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GLACIAL HISTORY OF SLEEPING BEAR DUNES NATIONAL LAKESHORE, NORTHWESTERN LOWER MICHIGAN: EVIDENCE FOR LATE PLEISTOCENE AND HOLOCENE LAKE LEVEL CHANGES

presented by

Todd Erick Wallbom

has been accepted towards fulfillment
of the requirements for

M.S. degree in Geological Sciences

Major professor

Date April 24, 2000

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Todd Erick Wallbom

A THESIS

Submitted to
Michigan State University
in partial fulfillment of the requirements
for the degree of

MASTER OF SCIENCE

Department of Geological Sciences

2000

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ABSTRACT

GLACIAL HISTORY OF SLEEPING BEAR DUNES NATIONAL LAKESHORE, NORTHWESTERN LOWER MICHIGAN: EVIDENCE FOR LATE PLEISTOCENE AND HOLOCENE LAKE LEVEL CHANGES

By

Todd Erick Wallbom

Despite several decades of research, fundamental questions have challenged glacial geologists working in the Sleeping Bear Dunes National Lakeshore ("the Lakeshore") area located in northwestern lower Michigan. These include: 1) what is the origin of the high, prominent, transverse ridges that characterize the Lakeshore, 2) did Greatlakean ice extend into the region, and 3) what glacial lake stage followed the ice.

Recent surficial mapping at the Lakeshore by the author suggests the following: that the transverse ridges oriented roughly northwest-southeast likely represent subaqueous crevasse-fill deposits formed within disintegrating Port Huron ice over Port Huron-age till; that there is no conclusive evidence for the Greatlakean ice advance but it probably did extend as thin lobes into the lowlands; and a strand at about 222 to 225 m elevation extends along many ridges at the Lakeshore and probably represents the Calumet stage of glacial Lake Chicago.

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INTRODUCTION

General Statement

Sleeping Bear Dunes National Lakeshore (or "the Lakeshore"; T.27 to 32N.,R.13 to 16W.) is located in northwestern lower Michigan (Figures 1, 2) and is managed by the National Park Service (NPS). It includes about 30 km of scenic Lake Michigan shoreline in Leelanau and Benzie Counties and covers more than 72,000 acres of land and deeded lake surface area, as well as North and South Manitou Islands. The Lakeshore area is mostly open wilderness with a few old-growth forest stands, and includes some undeveloped land, active orchards, and abandoned fields being returned to pre-settlement conditions by the NPS.

The Lakeshore is characterized by a series of 40 to 129 m high, prominent ridges that traverse the region, are oriented roughly northwest to southeast, and often form dramatic headland bluffs along the Lake Michigan shoreline. Between the ridges are broad, open, arcuate-shaped, lowland plains that inland from shoreline embayments between the bluffs. Capping many of the bluffs are extensive dune fields that include Michigan's most famous dune, the Sleeping Bear Dune, from which the name of the Lakeshore is derived.

Although it has long been recognized that glaciers from the Lake Michigan Lobe (LML) intermittently covered the Lakeshore and that most of the landforms there are of glacial origin (Leverett and Taylor, 1915; Waterman, 1926; Dow, 1938; Calver, 1947; Melhorn, 1954; Taylor, 1990), many questions still remain regarding

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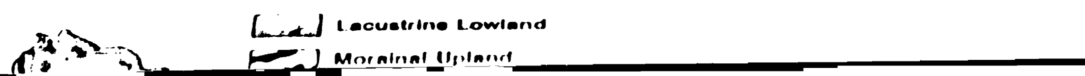
regarding the deglaciation history of the region. These include the following: 1) how and when were the prominent ridges at the Lakeshore formed, 2) did Greatlakean ice advance into the region, and 3) what were the elevations of lake levels in the northern Lake Michigan basin pre- and post-deglaciation?

Purpose of this Investigation

Recent geologic investigations (of glacial deposits and water-plane elevations at the Lakeshore) by the author have uncovered evidence regarding deglaciation and relative position of the ice margin since about 13,000 yrs B.P. (Schneider, 1990; Schneider and Hansel, 1990). These findings have resulted in the identification and facies descriptions of fourteen glacial and post-glacial units at the Lakeshore along with landform interpretations and relative time of formation estimates. In addition, they have generated new elevation data and interpretations for five strandlines in the region.

The key issues being addressed in this thesis are: 1) map and describe the surficial geologic deposits at the Lakeshore and discuss the stratigraphy, time of deposition, and origin of the ridges; 2) define the extent of the Greatlakean ice-margin; and 3) outline the glacial and post-glacial lake history in the Lakeshore area from glacial Lake Chicago to modern Lake Michigan with emphasis on differentiating high lake levels of glacial Lake Chicago.



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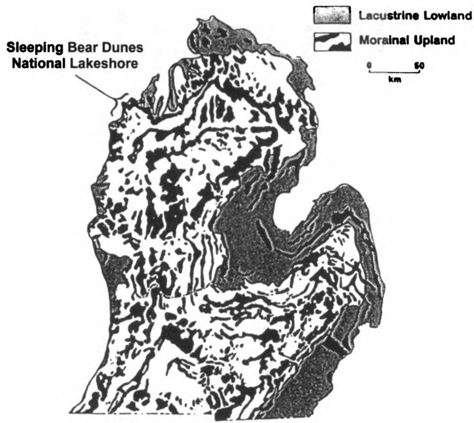

Figure 1. Map showing location of study area (from "Geology of the Southern Peninsula of Michigan", Leverett and Taylor, 1915).

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Historical Background

The Lake Michigan Lobe and its fluctuations

The State of Michigan was last glaciated during the Late Wisconsinan when the Laurentide ice sheet was at its maximum around 20,000 to 12,000 yrs B.P (Melhorn, 1954; Hough, 1958; Evenson, 1973; Drexler, 1975; Eschman, 1985; Hansel et al., 1985). During this period, the LML ice-margin underwent multiple oscillations leaving behind a complex sedimentological record in eastern Wisconsin, Illinois, Ohio, and southern Michigan, for which a wealth of information is available (Chamberlin, 1877; Alden, 1904; Leverett and Taylor, 1915; Alden, 1918; Thwaites, 1943; Thwaites and Bertrand, 1957; Evenson, 1973; Evenson et al., 1974; Mickelson and Evenson, 1975; Farrand and Bell, 1982; Eschman, 1985; Hansel et al., 1985).

During the Late Wisconsinan (Woodfordian Substage, or oxygen isotope Stage 2) along the eastern side of the Lake Michigan basin, four major episodes of LML ice advances (Figures 2, 3) were marked by a series of morainal ridges (terminal, recessional) built along the ice front (Farrand and Eschman, 1974; Eschman, 1985). These are, from oldest to youngest, the Tekonsha, Sturgis-Kalamazoo, Lake Border, and the Port Huron morainal complex; part of which includes the Manistee moraine (Figure 4), culminating roughly at about 21,000, 15,500, 14,100, and 13,300 to 12,200 and 12,000 to 11,700 yrs B.P. (for the Port Huron morainal complex) respectively, until the state was free of ice at around 10,000 yrs B.P. (Thwaites, 1943; Bretz, 1951; Thwaites and Bertrand, 1957;

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Broecker and Farrand, 1963; Evenson, 1973; Farrand and Eschman, 1974; Eschman, 1985; Hansel et al., 1985; Monaghan et al., 1986; Monaghan and Hansel, 1990; Taylor, 1990). Advances were rapid over the basin; Kempton and Gross (1971) estimate an advance rate of 62 m/year for the LML between 23,000 and 18,000 yrs B.P.

A period of retreat, named the Erie Interstade, occurred at about 15,600 yrs B.P. (Dreimanis and Goldthwait, 1973; Monaghan et al., 1986) and was followed by several more closely spaced ice advances. During the Erie Interstade, the Lake Michigan basin was, for the most part, ice-filled as no clear record exists of a glacial lake occupying the basin (Eschman, 1985).

Following the Erie Interstade, several more LML advances occurred in the roughly two-thousand year period between about 15,500 and 13,000 yrs B.P. (Figure 3). During this period, named the Port Bruce stade (Farrand and Eschman, 1974), the front of the LML fluctuated repeatedly with only minor oscillations in eastern Wisconsin, northeastern Illinois, and northwestern Indiana (Leverett and Taylor, 1915; Monaghan et al., 1986; Monaghan and Hansel, 1990).

A major advance at about 14,100 yrs B.P. reworked lake sediment and deposited it in the Lake Border moraine along the rim of the eastern Lake Michigan basin. Following this, the ice rapidly retreated north of the Glacial Grand Valley during the Mackinac Interstadial (Leverett and Taylor, 1915; Melhorn, 1954; Larson et al., 1994), which lasted until about 13,300 yrs B.P.

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(Farrand and Eschman, 1974; Fullerton, 1980; Eschman, 1985; Hansel et al., 1985).

Extent of ice advances in northwestern lower Michigan. Following the Mackinac Interstadial, at least three main LML advances occurred in the Lake Michigan basin prior to final retreat (Figure 3). These include at least two closely-spaced Port Huron ice advances and the Greatlakean, or Two Rivers, ice advance. These advances occurred between about 13,300 to 12,200 yrs B.P. and between about 12,000 to 11,700 yrs. B.P. respectively (Broecker and Farrand, 1963; Evenson, 1973; Farrand and Eschman, 1974; Eschman, 1985; Taylor, 1990; Larson et al., 1994).

