, Wisconsin, and in the Fox River valley between Green Bay and Appleton (fig. 32). A forest bed in the northern part of the Lower Peninsula of Michigan near Cheboygan, once thought to be older than Two Creeks, is now thought to be of the same age (Larson and others, 1994).

At the type section, the base of the bluff is made up of what we now call Ozaukee diamicton that extends about 6 ft (2 m) above beach level near the north end of the section (fig. 33). This diamicton drops below lake level to the south. This is overlain by clayey silt and fine sand lake sediment that coarsens upward to sand and gravel that we interpret to be shallow water and beach sediment. Snails from just below the forest bed have recently been dated at between about 12,400 and 12,100 radiocarbon (14,500 and 13,900 calendar) years old (Rech and others, 2011). They indicate a mixed tundra-taiga transitional environment. The forest bed lies on this sand and gravel, and consists of about an inch (a few centimeters) of organic material that is mostly black spruce

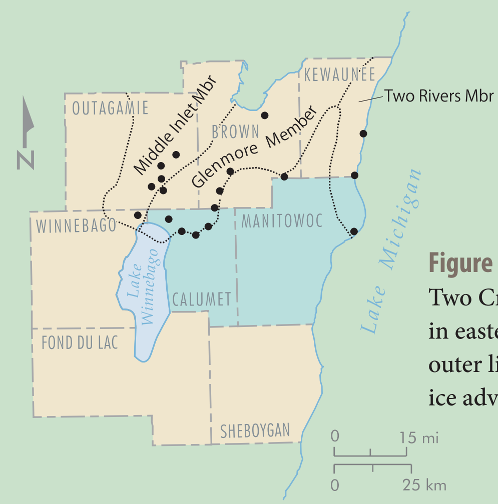


Figure 32. Map showing known Two Creeks Forest Bed localities in eastern Wisconsin and the outer limit of the post-Two Creeks ice advance.

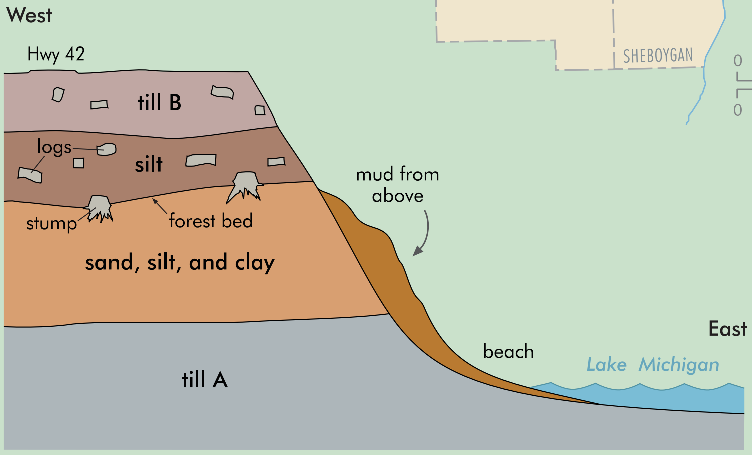


Figure 33. Sketch of exposure at Two Creeks Buried Forest Ice Age Reserve site just south of the Manitowoc-Kewaunee County line at the shore of Lake Michigan.

7 FOCUS BOX More about Two Creeks Forest Bed, *continued*

(*Picea mariana*) needles, twigs, and branches (figs. 34, 35). Very limited amounts of white spruce, balsam fir, tamarack, and aspen have been found (Mode, 1989). This is overlain by silty lake sediment that is 6 to 10 ft (2 to 3 m) thick, which in turn is overlain by Two Rivers diamicton. A thin sand and gravel unit of less than 3 ft (1 m) overlies the diamicton. Throughout the bluff, the layers have been deformed by overriding ice. The diamicton contains numerous pieces of wood picked up as the glacier moved across the forest bed.

Glacial history told by tree rings and trunk orientation

When Valders ice retreated from northeastern Manitowoc County, Lake Michigan was at the Glenwood level, about 60 ft (18 m) above the present level of Lake Michigan. Silt and clay were deposited in the deep water until the glacier retreated far enough north to allow lake level to drop (see focus box 3, "Former Levels of Lake Michigan"). When water level dropped to below present level, a spruce forest began to grow. Based on studies of revegetation in southeast Alaska (Lawrence, 1958), it may have been 50 years or so before spruce trees became the dominant species. The waterlogged, gray 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) estimates that the forest grew for at least 252 years, and that the oldest tree reported has 234 annual growth rings. 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 exposed during construction of Interstate Highway 43 had 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 ice by lake sediment, tree trunks are commonly oriented toward the southwest, indicating the ice flow direction (Black, 1970).

Climate conditions suggested by fossilized snails and beetles

Although the forest bed has relatively poor preservation and little diversity of fossil pollen (Wiese, 1979), the base of the forest bed contains both aquatic and air-breathing snails. The presence of both types of snails indicates a transition from frozen tundra to spruce.