The Port Huron stade ice advances, between 13,300 and 12,200 years B.P. (Farrand and Eschman, 1974; Fullerton, 1980; Hansel et al., 1985; Blewett and Winters, 1995), occurred at least twice in southern Michigan. The first Port Huron ice advance formed the outer Port Huron moraine in Michigan (Figures 2, 3) between 13,300 and 13,000 yrs B.P. (Farrand and Eschman, 1974; Hansel et al., 1985; Blewett, 1990) and can be traced from upstate New York to eastern Wisconsin. The ice margin then stagnated and retreated northward from the outer Port Huron moraine (Bretz, 1959; Hough, 1963; Farrand and Eschman, 1974; Hansel et al., 1985; Taylor, 1990) and meltwater quickly built outwash of the Mancelona Plain which now occupies the swale between the outer and the inner Port Huron moraines (Leverett and Taylor, 1915; Martin, 1957; Evenson, 1973; Farrand and Eschman, 1974; Blewett, 1990; Blewett and Winters, 1995).

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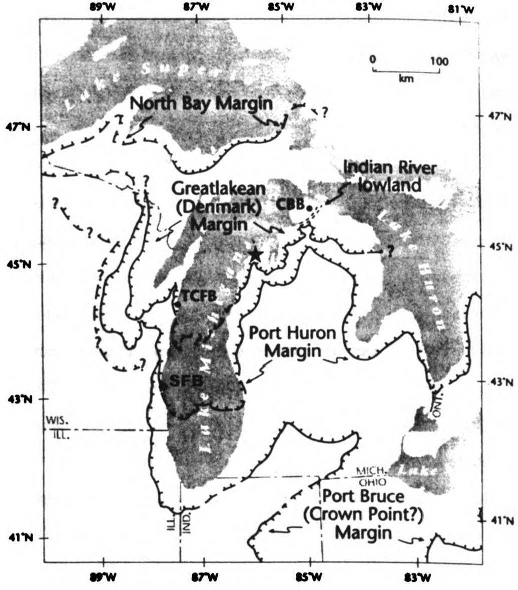

Figure 2. Map showing the southern margin of landforms associated with Port Bruce, Port Huron, Greatlakean, and North Bay stades. Also shown are the locations of Sleeping Bear Dunes National Lakeshore (star), Cheboygan bryophyte bed (CBB), Indian River lowland, Two Creeks forest bed (TCFB), and St. Francis Bluffs (SFB). (adapted from Larson et al. 1994 and Rovey II and Borucki, 1995).

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The second Port Huron advance (Figures 2, 3) occurred when ice advanced to the position of the Manistee moraine between 13,000 to 12,200 yrs B.P. (Evenson, 1973; Farrand and Eschman, 1974; Eschman, 1985; Hansel et al., 1985; Taylor, 1990; Blewett and Winters, 1995). This moraine is the western continuation of the inner Port Huron moraine in northeastern Michigan (Leverett and Taylor, 1915; Melhorn, 1954; Hough, 1958; Evenson, 1973; Farrand and Eschman, 1974; Drexler, 1975; Eschman, 1985; Hansel et al., 1985; Taylor, 1990) and marks the final ice advance of the LML during the Port Huron stade in northwestern lower Michigan (Melhorn, 1954; Hansel et al., 1985; Taylor, 1990). It is also the most conspicuous topographic feature within the Lakeshore and was first described and mapped by Leverett and Taylor (1914; 1915). Since then, its position has been only slightly modified on maps produced by Melhorn (1954), Martin (1957), and Farrand and Bell (1982).

The Manistee moraine, as mapped by Melhorn in 1954 (Figure 4), trends southwest to northeast and eventually parallels the Lake Michigan shoreline from the city of Manistee (not shown) northward to Glen Lake, Leelanau County. From Glen Lake, it bends eastward around the head of Grand Traverse Bay, and then turns northward again near the south end of Elk and Torch Lakes where it becomes subdued near Clam Lake in southwestern Antrim County. Near Rapid City the Manistee moraine bends sharply to the northeast where it appears as a series of isolated ridges. At this point, these ridges are no longer collectively named as part of the Manistee moraine but instead as part of the inner Port Huron morainal system within a small outwash plain located behind the inner

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Port Huron moraine. Despite the name change, many workers believe that both the Manistee and the inner Port Huron moraines were formed contemporaneously (Farrand and Eschman, 1974, Hansel et al., 1985, Larson et al., 1994).

Following the Port Huron stade and construction of parts of the Mancelona Plain and the Manistee moraine (including the inner Port Huron moraine to the east), the ice margin retreated north during the Twocreekan Interstadial (Figure 3) at about 11,800 yrs B.P. (Evenson, 1973; Farrand and Eschman, 1974; Hansel et al., 1985). Port Huron meltwater probably completed building the remainder of the Mancelona Plain during this time (Leverett and Taylor, 1915; Blewett, 1990; Blewett and Winters, 1995).

The final advance of the LML was during the Greatlakean stade (often referred to by earlier workers as the "Valders" glaciation) when ice advanced southward into the basin (Figure 2) between about 11,800 and 11,200 yrs B.P. (Thwaites and Bertrand, 1957; Evenson, 1973; Schneider, 1990; Schneider and Hansel, 1990).

Arguments on the extent of Greatlakean ice in the Sleeping Bear Dunes National Lakeshore region. The advance of Greatlakean-Valders ice is not marked by any distinctive landforms in northwestern Michigan or in eastern Wisconsin and so its ice margin extent, thickness, and correlative till have been debated for several decades (Melhorn, 1954, Thwaites, 1957, Evenson, 1973, Taylor, 1990, Larson et al, 1994). Some researchers have pointed out that ice from the Greatlakean stade ba

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stade barely extended out of the Lake Michigan basin and then only for a few km as tongues of ice into low areas leaving scattered till deposits superimposed on older Port Huron drift (Thwaites and Bertrand, 1957; Evenson, 1973; Hansel et al., 1985; Larson et al., 1994). On the other hand, Melhorn (1954) suggested that ice of the Greatlakean (Valders) stade was extensive, advanced at least as far as the inner Port Huron (Manistee) moraine, and left a thin carpet of red clay till over the browner, coarser Port Huron-age till. In addition, Taylor (1990), from his studies in a few areas of the Lakeshore, identified a patchy red, silt clay till deposit with the "same character" as Melhorn's "Valders" till (pg. 106) superimposed over older Port Huron-age deposits located several km inland. Based on this evidence, he argued that the till is likely Greatlakean in age and is widespread at the Lakeshore. Therefore, he concluded that Greatlakean ice covered large regions proximal to the shoreline.

Elsewhere in the Great Lakes basin, Greatlakean ice is well recorded by stratigraphic evidence at Two Creeks forest bed (Figures 2, 3) located in eastern Wisconsin (Broecker and Farrand, 1963; Evenson, 1973; Schneider, 1990) and Cheboygan bryophyte (moss) bed (Figure 2) located about 200 km to the northeast in Cheboygan, Michigan (Larson et al., 1994). At both places, a red till is superimposed on lacustrine deposits that in turn are superimposed on the organic beds. The age of the Two Creeks forest bed is given as 11,850 ± 100 yrs B.P (Broecker and Farrand, 1963), and 11,910 ± 120 yrs B.P. (Schneider, 1990). The bryophyte bed is given as 12,000 to 11,700 yrs B.P. (Larson et al., 1994).

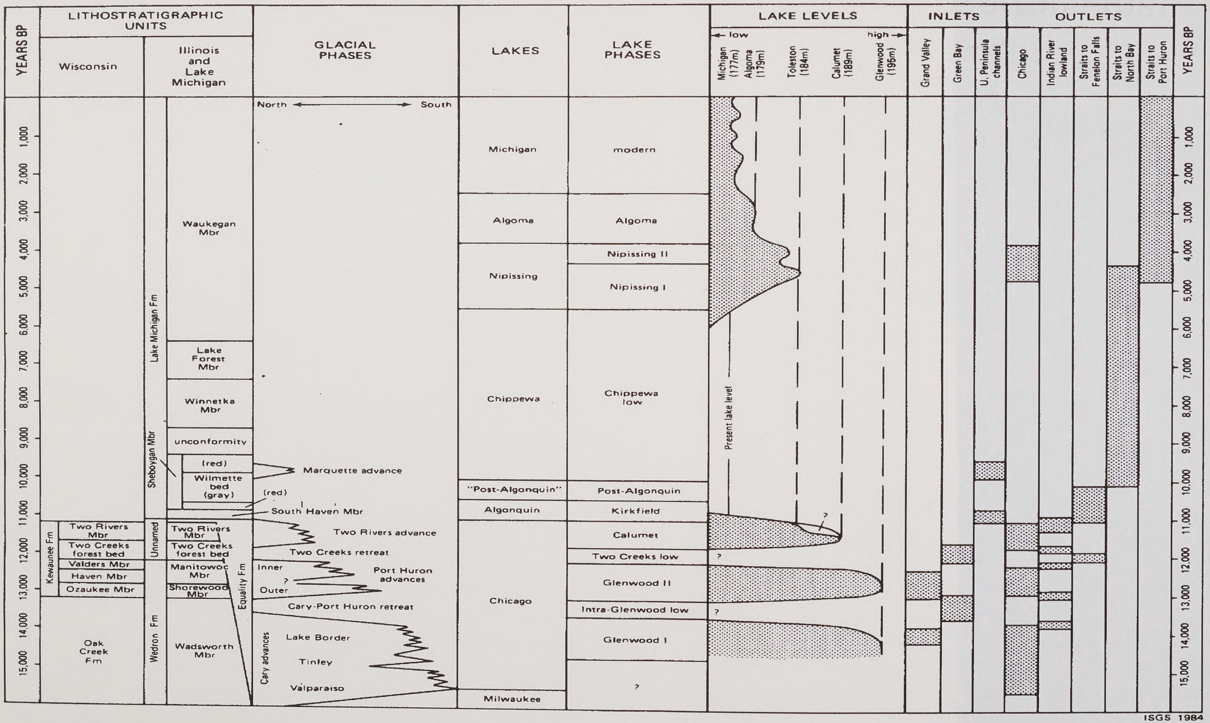

Figure 3. Glacial time/stratigraphy of the Lake Michigan basin. Shaded area indicates glacial lake stage, inlet, and outlet activity. Radiocarbon ages are estimates (from Hansel et al., 1985)

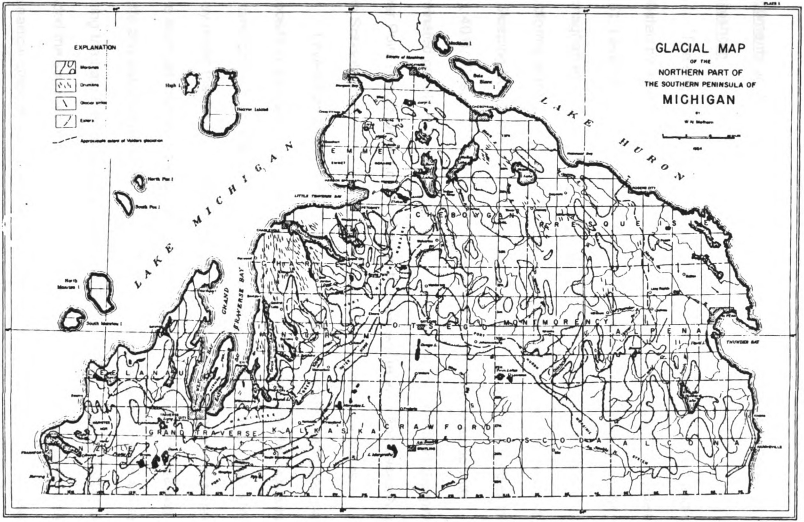

Figure 4. Glacial map showing distribution of moraines in the northern part of the lower peninsula of Michigan (Melhorn, 1954).

Arguments on the origin of the ridges at Sleeping Bear Dunes National Lakeshore

The origin of the ridges at the Lakeshore have been described and/or debated by various researchers for over 175 years (Schoolcraft, 1821; Desor, 1851; Leverett and Taylor, 1915; Dow, 1938; Calver, 1947; Melhorn, 1954; Monaghan et al., 1986; Taylor, 1990). Schoolcraft (1821) was the first to study landforms at the Lakeshore and interpreted sandy deposits in the exposed bluff at Sleeping Bear Point as "a bank of sand 200 feet high and 8 or 9 miles long" (pg. 401). Desor (1851) also noted thick sand accumulations along with laminated pebble-free clay occurring near the bases in the bluffs at Sleeping Bear Point (Figure 5) and other bluffs in the Lakeshore area and basically agreed with Schoolcraft that these bluffs were nothing more than large dunes.

Leverett and Taylor (1915) described similar laminated sand and clay deposits in the Lakeshore bluffs and were the first to state that the deposits were derived glacially yet expressed "much uncertainty" as to the origin (pg. 314). They noted that the elevations of the clay are sometimes 60 m above present lake level yet interpreted it as having been deposited in water and mapped the bluffs and associative ridges as part of the Manistee moraine (Figure 4) formed during the Late Pleistocene (Leverett and Taylor, 1914). Leverett (1914) also added the suggestion that deposition occurred not in a large proglacial lake but "in narrow strips of water between lobes of ice that occupied the great valley-like lowlands

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lowlands now occupied by the arms of Grand Traverse Bay, Pine Lake, Torchlight Lake, and lesser lakes of the Grand Traverse region." (pg. 315).

Dow (1938), like Leverett and Taylor (1915), noted the outcroppings of a thick pink laminated clay spread throughout the bluffs and speculated that "...they (the ridge clays) indicate a widespread deposit, or (date) earlier than the surface morainal drift, but rest upon older glacial deposits." (pg. 429).

Calver (1947), working on the geomorphology in Benzie County and other nearby sections of the Lakeshore, was more intrigued by the morphology and apparent linearity of the local ridges and expressed doubt that they represent moraines. Instead, he suggested that "these ridges may have formed as crevasse fillings in a stagnant block of ice that once occupied the Platte Valley." (Part II, pg. 13). Melhorn (1954), on the other hand, expanded on Calver's work by including some sedimentologic and interpretive discussions about the ridges and described them as composed mainly of red-clay till and formed by the final advance of the LML (Valders/Greatlakean). Near Honor, Taylor (1990) noted thick deposits of "outwash sands and gravels" overlain by red, sandy, pebble-rich till underlying the crests of many of the ridges (pg. 106). While no explanation is provided for the origin of the "outwash", he suggests that the till probably represents some of the "older sandy tills of probable Port Huron age" (pg. 106).

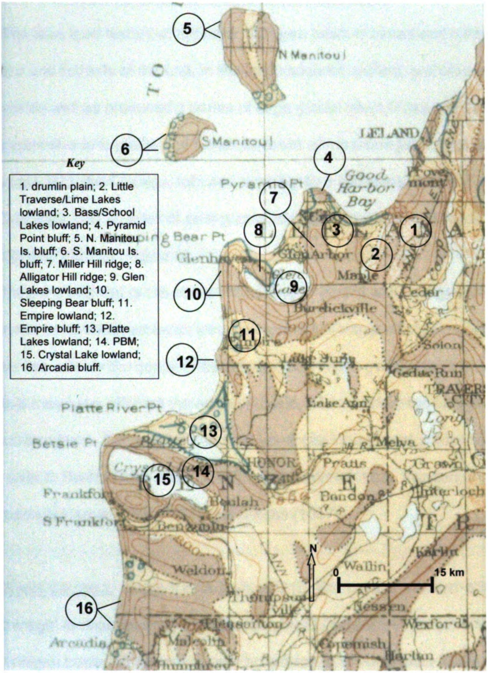

Figure 5. Map of the Lakeshore and vicinity showing major landforms (slightly modified from Leverett and Taylor's map, 1915).

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Glacial and post-glacial lake levels in the Lake Michigan basin

The lake level history of the Lake Michigan basin is varied and complex. Advances and retreats of the LML in the basin scoured, gullied, and dissected landforms as well as produced a series of large glacial lakes (Figure 3) that formed extensive lake plains and beach deposits within the region . These lakes included glacial Lake Chicago, followed possibly by a low interphase lake named Lake Chippewa, and a series of post-glacial interbasin lakes named Algonquin, Nipissing, and Algoma (Figure 3).

Proper treatment of the entire chronology of glacial and post-glacial lake levels in the Lake Michigan basin would require a full-length discussion and therefore lies outside the scope of this study. Only those general facts about levels that may have affected the northern basin, and therefore the Lakeshore, are described here. For excellent summaries of glacial and post-glacial lakes, please refer to Bretz (1951; 1959; 1963), Hough (1958; 1963), Farrand (1982), and especially Hansel et al. (1985) and Larsen (1987).

Glacial Lake Chicago. Leverett and Taylor (1915) first coined the term 'glacial Lake Chicago' to describe the proglacial lake that formed when ice sat in the Lake Michigan basin. They recognized the existence of three main levels or "stages" of this lake in the southwestern Lake Michigan basin: the Glenwood level at 195 m above sea level (a.s.l), the Calumet level at 189 m, and the Toleston (Algonquin?) level at 183 to 184 m (Figure 3).

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Glenwood. The Glenwood phase consisted of two lake levels, Glenwood I and II, both at 195 m, and controlled by an outlet at Chicago. These levels were produced between 14,500 to 12,500 yrs B.P. as the LML retreated northward from the Lake Border moraine (Leverett and Taylor, 1915; Hansel et al., 1985; Taylor, 1990).

Calumet. The main Calumet phase, at 189 m, was produced at the onset of the Greatlakean stade approximately 11,800 yrs B.P. (Evenson, 1973; Farrand and Eschman, 1974; Schneider and Hansel, 1990). According to Hansel et al. (1985), the Calumet phase occurred when LML ice readvanced to the position of the Manistee moraine in northwestern Michigan and the Two Rivers moraine in northeastern Wisconsin. As previously stated, this advance of the lobe did not build the Manistee moraine since the moraine was built during the Port Huron stade. The advance of the lobe during the Calumet phase subsequently blocked outlets at the Straits of Mackinac and Indian River thereby causing Lake Chicago waters to rise to a high lake level but still several meters lower than the Glenwood high lake level (Hansel et al., 1985).