Similarly, fossilized beetle remains from the site have been interpreted by Morgan and Morgan (1979) and by Garry and others (1990) as indicating a more boreal

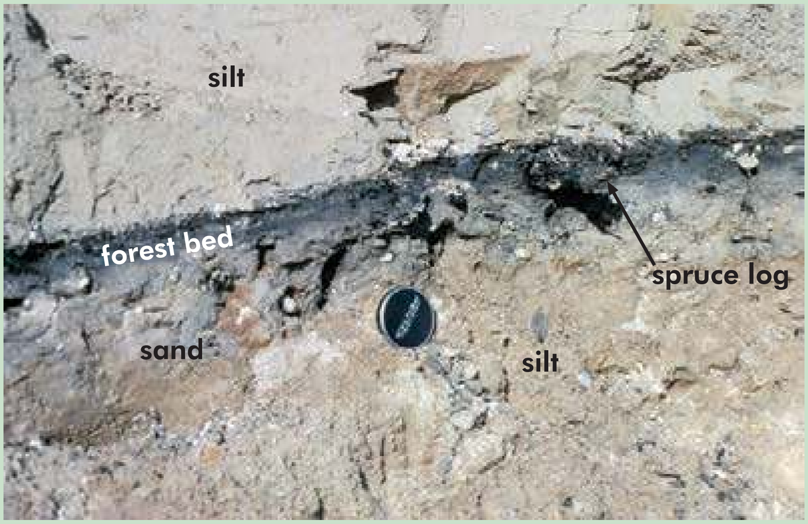


Figure 34. Photo of Two Creeks Forest Bed along Lake Michigan (NE, NE, NE section 2, T21N, R24E). Lens cap is about 2 inches across.

Figure 35. Photo of rooted stump in Two Creeks Forest Bed exposed in bluff along Lake Michigan (NE, NE, NE section 2, T21N, R24E).

7 FOCUS BOX More about Two Creeks Forest Bed, *continued*

climate (long winters and short summers) than 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 site at Valders. Leavitt and others (2006) measured carbon and oxygen isotopes at several Two Creeks–age sites and found them all to represent cool and dry conditions, but did find minor isotopic differences from site to site that they attributed to elevation.

Lake levels and driftwood

Glacier ice eventually advanced into the north end of Lake Michigan basin, damming the northern outlet of the lake 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, about 40 ft above present lake level. Driftwood is abundant in the lake sediment as well as in the Two Rivers diamicton above. Lake Michigan Lobe ice then overrode the site, advancing as far as present-day Two Rivers, and the Green Bay Lobe advanced southward to the Denmark Moraine and in the Fox River basin, to the present location of Appleton (Schneider and Hansel, 1990).

New sites found during project

Two new Two Creeks Forest Bed sites were found during this project in Calumet County. One site is located north of Hilbert (SW1/4 NE1/4 sec. 19, T20N, R19E). Two adjacent boreholes were drilled in the broad, low end moraine that marks the farthest advance of the Glenmore ice into the Brillion basin. Small pieces of spruce wood were encountered in red clayey diamicton at a depth of about 9 to 13 ft (3 to 4 m) in an augered borehole. The horizon containing the wood is overlain by red clayey (Glenmore) till. The wood is dated at 11,690 ±70 radiocarbon (about 13,680 calendar) years old (Beta – 10481). Spruce wood, inter-layered with red clayey diamicton with dispersed organics, and a gleyed (greenish-gray 5GY 5/1) soil horizon, was also encountered in an adjacent borehole from a depth of 30 to 34 ft (9 to 10 m). The wood in this horizon appears to be tree trunks or large branches. The wood appears fresh, yellow in color and has distinct small growth rings. The wood is dated at 12,110±70 radiocarbon (about 14,102 calendar) years old (Beta – 11558). Underlying the Two Creeks horizon is red clayey (Chilton) till to a depth of about 86 ft (26 m). The till contains small twigs and dispersed pieces of mosses, grass, and stems. Although not radiocarbon dated, the organic matter in the Chilton till at this site appears similar to that found at Sherwood, which were dated at 13,370±90 radiocarbon (about 15,460 calendar) years old (Beta – 119360). Below the Chilton till at a depth of 86 to 88 ft (26 to 27 m) is a soil horizon developed in the grayish-brown, silty-sand diamicton that is till of the Cato Falls Member of the Hayton Formation.

A second Two Creeks site was found north of Brillion, in southern Brown County, in SE1/4, SE1/4 sec. 16, T21N, R20E. A borehole was drilled in the discrete, sharp-crested, narrow Denmark moraine, which marks the farthest advance of the Glenmore ice onto the Niagara Escarpment. Small traces of dispersed organics (stem, leaves, grass) were encountered incorporated into red clayey diamicton (interpreted to be Glenmore till) at a depth of about 13 to 25.5 ft (4 to 7.8 m) in the borehole. The horizon containing the traces of organic material is overlain by red clayey diamicton that is similar in appearance but lacks any organic material. At a depth of 25.5 to 27 ft (7.7 to 8.2 m), an olive-gray to black paleosol, containing small fragments (0.2 to 0.7 in (0.5 to 2 cm) long) of dark brown wood was encountered. The wood is radiocarbon dated at 11,210±100 B.P. (13,160 calendar) (Beta – 177408). Underlying the paleosol horizon containing wood, is an organic mat, about 2.5 in (6 cm) thick, consisting predominantly of mosses. The moss horizon is dated at 11,820±100 radiocarbon (about 13,830 calendar) years old (Beta – 177409). Underlying the mossy mat is a layer about 10 inches (25 cm) thick of dark gray and black sand with small wood pieces, and dark gray and black, mixed silty-sand, and clay diamicton, with distorted black organic laminae and a few red clay laminae. A layered bed of variable textural composition (silt, clay, silty clay, sand) extends from about 28 to 32 ft (8.5 to 9.7 m), and is underlain by uniform red clay diamicton (interpreted to be Chilton till) to a depth of 47 ft (14.3 m), the maximum depth of the borehole.

The diamicton in map units *Gdhe* and *Gdr* is generally interpreted as basal till. Unit *Ggt* (Glenmore gravel and sand in outwash terraces) is mapped in Manitowoc County. This sediment is moderately well-sorted, well-stratified gravel and sand that was deposited by glacial streams, and that is now in a terrace above a modern floodplain. The gravel and sand was deposited by braided streams in front of end moraines or in valleys running parallel to the glacier margin. The sediment in map unit *Gslp* (Glenmore silt and sand in lacustrine plains) is moderately well-sorted silt, sand, and clay, that was deposited in standing water in low places in the landscape where lakes were dammed by the retreating ice margin.