The Calumet rise drowned the forest at Two Creeks, Wisconsin (Figure 2) before the Two Rivers Till was deposited (Evenson, 1973) and buried interglacial organic matter with lake sediment at Cheboygan, Michigan (Figure 2) before the Munro Till was deposited (Larson et al., 1994) by Greatlakean ice. The Calumet level remained stable until the Straits of Mackinac were again deglaciated. After which it dropped several meters to the Toleston/Algonquin level.

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Toleston/Algonquin. The Toleston/Algonquin level, at 184 m, was produced between 11,500 and 11,200 yrs B.P. (Larsen, 1987; Taylor, 1990) as the LML ice margin retreated north during the Two Creekan and opened the Indian River lowland (Figure 2) and the Straits of Mackinac. At that time, water flowed out from the Lake Michigan basin through the Indian River lowland, through the Straits of Mackinac, and out into the Lake Huron basin as the two lakes became confluent. This formed the main phase of glacial Lake Algonquin (Hansel et al., 1985; Larsen, 1985).

Chippewa. After 10,000 yrs B.P. Lake Algonquin levels dropped to the Chippewa Low (140 to 160 m; or below present lake level) as the ice front continued its retreat northward and opened an outlet at North Bay, Ontario (Hough, 1958; 1963; Larsen, 1985; Larsen, 1987). This drop marked the transition from glacial to post-glacial lakes as ice positions ceased to be a factor, and isostatic rebound, climate changes, and outlet elevations became more significant variables during the Early-to-Middle Holocene.

Post-glacial lakes. A series of post-glacial lakes also formed in the Lake Michigan basin. These include:

Nipissing. The Nipissing phase was produced at 8,000 to 4,000 yrs B.P. (Hansel et al., 1985; Larsen, 1985; 1987) with Nipissing I, at 184.5 m, and Nipissing II, at 180.5 m (Larsen, 1985) as Lakes Superior, Michigan and Huron became confluent. During this time, the lake level was becoming progressively less influ

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less influenced by rising spillways to the north (at North Bay) and more controlled by outlets to the south at Chicago and Port Huron (Hough, 1958; Larsen, 1985).

Algoma. The Algoma level was produced at 181.5 m after 4,700 yrs B.P. as the North Bay sill isostatically rose above the Port Huron sill resulting in the downcutting of unconsolidated sediments at Port Huron which caused the lake level to fall (Hansel et al., 1985; Larsen, 1985). This event forever abandoned the bedrock-floored Chicago outlet (Hough, 1958) and forced the cut-off between the Michigan-Huron basins with the Superior basin. Subsequently, the Michigan and Huron basins formed one single lake similar to its present configuration at about 2,500 yrs B.P. (Eschman, 1985).

Arguments on glacial to-post-glacial lake levels in the Sleeping Bear Dunes National Lakeshore region. Leverett (1915) stated that the oldest strandline that he found north of the Manistee moraine represents the Toleston (Algonquin?) level. In addition, many researchers since him agree and it seems to be likely that Glenwood I and II levels should not be present on slopes north of the Manistee moraine because: a) they are likely buried beneath a mantle of Two Rivers drift or were eroded away by advancing Greatlakean ice and; b) the bluffs along the Lake Michigan shoreline are higher than the limit of Glenwood I and II transgression and bluff recession has no doubt eradicated much of the evidence of any paleostrandlines (Evenson, 1973; Farrand and Eschman, 1974; Mickelson and Evenson, 1975).

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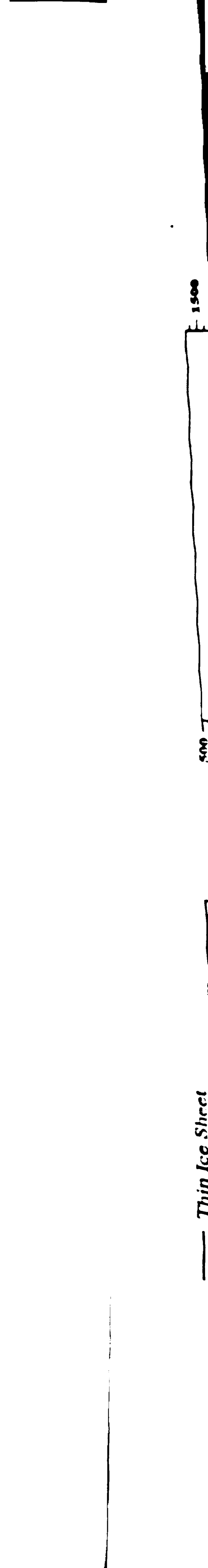
Conversely, Taylor (1990) suggested that Glenwood strandlines survived the ice advance during the Greatlakean in the Lakeshore area and interpreted 225 to 235 m high scarps, terraces, deltas, and other related phenomena developed on some of the ridges of the Manistee moraine around Benzie County as formed by the Glenwood II level of Lake Chicago (Figure 6). Taylor next interpreted 210 to 215 m high scarps, terraces, and deltas as indicative of the Calumet stade which is 21 to 26 m higher than the 189 m horizontal Glenwood stade for the southern Lake Michigan basin demonstrated by Evenson (1973). Using a different approach, Clark et al., (1994) have argued for Glenwood shorelines north of the Manistee moraine and based their argument on computer modeling.

Clark and others (1994) calculated shorelines for Glenwood using a numerical Earth model with varying elastic and viscosity structure. They ran the algorithm based on two simulations (thin vs. thick ice) which took into account the ice-sheet load and subsequent rebound of the land surface as the ice sheet is removed. From this model, they drew predicted Glenwood and Algonquin lake level curves which were compared to historic lake-level gauge data and timing of known outlet activity.

In general, shoreline data from Clark and others (Figures 6-8) for a thick (~2,000 m) ice sheet are probably high estimates, whereas their data for thin (~700 m) ice conforms better to field observations made by Larsen (1985) and Taylor (1990). Clark et al. concluded that the thin-ice model works better when predicting lake level curves, even more so once the ice thickness was increased by 30%. Therefore, using this thin-ice model, the shoreline curve for Glenwood project into the Lakeshore area (~400 km on the transect line as indicated; Figure 6) and predict a Glenwood level there (Figure 6) at about 260 m, while the shoreline curve for Algonquin (Figure 7) at the Lakeshore is predicted at about 191 m (1994).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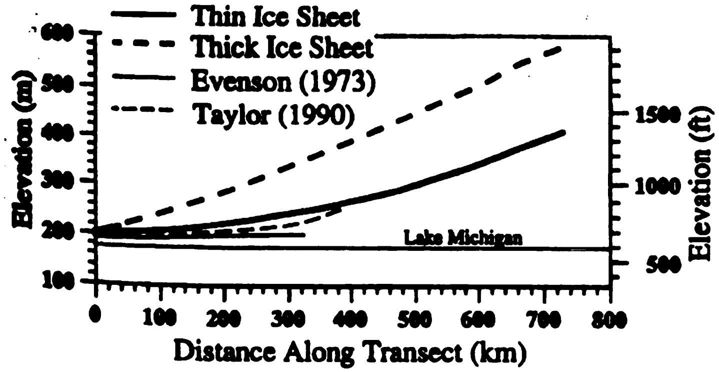


Figure 6. Comparison of Clark et al.'s (1994) predicted Glenwood height/distance profiles to Evenson (1973), and Taylor (1990).

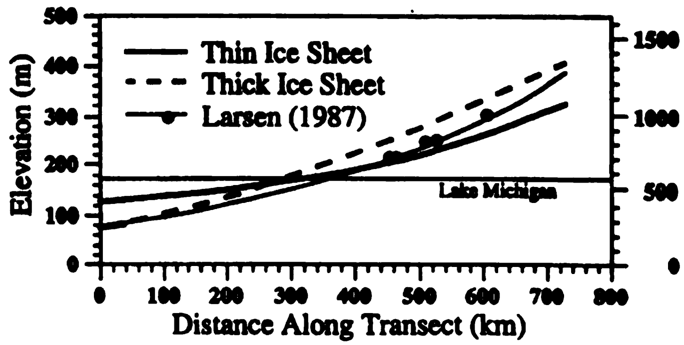


Figure 7. Comparison of Clark et al.'s (1994) predicted Algonquin shorelines to data of Larsen (1987).

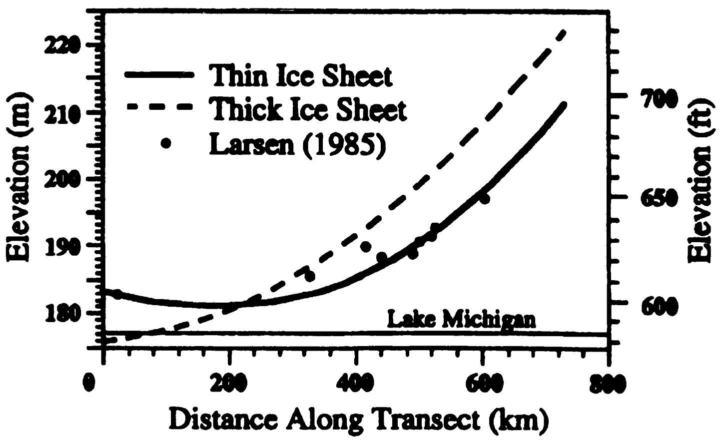
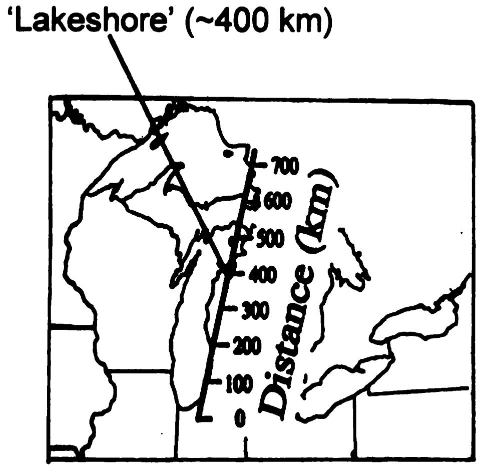


Figure 8. Comparison of Clark et al.'s (1994) predicted Nipissing shorelines to data of Larsen (1987).