Two Rivers Member (Lake Michigan Lobe)

The Two Rivers Member is the youngest member of the Kewaunee Formation in the Lake Michigan Lobe and overlies the Two Creeks Forest Bed. The diamicton in map units *Tdh* and *Tdhe* (Two Rivers diamicton in areas of hummocky topography and hummocky end moraines, respectively) is clayey silt that is reddish brown (most commonly 5YR 5/4) and crudely stratified or unstratified. The diamicton is mostly till under a 3- to 6-ft (1- to 2-m) thick layer of mudflow deposits, especially in moraines. Some of the diamicton includes spruce wood and mosses of the Two Creeks Forest Bed that are dispersed or found at the base of the unit. Ten samples of the Two Rivers diamicton have a mean magnetic susceptibility of 1.2 x10^{-3} MKS units. The diamicton is generally compact and uniform, except in the upper few feet where sand lenses and other discontinuities may occur. Classified using the USCS, the diamicton is generally lean clay (CL) but some samples are fat clay (CH). The 10 samples analyzed for grain size distribution have a mean sand:silt:clay ratio of 26:42:33. There are also sand and gravel deposits included in the Two Rivers Member, in three map units on plate 1. The sediments in all the units are moderately well-sorted, well-stratified gravel and sand deposited by glacial streams. Unit *Tgpp* (Two Rivers gravel and sand in pitted plain) has more than 20 percent of original floodplain interrupted by depressions formed by melting ice blocks, and unit *Tgh* (Two Rivers gravel and sand) has less that 20 percent of the original stream bed surface remaining uncollapsed. Unit *Tgt* is a terrace above modern stream. The sediment in all *Tg* units was deposited by braided streams in front of end moraines or in valleys running parallel to the glacier margin.

Dave Mickelson

Glacial history

Summary of glacial events

Much of the record of early glaciations in Calumet and Manitowoc Counties has been erased by more recent glaciations. Glacial deposits older than 130,000 years are present in Illinois and southern Wisconsin (Illinoian Glaciation), so glaciers must have covered this area also. At its maximum extent during the last glaciation (late Wisconsin Glaciation) the Green Bay Lobe was about 80 mi (125 km) wide and 125 mi (200 km) long, and the 125-mi (200-km) wide Lake Michigan Lobe advanced over a hundred miles (several hundred kilometers) farther south into Illinois. Flowlines for ice in both lobes began at the probable ice divide near James Bay, Canada, about 870 mi (1400 km) north of the terminus of the lobe (Dyke and Prest, 1987). Green Bay Lobe ice flowed southward across the east end of Lake Superior and then was channeled southwestward through the Green Bay lowland. The Green Bay Lobe advanced into the lowland in northeastern Wisconsin several times during the late Wisconsin. Farther east, Lake Michigan Lobe ice flowed along a roughly parallel path.

Mapping done for this report can be used to interpret not just the chronology of ice advances and retreats, but to better understand how the glacier itself was behaving (focus box 8).

No radiocarbon dates record early advances of the Green Bay Lobe, but early advances of the Lake Michigan Lobe are radiocarbon dated in Sheboygan County. Three reddish-gray diamicton units are present at depth near Plymouth. Two units appear to have been deposited before 34,610±390 radiocarbon years ago, and another was evidently deposited between 34,610±390 (39,800 calendar) and 26,400±920 radiocarbon (31,000 calendar) years ago (Chapel, 2000; Carlson and others, 2011). Ice reached Illinois by about 25,600 radiocarbon (30,000 calendar) years ago (Hansel and Johnson, 1992). Model results suggest that during the late Wisconsin, the Lake Michigan Lobe advanced into Wisconsin before the Green Bay Lobe (Cutler and others, 2001).

There is a gray, very compact, silty diamicton that has been found in a few places in southeastern Wisconsin beneath the Kewaunee and Holy Hill Formations (Alden, 1918; Bleuer, 1971; Fricke, 1976; Nemchak, 1977) in addition to the Hayton diamicton described in the previous section, and what is presumably the same unit near Horicon (Battista, 1990). It is not clear if the deposition of this diamicton predates, postdates or is contemporaneous with the deposition of the Plymouth diamicton units, or for that matter whether all of the compact, gray units are all from a single glacial advance.

When ice of the late Wisconsin or possibly earlier glaciations advanced south, the ancestral South Branch Manitowoc River, which now flows northward, likely flowed southward to the Sheboygan River and carried large amounts of proglacial meltwater. Proglacial meltwater cut or enlarged the bedrock river channel. As the glacier advanced over the channel, it streamlined the walls of the river channel or formed rock-cored drumlins, as seen for example, where Highway H now crosses the South Branch Manitowoc River, about 4 mi (6.5 km) east of the village of Brothertown (fig. 20).

The ice continued to advance until it reached its maximum extent near Madison, Wisconsin. There it formed the Johnstown moraine (fig. 10A), between 25,000 and 18,000 radiocarbon (30,000 and 21,500 calendar) years ago, and deposited sandy, brown till of the Holy Hill Formation (Colgan and Mickelson, 1997; Colgan, 1999; Syverson and others, 2011). The Green Bay Lobe likely began its retreat as early as 18,000 calendar years ago (Attig and others, 2011; Colgan, 1999; Colgan and others, 2002).