The Lakeshore is located at about 400 km on the transect line as shown.

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SECTION II: GEOLOGIC UNITS AND LANDFORMS

Methods

Field mapping.

Field mapping of the Lakeshore geology, supported by funding provided by the Educational Mapping Program (EDMAP), was done over the summers of 1997 and 1998 and employed the most recent U.S. Geological Survey (U.S.G.S.) 7.5 minute topographic quadrangles available (1982–1983). The quadrangles (Figure 9), covering parts of Leelanau and Benzie Counties, are as follows: Glen Haven, Glen Arbor, Good Harbor Bay, Empire, Burdickville, and Beulah. Supplemental adjacent quadrangles covering parts of Grand Traverse and Manistee Counties were also used (Figure 9). They include: North Manitou Island, South Manitou Island, Maple City, Platte River, Lake Ann, and Frankfort. Recent aerial photographs and Leelanau County soil survey maps (1973) and parent material descriptions were also used to assist mapping.

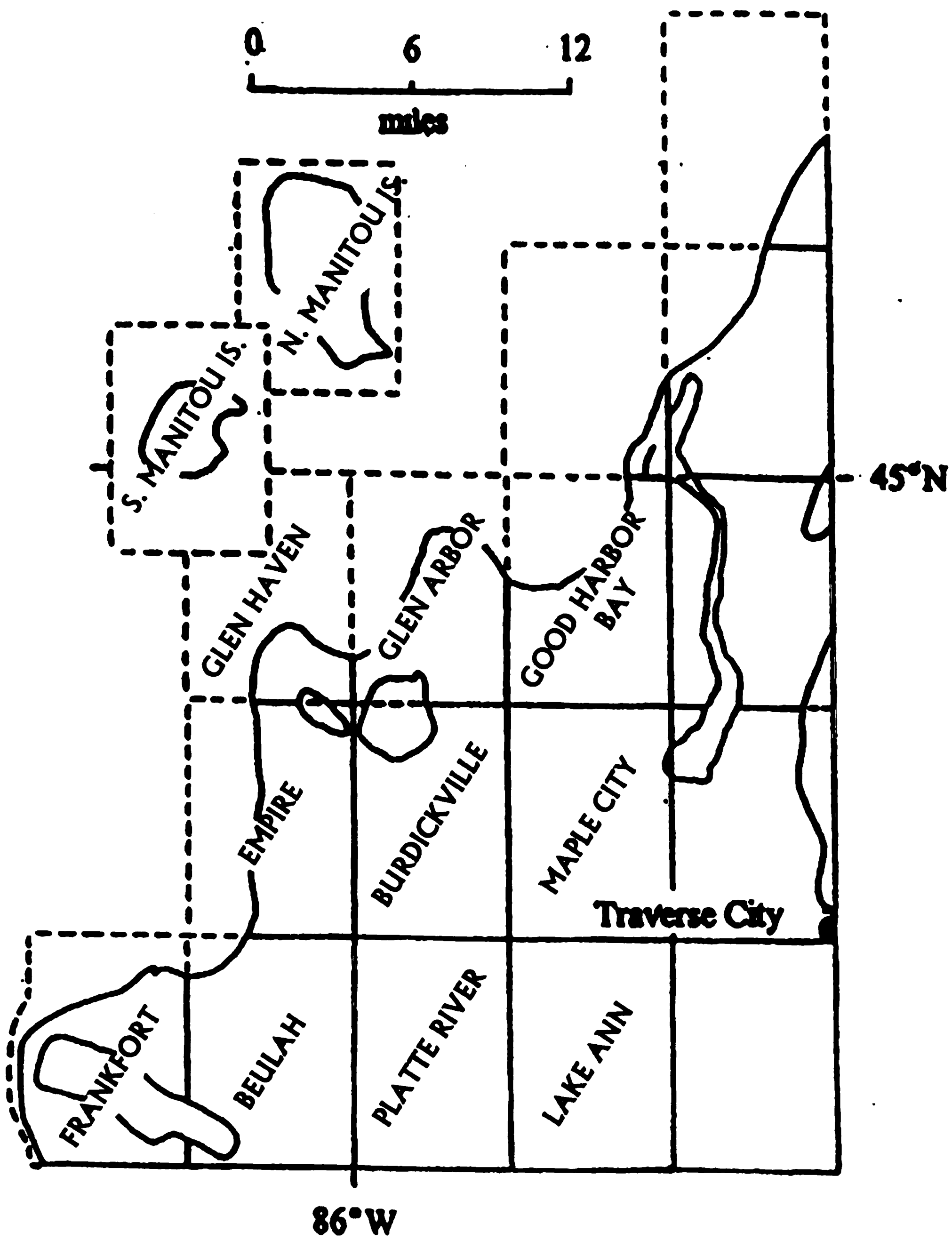

Figure 9. U.S.G.S. 7.5 minute Quadrangles associated with the Lakeshore.

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Surficial deposits.

Surficial deposits were studied in exposures along the Lakeshore, in gravel pits when available, and in shallow (0.25 m to 1 m), hand-dug burrow holes. Special attention was paid to sediment texture and the relative proportions (e.g. "silty-clay"; or, "fine sand with little silt, trace gravel"; and so on), and composition and clast lithology was also noted (granitoid, dolomite, limestone breccia, etc.).

Well-log data.

Deep wells are scarce in the Lakeshore area and their logs provide few details regarding the geologic characteristics of the unconsolidated deposits beneath the surface. Most deep wells are primarily exploratory in nature for oil and gas and in most cases the well logs reflect a general lack of interest by the driller in describing the glacial overburden. Three wells (here termed as 'b' for bedrock wells) in the Lakeshore area were drilled as far as the bedrock and are useful for determining the overall thickness of the unconsolidated sediment overlying bedrock. The locations of these wells (Figure 11) are :

b#1: SW¼,SW¼,NW¼,Sec.35,T.28N.,R.12W.

b#2: SE¼,SE¼,NE¼,Sec.5,T.29N.,R.12W.

b#3: SW¼,SW¼,NW¼,Sec.34,T.28N.,R.14W.

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In addition, four more wells (here termed as 's&g' for sand and gravel wells drilled for water) stopped shy of the bedrock but provide at least minimum thicknesses for the unconsolidated sediments. The locations of these wells (Figure 11) are:

s&g#1: N½,SW¼,SW¼,Sec.4,T.29N.,R.12W.

s&g#2: NW¼,SW¼,SW¼,Sec.16,T.30N.,R.12W.

s&g#3: NE¼,NE¼,SE¼,Sec.33,T.30N.,R.12W.

s&g#4: NE¼,NE¼,NW¼,Sec.30,T.29N.,R.13W.

Other deep well records are available for Leelanau and Benzie counties but lie outside the mapped area. Many of these are municipal water wells and, in some cases, provide lithofacies descriptions and thicknesses of the glacial overburden. These wells were useful to help describe the different surficial geologic units in the Lakeshore area and to help construct cross-sections.

Measurement of sections.

Five lakeshore sections were measured at the Pyramid Point bluff and each section was positioned strategically approximately 350 m apart based mostly on good exposures and consistent distances. Sections were measured using a meter-tape from the top down to the beach and thickness of the measured units was adjusted using a simple trigonometric nomogram that factored in slope angle from the horizontal plane along with the average dip of the beds

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the beds. Paleocurrent measurements were taken of current indicators that include ripples, foresets, cross-beds, and aligned clasts using a Brunton compass where feasible. Lithologic descriptions were made using the methodology (Table 1) outlined in Eyles et al. (1983) which assigns a simple two-to-three-letter alphabetical nomenclature to a unit.

Analysis of geomorphology

Geomorphic analysis consisted mostly of using 2° topographic maps of the Lake Michigan basin to identify ridges around it and the Lakeshore topographic maps to distinguish and describe the main landforms there. These landforms included drumlins, ridges, lowlands, embayments, coastal and perched dunes, as well as erosional landforms such as the bluffs and scarps.

In addition, geomorphology was also used to determine the elevations of former glacial and post-glacial lake-level strandlines at the Lakeshore. A strandline is formed generally during a high lake-level and may be represented by a variety of landforms that are identified mostly from morphological attributes (Larsen, 1985; Larsen, 1987; Taylor, 1990). These landforms included: 1) raised wave-cut scarps in areas where no fluvial activity occurred, 2) deltas, terraces, benches, curvilinear beach and dune deposits, 3) large coastal swamps and lacustrine plains, and 4) well-developed wave-cut scarps and benches (Leverett and Taylor, 1915; Bretz, 1951; Evenson, 1973; Taylor, 1990; Colman et al., 1994). Once these landforms were identified and mapped, the general elevations of their surfaces were read directly from the contours on the topograp

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topographic maps. These elevations are therefore representative of lake-level strandlines.

Due to the fact that strandline elevations were taken mostly from topographic maps with a 5 m contour interval, exact elevations for former lake levels were difficult to determine. Therefore, the paleolake strand data given in this thesis are only estimates.

<u>Sampling Strategy.</u> Strandline data (elevations) were assembled, using the criteria described above, from twenty-five sites (with over 70 data points; Table 3) to cover about 130 km of Lake Michigan shoreline from Leland southwest to Elberta, with North and South Manitou Islands included. These elevations, as stated earlier, were read mostly from the five-meter contours on the topographic maps used in this study and were then plotted on a scatterplot graph to show their general spatial distributions. Therefore, the margin-of-error for any strandline elevation using this approach is ± 5 m. Lines were then drawn that best-fit reasonable plot groups. Group criteria were based on finding the most number of points that fell on a line drawn roughly horizontally. This line therefore suggests an average strandline elevation for a paleolake level. This simple group method was done using the general distribution of points that best represented paleolake elevations. A point distribution of 6 points or more that fell on or around the same line would suggest a well-defined strand, 5 to 3 points would be considered a poorly defined strand, and less than 3 points on a line would suggest a possible strand.

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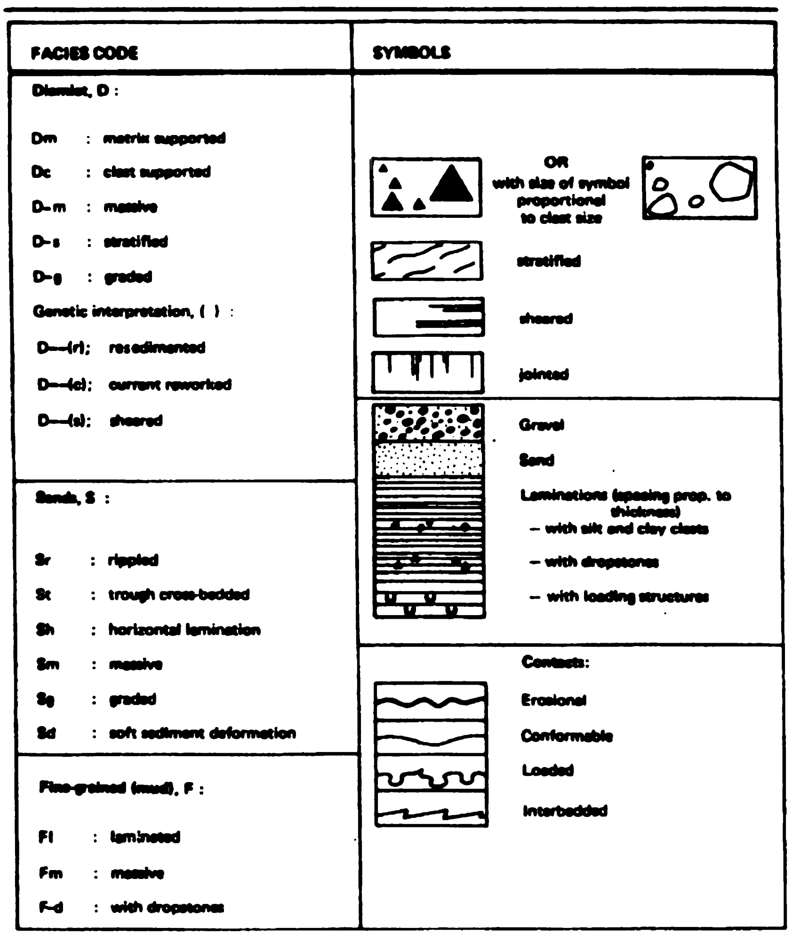
Table 1. Glacial lithofacies codification (adapted from Eyles et al., 1983).

LITHOFACIES

DIAMICT D		SAND	S
Dmm	matrix-supported, massive	Sm	massive
Dms	" " , stratified	Sr	rippled cross-stratified
Dmg	" " , normally-graded	Sh	horizontally laminated
Dcs	clast-supported, stratified	St	trough cross bedded
MUD	F	GRAVEL	G
Fm	massive	Gm	massive
Fl	laminated	Gms	massive, sand supported
Fld	laminated silty-clay/dropstones	Gc	clast-supported

Table 2. Diamict lithofacies code and symbols in addition to those in Table 1 (from Eyles et al., 1983).



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Map assembly

Final paper copies of the completed maps were transferred to mylar and hand-colored with all pertinent metadata and unit descriptions added. Three cross-sections (A-A', B-B', C-C') across representative areas were also constructed and included on the map. Finally, the completed map was digitized into an Arc/Info–ArcView™ (Version 3.1) format and was then overlain with the most recent Digital Elevation Models (DEM) available from the U.S.G.S..

A color copy of the digitized surficial geologic map of the Lakeshore is shown as Figure 12. Other images in this thesis are also presented in color. In addition, 7.5" quadrangle-sized copies of the map are available (1:25,000 scale) as an open-file report (EDMAP#1434-HQ-97-AG-01730) at the Department of Environmental Quality (DEQ), Michigan Geological Survey, Lansing, Michigan.

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Bedrock geology

No bedrock crops out in the Lakeshore region, however rock units are exposed to the east and northeast around Petosky, Charlevoix and Mackinac Island. The bedrock geology of the Lakeshore, shown in Figure 10, is based on Milstein's (1987) bedrock geology map of Southern Michigan published at a scale of 1:500,000. The map shows that the area is underlain mostly by northeast striking Mississippian to Devonian sedimentary rocks (Melhorn, 1954; Martin, 1957; Milstein, 1987) that dip toward the southeast at about 5°. The bedrock includes, from oldest to youngest, the Detroit-Traverse Group and the Antrim-Ellsworth-Coldwater (shales) Group (Milstein, 1987).

The Detroit-Traverse Group is mostly Devonian in age and consists of white to black dolomite and limestone. Pebbles, cobbles, and boulders from this group are common within the glacial deposits around the Lakeshore and are generally prominent along most of the shoreline, especially areas with high bluffs.

The Antrim-Ellsworth-Coldwater Group is Devonian to Mississippian in age, overlies the Traverse Group, and consists mostly of light-brown to black shale with nodules of limestone and pyrite cubes. The Antrim shale is thin-bedded, fissile, and dark-brown-to-black while the Ellsworth and Coldwater shales consist of mostly soft, semi-indurated green shale with bands of calcareous material (Melhorn, 1954; Martin, 1957). In general, rocks from this group are scarce within glacial deposits at the Lakeshore.

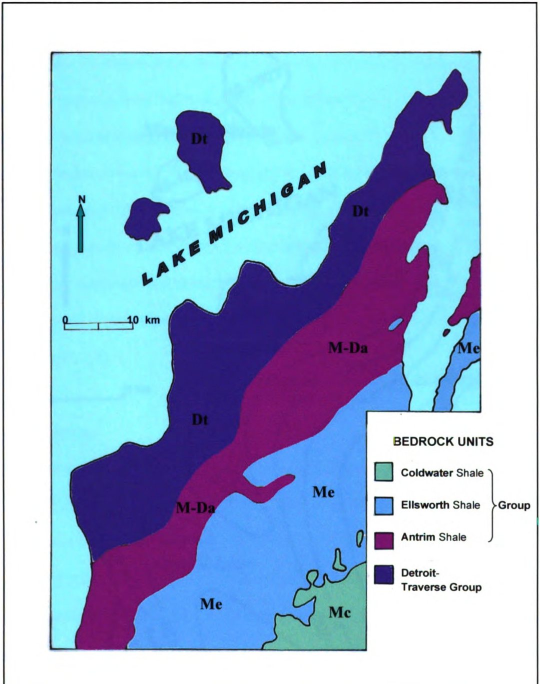


Figure 10. Map of bedrock units at the Lakeshore and vicinity (from Milstein et al., 1987).