Between retreat from the glacial maximum terminal moraine and subsequent deposition of reddish-brown clay till, the ice margin apparently retreated more than 220 mi (350 km) (fig. 10B). This retreat phase, which ended about 16,000 calendar years ago, is called the Mackinaw Interstadial, and in older literature the Cary-Port Huron Interstadial. There is evidence that ice retreated far enough north in the Lake Michigan basin for lake level to drop well below present during this time (Monaghan and Hansel, 1990). During the next readvance, reddish-brown clayey sediment was transported from the Lake Superior basin into the Lake Michigan and Green Bay basins (Murray, 1953). All tills deposited after this are clayey and reddish-brown, presumably because this sediment was picked up and redistributed by the ice. Although the sediment could have come from the Lake Superior basin into the Green Bay and Lake Michigan basins through subglacial tunnels, based on how well sorted these fine sediments are, it seems likely that they were first deposited into a lake before they were redistributed by the glacier.

Timing of the Mackinaw Interstadial events is not entirely clear because the events were closely spaced in time, and radiocarbon dates in this time period are difficult to interpret.

8 FOCUS BOX Paleoglaciology: Reconstructing ice surfaces

Reconstruction of ice surface profiles, basal shear stress, and other characteristics of former glaciers is called "paleoglaciology." We learn about former glaciers by relating knowledge from the study of present-day glaciers to the distribution of glacial sediments and landforms, and to glacial chronology. Radiocarbon dates on overridden organic matter and stratigraphic information allow estimation of the rate of ice movement, and thus the dynamics of former glaciers.

The average radiocarbon age of the Two Creeks Forest Bed is about 12,000 radiocarbon years, which is calibrated to about 13,900 calendar years ago (Maher and Mickelson, 1996). The radiocarbon age of leaves and twigs in pond sediments beneath the Chilton and Valders till have a calibrated age of about 15,000 calendar years. Rapid ice margin oscillations with advance and retreat rates on the order of about 0.3 mi (0.5 km) per year are indicated by the margin positions of the Green Bay Lobe for the period 16,000 to 13,800 calendar years ago. Glacier bed conditions (rapid sliding and bed deformation) may have contributed to these rapid margin oscillations (Alley, 1991) or there may have been unstable surge advances (Colgan, 1996). This period of rapid ice-margin oscillations coincides with the rapid climate change indicated in the Greenland ice cores at about 14,000 calendar years ago (Cuffey and others, 1995).

Socha and others (1999) reconstructed the ice surface morphology for two readvances—the post–Two Creeks Glenmore and the pre–Two Creeks Chilton advance—during this period of rapid ice-margin oscillations to determine ice thickness and to evaluate the role of low driving stress in the dynamic behavior of the ice margin. These readvances took place during a period of rapid deglaciation when the margin was fronted by a large proglacial lake. The reconstruction indicates that the ice was very thin, the surface slope was low, and that the driving stress was very low.

Chilton advance

The reconstructed ice surface for the Chilton advance rises from about 920 ft (280 m) at the south end of Lake Winnebago to about 1,180 ft (360 m) where it overtopped the Niagara Escarpment. At the southeast end of Lake Winnebago, the slope of the moraine against the escarpment is about 46 ft/mi (8 m/km) within 10 mi (3.4 km) of the terminus (Colgan, 1996). Near the north end of the lake, the Brillion Sublobe protruded through a breach in the Niagara Escarpment. North of the Brillion basin, the ice came up onto the escarpment, but did not completely override it. Although the relief is low, a maximum of about 300 ft (100 m), the ice margin was deflected by the topography both here and along Lake Winnebago. This indicates the ice was very thin. The correlative Valders advance of the Lake Michigan Lobe also had a different margin configuration compared to earlier and later advances. A lobe of ice extended toward the southwest from Lake Michigan depositing thin clayey till more than 6 mi (10 km) beyond the previous western limit of Lake Michigan Lobe ice, yet its extent in the main part of the Lake Michigan basin was less than earlier advances.

The reconstruction for the Chilton advance indicates that within 22 mi (35 km) of the terminus in the Lake Winnebago lowland, the average slope of the ice was 10 ft/mi (2 m/km), the thickness was less than 650 ft (200 m), and the driving stress was about 2 kPa. Ice protruding southeast into the Brillion basin about 12 mi (20 km) was slightly steeper. The surface slope in the Brillion basin was on average about 15 ft/mi (3 m/km), the thickness was less than 490 ft (150 m), and the driving stress was about 3 kPa. Ice that came up on the Niagara Escarpment had an average surface slope of about 57 ft/mi (11 m/km), was less than 490 ft (150 m) thick, and had a driving stress of about 12 kPa within a few miles of the terminus.

Glenmore advance

The Glenmore ice margin has been mapped in the field on the east side of the lobe, but not in the lowland or on the west side where it is difficult to trace because of younger lake sediment. The moraine is also difficult to trace along the west side of the lobe because of bedrock relief. The Glenmore margin did not go as far south into the Lake Winnebago lowland or the Brillion basin, nor as far south on the upland as the Chilton ice. Based on the segment of Denmark Moraine that has been mapped, it appears that the ice during the Glenmore advance had similar characteristics to the ice during the Chilton advance. Within 9 mi (15 km) of the terminus, in the main lobe in the Lake Winnebago lowland and in the Brillion Sublobe, the estimated average surface slope was 10 to 15 ft/mi (2 to 3 m/km), the thickness was less than 490 ft (150 m), and the driving stress was about 2 kPa. The driving stresses

8 FOCUS BOX Paleoglaciology: Reconstructing ice surfaces, *continued*

calculated for both advances are extremely low. Qualitatively, the fact that the ice was deflected by the low bedrock escarpment, supports the low profile, thin ice reconstruction. Inferred from the reconstructions is that the bed was smooth and well lubricated, consistent with a margin advancing over lake sediment. On the upland where the bed consisting of sandy till and bedrock was more permeable and likely better drained, ice surface slopes were steeper and the driving stress was slightly higher.