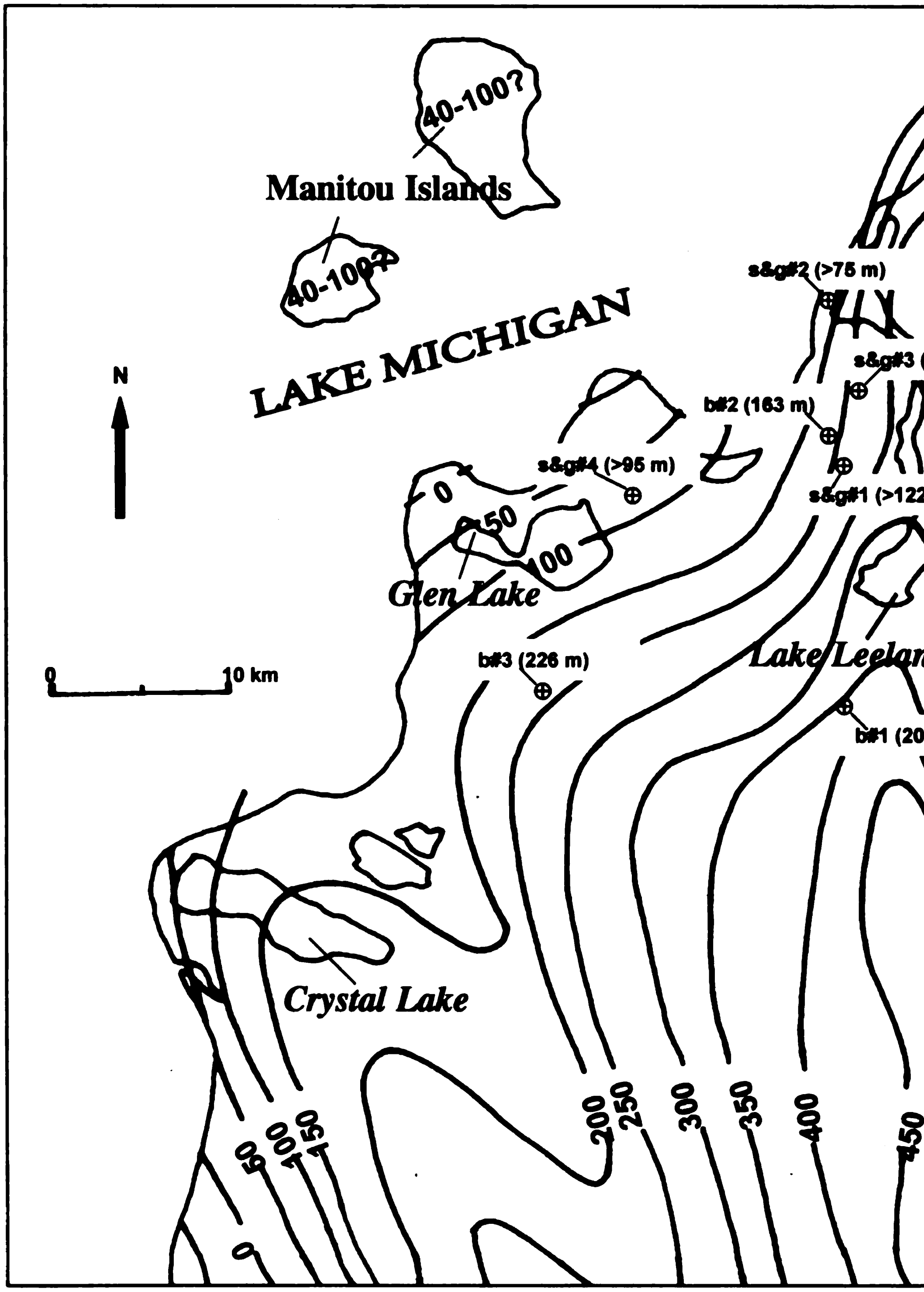


Figure 11. Map showing elevation of bedrock surface and well locations with spot thicknesses of glacial drift (in meters a.s.l.) at the Lakeshore and vicinity (from the Hydrologic Atlas of Michigan, Geologic Survey Division, Michigan Department of Natural Resources, 1981).

Drift thickness. The elevation of the bedrock at the Lakeshore is generally not well known and therefore only general estimates on drift thickness are possible.

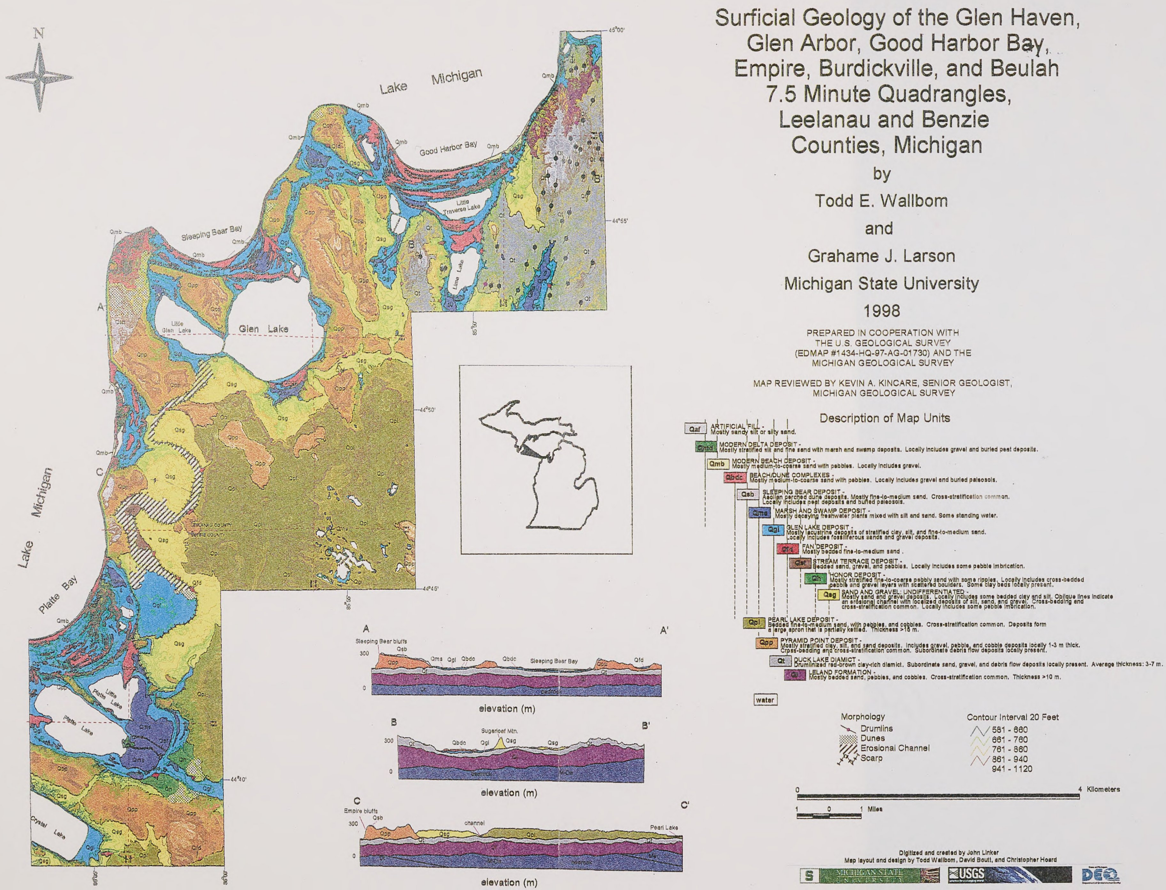
Figure 11 shows the bedrock surface elevation for the Lakeshore and vicinity as redrawn from the Hydrologic Atlas of Michigan (1981). Drift thickness, calculated by subtracting the elevations of the land surface taken off the U.S.G.S. 7.5" quadrangle maps from elevations of the bedrock surface reported in the atlas, ranges from 50 to 322 m and thin towards the east and northeast. Drift thickness ranges from 60 to 180 m at the Manitou Islands (Martin, 1957). It is thickest near the mainland shoreline where it is about 230 m (well b#3) and thins towards the east near Lake Leelanau where it ranges from 75 to 200 m (wells b#1,2; s&g#1,2,3).

Surficial geology

Fourteen surficial geologic units were identified and mapped at the Lakeshore (Figure 12) and are described (oldest to youngest) as follows (all names given in this description are informal):

Leland formation. The "Leland formation" is a 5 to 120 m thick continuous layer directly overlying the bedrock. It consists predominantly of massive, planar-to-cross bedded, moderately sorted, pebbly-to-cobbly, medium-to-coarse sand. The sand facies is mostly composed of rounded-to-subrounded quartz whereas the pebbles and cobbles are composed chiefly of subangular-to-subrounded carbonates and some crystallines. Cross-stratification within the sand is common and includes cross-sets and truncated beds. Also common are

Figure 12. Surficial Geology of the Glen Haven, Glen Arbor, Good Harbor Bay, Empire, Burdickville, and Beulah 7.5 minute Quadrangles, Leelanau and Benzie Counties, Michigan (Wallbom and Larson, 1998).