Modern equivalents

The subglacial conditions at the margin of the Green Bay Lobe during the Chilton and Glenmore advances were probably similar to the basal conditions near the grounding line of Ice Stream B, West Antarctica. The low driving stresses calculated indicate basal conditions similar to Ice Stream B's ice plain, which is the transition zone between grounded ice in the ice stream and floating ice in the ice shelf (Paterson, 1994). The Green Bay Lobe terminated in shallow water that was probably never deep enough to float the ice, so slopes were low, but not zero. The ice was grounded but the effective pressure was probably very low. High subglacial porewater pressure would likely have developed as the ice advanced into the proglacial lake and overrode lake sediment and shale. Ice deformation would have been an insignificant component of flow. The motion of the glacier was likely due to basal sliding, sediment deformation, or a combination of the two mechanisms.

It is also possible that margin configurations for the Chilton and Glenmore advances represent a post-surge terminus, such as suggested by Wright (1971) and Mickelson and others (1981) for readvances of the Lake Michigan and the Green Bay Lobes that occurred after 16,000 calendar years ago. The ice on the upland may have been in a relatively stable position, while ice in the center of the lobe surged out into the lowland. The mechanisms of surging for soft bed glaciers are poorly understood (Paterson, 1994), but it is possible that changing bed conditions associated with the saturated clay till in the lowland could have facilitated a surge.

It may be that the strange configuration of the Valders advance out of the Lake Michigan basin was also a surge, perhaps triggered by the same change in basal conditions as the Chilton advance. Without a better understanding of the mechanisms of surging in soft bed glaciers, and a chronology much more detailed than radiocarbon methods allow, this cannot be tested.

The marginal areas of the Laurentide Ice Sheet were complex and dynamic environments. Models such as surge glaciers and ice plains are supported by geomorphic evidence. While many lobes of the Laurentide Ice Sheet may have acted much like large outlet glaciers during the glacial maximum, it is apparent that other styles of advance and retreat were operating during the rapid deglaciation that followed the major shift in climate that began 16,000 calendar years ago.

Organic matter from the northern part of Michigan's Lower Peninsula was thought to have been deposited during this time (Farrand and others, 1969), but this site has since been redated and it appears that it is Two Creeks age, about a thousand years younger (Larson and others, 1994). Maher and Mickelson (1996) and Maher and others (1998) describe a deposit at the Valders Lime and Stone Quarry where they interpret an ice-free period ending about 13,000 radiocarbon (16,000 calendar) years ago (fig. 10B) (also see focus box 9, "Tundra Environments").

After the Mackinaw Interstadial, ice readvanced over 150 mi (250 km) and the Silver Cliff and Kirby Lake tills were deposited on the west side of the Green Bay Lobe, the Branch River and Chilton tills were deposited on the east side of the Green Bay Lobe, and the Ozaukee and Valders tills were deposited by the Lake Michigan Lobe (fig. 10C). This cold time is known as the Port Huron Phase (Hansel and Johnson, 1992), and it consists of two glacier advances separated by a short retreat interval. Moraines from these advances can be traced from Minnesota to Michigan and Ontario (Mickelson and others, 1983). The Lake Michigan Lobe advanced southward to just south of Milwaukee and deposited reddish-brown, clayey Ozaukee Member. This is well exposed along the lakeshore bluffs, but it is not exposed at the surface in Manitowoc County because deposits of the Valders Member cover it. The time equivalent of the Ozaukee advance in the area covered by the Green Bay Lobe may be the Branch River advance, or what is mapped as Branch River Member on plate 1 could be somewhat younger.

Whenever the northern outlet of Lake Michigan was blocked by glacier ice, the level of Lake Michigan rose again to the Glenwood level (fig. 10C).

9 FOCUS BOX Tundra environments

Organic matter has been found in Manitowoc and Calumet Counties that accumulated some time after the glacial maximum and before the Two Creeks Forest Bed. The organic horizons, found below the Valders and Chilton tills, provide insight into the environmental conditions and the deglaciation history of the area (Maher and Mickelson, 1996).

Pollen, plant remains, mollusks, and ostracodes (small bivalved crustaceans) were found below Valders till in a **glaciolacustrine** unit at Valders Lime and Stone Quarry, in central Manitowoc County. Radiocarbon dates on the organic matter indicate that it was deposited about 14,500 radiocarbon (15,800 to 17,700 calendar) years ago (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 found in tundra and open-forest tundra in northern Canada, were also found in the glaciolacustrine unit, and indicate a cold, open environment (Maher and Mickelson, 1996; Maher and others, 1998).

Plant remains found at a quarry in Sherwood in Calumet County (SW1/4, SE1/4 sec. 30, T20N, R19E) provide an indication of the environment before the ice of the Green Bay Lobe advanced over the area and deposited red clayey Chilton till (Socha, 2007). Quarry operations expose sediment in the end moraine that marks the farthest advance of the Chilton ice onto the Niagara Escarpment. The uppermost sediment is Chilton till, the upper portion of which is fractured and contains the modern soil profile. Dispersed in the Chilton till, and also associated with gray silt stringers, are small pieces of plant remains. The Chilton till is underlain by 1 to 3 ft (less than 1 m) of gray fine sand and silt of the High Cliff Member of the Hayton Formation. Locally (on the stoss-side of a bedrock bump), concentrations or mats of plant remains are present. They are also dispersed in the gray silt. The plant remains, which include leaves, stems, mosses, and twigs, are radiocarbon dated at 13,370±90 B.P. (16,302±428 calendar years) (Beta – 119360). Underlying the gray fine sand and silt is about 1.5 ft (0.5 m) of grayish brown silty-sand diamicton of the Cato Falls Member of the Hayton Formation. The Cato Falls till overlies dolomite bedrock that varies from striated and polished to rough and broken. Striations on the bedrock trend from 125 to 230 degrees.

The gray silt and fine sand is interpreted to be primarily aolian in origin, and was likely deposited by winds near the retreating margin of the ice that deposited Cato Falls till. Locally the windblown gray silt and fine sand may have been deposited in water or slumped into shallow ponds. The plants growing on the gray fine sand and silt were subsequently overridden by the Chilton ice advance. The radiocarbon date on the plants is an indication of when the ice reached its farthest extent onto the escarpment. Small pieces of plant remains have been observed dispersed in the gray silt and fine sand at several other locations in Calumet and adjacent counties. The plant remains found at Sherwood indicate a cold, arctic environment. The plant remains are composed of arctic taxa, with abundant *Dryas integrifolia* leaves, and common seeds of Silene acaulis, much moss, and no evidence of trees (Richard Baker, personal communication, 1998). Barrens and cliffs are the general habitat for both *Dryas integrifolia* and *Silene acaulis* (Maher and others, 1998). Mosses and *Dryas integrifolia* are abundant at the site in Sherwood and are common in the Cheybogan bryophyte bed, but are rare at the site in Valders (Maher and others, 1998).

Thus, a cold, treeless, permafrost environment is indicated by both the Valders and Sherwood organic horizons until at least 13,000 radiocarbon (16,000 calendar) years ago. By 12,000 radiocarbon (14,000 calendar) years ago, permafrost was gone and spruce trees covered the landscape as evidenced by the Two Creeks buried forest.

Shorelines at this elevation south of Manitowoc developed after the onset of the retreat of the Port Huron Phase glacier because the wave-eroded surface is younger than the Valders diamicton (Evenson, 1973; Acomb and others, 1982). Water was dammed in the Lake Winnebago basin as well (Thwaites, 1943) (see focus box 3, “Former Levels of Lake Michigan”).

Mickelson and Evenson (1975) mapped relative ages of till surfaces in Manitowoc and Sheboygan Counties using depth of carbonate leaching. There is a clear difference between surfaces that are younger or older than the Two Creeks Forest Bed (less than 9 in (24 cm) or more than 28 in (70 cm), respectively). Building on this difference, McCartney and Mickelson (1982) and Need (1985) recognized an area where Branch River till is at the surface (plate 1) and they interpret this as a deposit older than the Valders-Chilton advance. Another interpretation is that the segment of end moraine composed of Branch River till could be continuous with the end moraine composed of Chilton till (Socha and others, 1999) (see focus box 5, “Unresolved Stratigraphic Questions”).

The Chilton ice margin appears to have been contemporaneous with the Valders ice margin of the Lake Michigan Lobe (McCartney and Mickelson, 1982) and to represent the late Port Huron advance (fig. 10D). Black (1980) called the moraine between the two units the Menchalville moraine, and argued that it was an end moraine built by Valders ice. We interpret this as an interlobate moraine between the two lobes between Hayton Marsh, near Highway 151 (fig. 16), and Maple Grove (fig. 17). North of Maple Grove (fig. 17) it appears that the Valders ice margin was in a lake along much of its extent to where it rises to higher elevations about 3 mi (5 km) southwest of Denmark (plate 1). Note that the Chilton and Valders advance (fig. 10D) extended farther inland than did the early Port Huron advance (fig. 10C).

The Chilton ice advanced to the same position as the sandy till recessional moraine, and deposited red clay till on top of the sandy till moraine. Well logs indicate that the moraine at Chilton is a composite moraine, formed of red clay till at the surface, and underlain by brown sandy till of the Holy Hill Formation, and gray silty till of the Hayton Formation.

From its maximum upland extent near Chilton (plate 1), the Chilton ice margin retreated to the northwest. The South Branch Manitowoc River possibly cut through the moraine and flowed to the northeast to the confluence with the North Branch Manitowoc River. Water could drain to the east via the Manitowoc River when the Valders ice margin had retreated to the east of Cato Falls (now with bedrock at an elevation of about 790 ft). However, as retreat began, a large lake developed in the Manitowoc River basin in the present positions of Hayton and Collins Marshes that was dammed by retreating Valders ice. This was probably short lived, and it would have drained as soon as Valders ice retreated from the present position of the lower Manitowoc River. With further retreat of the Valders ice margin, water could drain to the east into Lake Michigan when the area near what is now Manitowoc became ice-free.

Ice retreated dramatically during the Two Creeks warm time. The northern outlet of Lake Michigan was ice-free and lake level likely dropped at least 300 ft (90 m), exposing the floor of Green Bay and likely a substantial amount of the bed of Lake Michigan (fig. 10E). By this time, climate had warmed and spruce forest had migrated into this area. Remains of this forest are preserved as the Two Creeks Forest Bed, one of the best studied geological sites in the Great Lakes area, and a site that is known to geologists around the world (McCartney and Mickelson, 1982; Maher and Mickelson, 1996) (see focus box 7, “More about the Two Creeks Forest Bed”). Spruce trees from the Two Creeks Forest Bed have been radiocarbon dated many times and have a fairly large range, but 12,050 to 11,750 radiocarbon (14,100 to 13,800 calendar) years ago seems to be the best estimate of when the forest grew here.

Ice readvanced, covering the forest bed and depositing a layer of till. In the area covered by the Green Bay Lobe, the tills are from the Middle Inlet and Glenmore Members; in the Lake Michigan basin, till of the Two Rivers Member covers the bed. When ice advanced to the Denmark moraine (Green Bay Lobe) and the Two Rivers Moraine (Lake Michigan Lobe), water was again dammed in front of the ice margin in the Green Bay–Fox River lowland, but water in Lake Michigan rose only to the Calumet level (fig. 10F). The Lake Michigan Lobe ice terminated in water, flooding the area from the Two Rivers lowland to the eastern side of Lake Michigan. The Green Bay Lobe ice, which deposited the Glenmore till, advanced as far south as what is now the north end of Lake Winnebago and as far southeast as the Villages of Hilbert and Brillion in the Brillion basin (plate 1). The position of the ice margin in the Brillion basin is not marked by a distinct ridge, possibly because the ice advanced into the broad lowland area and filled in the river valley. Ice that reached the escarpment built a distinct fairly sharp-crested moraine such as the moraine north of Brillion (plate 1).

North of the village of Hilbert, in the NW1/4 sec. 19, T20N, R19E, the Two Creeks Forest Bed was encountered at an elevation of about 810 ft in a borehole drilled north of the North Branch Manitowoc River (see focus box 7, "More about the Two Creeks Forest Bed"). Chilton till directly underlies the forest bed, indicating a lake was not present in front of the Glenmore margin as ice advanced into the Brillion basin. When the Glenmore ice was advancing and when it reached its maximum extent, meltwater could drain into the North Branch Manitowoc River, and be carried to the Manitowoc River, which flowed eastward to Lake Michigan. The outlet of the Manitowoc River was not blocked by ice, because at this time in the Lake Michigan basin, the Two Rivers ice margin was north of the Manitowoc River outlet (plate 1).

The Glenmore ice did not advance far beyond the north end of present day Lake Winnebago (plate 1), so the main part of the Lake Oshkosh basin was ice-free and was draining to the west, with only meltwater from the Brillion sublobe draining to the east. Once the Glenmore ice margin started to retreat north, off the Denmark moraine, low areas near Sherwood were exposed and glacial Lake Oshkosh could drain into the North Branch Manitowoc River channel in several places (fig. 21). The largest of these is in the W½ sec. 15, T20N, R19E (Woodville Township). The elongate features in sec. 23 and 25, T20N, R19E and in sec. 31, T20N, R20E may be erosional remnants of a flood channel. Large **lag boulders** are present in sec. 30, T20N, R20E. A smaller channel in sec. 28 and 29, T20N, R19E, likely formed when the ice margin was against the upland (plate 1).

The Glenmore ice margin continued to retreat to the northeast in the Fox River lowland and opened up the Neshota outlet at an elevation of 765 ft (233 m), the Kewaunee River at an elevation of 685 ft (209 m), and the Ahnapee River at an elevation 635 ft (194 m) (Hooyer and Mode, 2008). Shorelines and stratigraphy in the Green Bay–Fox River lowland are further described by Hooyer (2007) and Hooyer and Mode (2008).

Lake Michigan remained at the Calumet level until its northern outlet opened and the lake level fell again. Following retreat of the Glenmore and Two Rivers ice by about 11,000 radiocarbon (13,000 calendar) years ago, vegetation grew back onto the landscape very rapidly, because by this time climate had warmed, there was no permafrost, and there was a seed bank readily available. It seems likely that mastodons and mammoths along with many small mammals were occupying the landscape from the time of ice retreat. Ice did not enter eastern Wisconsin again, although the glacier did fill the east end of the Lake Superior basin and produced a moraine in the upper peninsula of Michigan about 9,900 radiocarbon (11,500 calendar) years ago (Lowell and others, 1999).

The post-glacial (Holocene) history of Manitowoc County has not been studied in detail. The bluffs along the Lake Michigan shoreline are continually experiencing instability, slumping, and retreat (Hadley, 1976; Chapman and others, 1997). In the Point Beach area deposition took place, forming the beach ridges and dunes we see today. Farther inland, rivers have established well-developed courses with floodplains (mapping unit *a*, **alluvium**, on plate 1). Beaches and the alluvial deposits associated with present streams and rivers are the youngest surficial deposits in the study area.

In retrospect

The landscape of Calumet and Manitowoc Counties is almost entirely derived from glaciers that covered the land beneath thick glacier ice from about 25,000 radiocarbon (30,000 calendar) years ago until it finally retreated out of the counties for the last time about 11,000 radiocarbon (13,000 calendar) years ago. The glacier left behind deposits of clay, sand, and gravel that have a great influence on our use of the land today.

As population grows and land use becomes more intensive, we must try to use the land wisely. One way to encourage this is to understand the limitations placed on land use by different surface materials.

Calumet and Manitowoc Counties are important because the best record of environmental conditions between about 13,500 and 11,500 radiocarbon (16,400 and 13,400 calendar) years ago anywhere in the Great Lakes area is preserved here. The climate record preserved in Manitowoc and Calumet Counties may help us understand future rapid climate change.

This report also provides an overview of how the landscape evolved and how glacial sediments are distributed today. Hopefully after reading it and using the map to better understand the distribution of soils and landforms, you will be able to contribute to wise land use decisions.

Glossary

alluvium – Sorted or semi-sorted clay, silt, sand, or gravel deposited by a stream, river, or other body of moving water, in a stream bed, or floodplain; especially silt or clay deposited during a flood.

B-horizon – A layer of soil formed by accumulation of weathering products of minerals (typically clay and iron) from above.

braided stream – Stream with multiple short-lived, migrating channels that flow together and then apart creating a braided pattern on the floodplain. Typical of outwash streams.

clay – Mineral soil particles less than 0.002 mm in diameter.

cohesion – The component of the shear strength of a sediment that is independent of friction between grains.

cross-bedding – Thin layers within a unit that are at an angle to the main bedding.

delta – A triangular or fan-shaped deposit of sediment formed where a river or stream enters a standing body of water such as a lake or ocean.

diamicton – Mix of unsorted clay- to boulder-size particles typically deposited by mudflows, landslides, or glaciers.

drumlin – A low, elongated or oval hill formed at the base of a glacier. The long axis is parallel to the flow path of the glacier and commonly has a blunt nose pointing in the direction from which the ice approached.

end moraine – Ridge of glacial sediment, typically diamicton, that is formed along the glacier margin. See *moraine*.

esker – A narrow, generally winding ridge of stratified gravel and sand deposited by a stream flowing in a tunnel beneath a glacier.

foreset beds – Inclined sediment beds accumulated as sediment grains tumble and slide down the steep frontal slope of a delta.

friction angle – The shear resistance caused by friction between grains in a soil.

glaciofluvial deposits – Alluvium deposited by streams flowing beneath or away from a glacier. The deposits are stratified and occur as moulin kames, eskers, and outwash plains, hummocky sand and gravel, etc.

glaciolacustrine – Sediment deposited in a glacial lake. The sediments are silt, clay, or fine sand and are laminated, thin-bedded, or massive, if deposited in deep water, or sand or gravel if deposited in beach or delta environments.

gleyed soil -– A soil having one or more neutral gray horizons as a result of waterlogging and lack of oxygen. The term "gleyed" also designates gray horizons and horizons having yellow and gray mottles as a result of intermittent waterlogging.

gravel – Rounded or angular fragments of rock ranging from sand size to 3 in (2 mm to 7.5 cm) across.

hummocky topography – Landscape with many small hills (hummocks) separated by depressions (kettles). Produced by collapse of a thick layer of sediment as buried glacier ice melts.

hydraulic conductivity – The ability of a soil or rock to transmit groundwater.

igneous rock – Rock formed from the cooling of molten materials (magma). Examples are granite and basalt.

interlobate – Between lobes of glacial ice. The Kettle Moraine formed in the interlobate zone between the Lake Michigan and Green Bay Lobes.

isostatic depression – Depression of the earth's crust due to the weight of overlying glacier ice.

isostatic rebound – Uplift of the earth's crust due to removal of weight of overlying glacier ice.

kettle – Depression in the land surface due to the melt-out of buried glacier ice.

lag boulders – Boulders left behind after the removal of smaller particles by flowing water.

loam – Soil material that contains 7 to 27 percent clay particles, 28 to 50 percent silt particles, and less than 52 percent sand particles.

metamorphic rock – Originally sedimentary or igneous rock that has been subjected to heat and high pressure and changed in composition and arrangement of mineral grains.

moraine – Ridge of glacial sediment, typically diamicton, formed along the edge of a glacier. Usually the same as *end moraine*.

Munsell color notation – A designation of color by degrees of the three simple variables – hue, value, and chroma. For example, a notation of 10YR 6/4 is a color of 10YR hue, value of 6, and chroma of 4.

outwash – Stratified sediment, usually mostly sand and gravel, carried away from a glacier by meltwater streams. Coarse particles are usually deposited nearer the glacier than the finer-grained sediments.

overconsolidation – A property of sediment that indicates it has been under greater load at some time in the past than it is when sampling takes place. This is often interpreted as being due to the weight of former glaciers.

paleosol – A soil with distinct morphologic features that formed under conditions that are no longer present at the site. Usually the soil-forming process was interrupted by burial.

palimpsest – A landscape that retains landforms from an older land surface.

peat – A deposit of plant remains that accumulated in a wet environment such as a bog or swamp.

pitted outwash – Deposit of outwash sand and gravel that has kettles developed in it, but that retains parts of its pre-collapse stream bed surface. Formed when outwash is deposited over melting glacier ice, and buried ice later melts out.

Quaternary Period – The most recent (and present) period of geologic time. It began about 2 million years ago.

sand – Individual rock or mineral fragments from 0.05 to 2.0 mm in diameter.

sediment – Particles that are transported by water, wind, glacier ice, or gravity, and are eventually deposited.

sedimentary rock – Rock that is made up of cemented sediment particles, organic matter, or chemical precipitate.

silt – Individual mineral particles that range in diameter from the upper limit of clay (0.002 millimeter) to the lower limit of very fine sand (0.05 millimeter).

spit – Elongate body of sand deposited by waves and currents, typically with a lagoon behind.

stratified – Arranged in strata, or layers. The term refers to geologic material. Layers in soils that result from the processes of soil formation are called horizons; those inherited from sedimentary processes are called strata, laminations, or beds.

striations – Scratches on a bedrock surface or on cobbles and boulders that are produced by differential movement of particles at the bottom of the glacier.

supraglacial sediment – Sediment that accumulates on a glacier surface and eventually is deposited as underlying glacier ice melts. Often forms hummocky topography.

terrace – A former floodplain surface that was abandoned as the stream cut its channel lower. Typical of outwash rivers.

till - A predominantly unsorted and unstratified, homogeneous mixture of sand, silt, clay, and gravel, that was deposited by or from glacier ice without being sorted by meltwater or sediment gravity flow processes.

unconformity – A period of time missing in the rock record. May be caused by or nondeposition or erosion.

Wisconsin Glaciation – The last major glacial event in central and eastern North America. The late Wisconsin Glaciation began about 30,000 calendar years ago and ended about 13,000 calendar years ago in Wisconsin, but ice cover lasted longer in Canada.

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