Surficial Geology of the Glen Haven, Glen Arbor, Good Harbor Bay, Empire, Burdickville, and Beulah 7.5 Minute Quadrangles, Leelanau and Benzie Counties, Michigan
by
Todd E. Wallbom
and
Grahame J. Larson
Michigan State University
1998
PREPARED IN COOPERATION WITH THE U.S. GEOLOGICAL SURVEY (EDMAP #1434-HQ-97-AG-01730) AND THE MICHIGAN GEOLOGICAL SURVEY
MAP REVIEWED BY KEVIN A. KINCARE, SENIOR GEOLOGIST, MICHIGAN GEOLOGICAL SURVEY
Description of Map Units
Qaf ARTIFICIAL FILL - Mostly sandy silt or silty sand.
Qmd MODERN DELTA DEPOSIT - Mostly stratified silt and fine sand with marsh and swamp deposits. Locally includes gravel and buried peat deposits.
Qmb MODERN BEACH DEPOSIT - Mostly medium-to-coarse sand with pebbles. Locally includes gravel.
Qbdc BEACH/DUNE COMPLEXES - Mostly medium-to-coarse sand with pebbles. Locally includes gravel and buried paleosols.
Qsb SLEEPING BEAR DEPOSIT - Aeolian perched dune deposits. Mostly fine-to-medium sand. Cross-stratification common. Locally includes peat deposits and buried paleosols.
Qms MARSH AND SWAMP DEPOSIT - Mostly decaying freshwater plants mixed with silt and sand. Some standing water.
Qgl GLEN LAKE DEPOSIT - Mostly lacustrine deposits of stratified clay, silt, and fine-to-medium sand. Locally includes fossiliferous sands and gravel deposits.
Qfd FAN DEPOSIT - Mostly bedded fine-to-medium sand .
Qst STREAM TERRACE DEPOSIT - Bedded sand, gravel, and pebbles. Locally includes some pebble imbrication.
Qh HONOR DEPOSIT - Mostly stratified fine-to-coarse pebbly sand with some ripples. Locally includes cross-bedded pebble and gravel layers with scattered boulders. Some clay beds locally present.
Qsg SAND AND GRAVEL: UNDIFFERENTIATED - Mostly sand and gravel deposits. Locally includes some bedded clay and silt. Oblique lines indicate an erosional channel with localized deposits of silt, sand, and gravel. Cross-bedding and cross-stratification common. Locally includes some pebble imbrication.
Qpl PEARL LAKE DEPOSIT - Bedded fine-to-medium sand, with pebbles, and cobbles. Cross-stratification common. Deposits form a large apron that is partially kettled. Thickness >15 m.
Qpp PYRAMID POINT DEPOSIT - Mostly stratified clay, silt, and sand deposits. Includes gravel, pebble, and cobble deposits locally 1-3 m thick. Cross-bedding and cross-stratification common. Subordinate debris flow deposits locally present.
Qt DUCK LAKE DIAMICT - Druminized red-brown clay-rich diamict. Subordinate sand, gravel, and debris flow deposits locally present. Average thickness: 3-7 m.
Ql LELAND FORMATION - Mostly bedded sand, pebbles, and cobbles. Cross-stratification common. Thickness >10 m.
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761 - 860
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Map layout and design by Todd Wallbom, David Boutt, and Christopher Hoard
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Little Traverse Lake
Lime Lake
Little Glen Lake
Glen Lake
Platte Bay
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Platte Lake
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44°50'
44°45'
44°40'
A
A'
Sleeping Bear bluffs
elevation (m)
B
B'
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Empire bluffs
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imbricated pebbles and cobbles as well as coarsening-upward sequences. Individual bed thickness ranges from 0.1 to 1.5 m. Other characteristics include: infrequent, gray-to-red, thin clay and silt clasts 1 to 4 cm in diameter; frequent, 1 to 5 cm thick and 8 to 22 cm long, gray-to-light tan silt and sand lenses; frequent, 1 to 14 cm thick and 4 to 60 cm long gravel lenses; and infrequent boulders 15 to 40 cm in diameter.

The Leland formation is also characterized by 1 to 3 cm thick and 1 to 2 m long units of massive-to-laminated silt and fine sand facies as well as rippled sand and 2 to 5 cm thick and 0.6 to 2 m long, gray-blue, wavy, laminated muds. In addition, the formation includes 0.05 to 1.0 m thick and 0.7 to 3 m long units of well-sorted sand, pebbles, and cobbles.

Well logs (b#2; s&g#3) show that the formation also includes massive, stony, reddish diamict layers, massive-to-laminated, reddish-brown clay and silt layers, and stratified, thinly-bedded sand and gravel layers that includes several interbedded "yellow" sand stringers.

There are two good exposures of Leland formation in the Lakeshore area. One is at the Leelanau County gravel pit (Figure 13) located about 3 km south of Duck Lake (NE¼,Sec.32,T.30N,R.12W.). The other is a 2.2 km long bluff (Figure 14) located at the northwest end of Pyramid Point about 6 km northeast of Glen Arbor (NE¼,Sec.36,T.30N,R.14W.). At the pit, Duck Lake diamict overlies about 10 m of the sand of the Leland formation (Figures 13, 15). In some places, northward dipping 0.5 to 2 cm thick and 2 to 6 cm long masses of Duck Lake diamict are thrust downward into the sand. The long axes of rock clasts within

Figure 13. Stratified sand and gravel of the Leland formation overlain by Duck Lake diamict at the Leelanau County gravel pit.

Figure 14. Laminated clay and silt of the Leland formation with some small-scale normal faulting at Pyramid Point. Note deformed beds near pencil.

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the unit have a general north-south orientation. At Pyramid Point, 5 to 12 m of the Leland formation is exposed and directly underlies the Duck Lake diamict (Figure 14). Here, it consists mostly of a blue-to-gray, laminated clay and silt layer which grades upwards to a more massive, blocky, and chaotic clayey-silt and fine sand layer and then into a pebbly, clayey-silt layer. The laminated clay and silt layer has frequent current ripples often in the form of ripple trains and is locally intercalated with 0.2 to 5 cm thick and 6 to 45 cm long ribbons of sand. The clayey-silt and fine sand layer includes frequent soft-sediment deformation structures such as contorted bedding (Figure 14) and frequent, 0.4 to 2 cm thick and 2 to 6 cm long, diamitic lenses that are generally planar. Also frequent here are 0.3 to 5 cm thick and 5 to 18 cm long rock clast horizons.

Duck Lake diamict. The "Duck Lake diamict" is a massive, red-to-reddish brown, poorly sorted, matrix-supported diamict. The diamict matrix consists chiefly of fine, rounded-to-subrounded sand composed mainly of quartz and some fine-grained sediments. The proportions of fine-grained sediments include mostly silt (55-75% of the total amount of fine-grained sediments within the diamict matrix) and some clay (25-45%). Clasts within the diamict are common, subrounded, 0.5 to 30 cm in diameter, and consist mostly of carbonates, some crystallines, and pink-to-red sandstones. Other characteristics of the diamict include: frequent, 0.3 to 6 cm thick and 2 to 38 cm long, gray-to-light-tan, fissile, silt stringers; frequent, 0.1 to 12 cm thick and 5 to 100 cm long, light-tan, fine-to-medium, sand stringers; and infrequent, 3 to 19 cm thick and 10 to 40 cm long pebble and cobble lenses.

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Locally, the diamict is more red than brown, sandier, and includes a greater frequency of rock clasts but in some areas it is mostly light-gray-to-light-blue. In a few places, it also includes poorly developed, discontinuous, rock-clast horizons.

The Duck Lake diamict extent is about 96 km^2 and is exposed mostly in the eastern uplands where it underlies an undulating, drumlinized surface. Well logs (b#2; s&g#1,2,3,4) and exposures of the unit show that its thickness generally ranges from 2 to 35 m.

There are two good exposures of Duck Lake diamict in the Lakeshore area. One is at the Leelanau County gravel pit whereas the other is at the northwest end of Pyramid Point. In the pit, about 1 to 5 m of the unit is exposed and directly overlies the Leland formation (Figure 15). The contact between it and the Leland formation is sharp and unconformable. Also, in some places weakly developed rock clast horizons are present near the contact. Above the contact, 1 to 2.5 m of massive, brown-red, silty-clay diamict is overlain by a more red, sandier, and more stratified diamict. The stratified diamict is 0.25 to 2 m thick and occurs above a thrust-induced contact with the core diamict. Locally above this contact and concentrated within the stratified diamict lies a zone of 0.5 to 2.3 m long thrust faults that dip northward at 3 to 7° and are spaced 0.5 to 1.2 m apart.

At Pyramid Point, a 2 to 11 cm thick and 0.3 to 1.6 m long bed of red clayey-silt diamict overlies the Leland formation (Figure 16). Within this diamict are infrequent, 1 to 3 cm thick and 2 to 5 cm long, blue-gray-to-red-brown, diamic

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diamict clasts as well as frequent pebbles and cobbles that consist of mostly carbonates and crystallines. The diamict shows increasing degrees of deformation approaching the contact with the overlying sands and gravels of the Pyramid Point deposit.

Pyramid Point deposit. The "Pyramid Point deposit" consists of stratified, planar-to-crossbedded, poor-to-moderately sorted, pebbly-to-cobbly, fine-to-medium, rippled sand and includes light-blue-to-red-to-reddish-brown, massive-to-laminated silt and clay layers. The sand consists mostly of rounded-to-subrounded quartz whereas the pebbles and cobbles are chiefly subangular-to-subrounded carbonates and some crystallines. Cross-bedded pebble and gravel layers are common and include cross-sets, frequent troughs, truncated beds, ripples, and coarsening-upward sequences. Thicknesses for the sand and sand and gravel beds range from 0.2 to 3 m. Thicknesses for the silt and clay layers range from 0.01 to 2.5 m. Other characteristics of the deposit are: infrequent, 1 to 3 cm wide and 3 to 8 cm long, blue-gray-to-red, thin silt and clay clasts and lenses; infrequent, 0.8 to 6 cm thick and 4 to 26 cm long, light-tan, fine, sand lenses; and infrequent boulders 0.2 to 1.3 m in diameter.

In some areas, the deposit contains more silt, clay, and fine sand as well as wavy laminated muds. In other places, the deposit is more pebbly and cobbly and includes 0.4 to 1.0 m thick and 0.1 to 4 m long beds of pebbles and cobbles with a coarse sand matrix. Locally, the deposit contains aligned pebbles and cobbles within a coarse sand matrix. In addition, some of these aligned, matrix-supported pebbles and cobbles are contained within diapiric-shaped horizons

Figure 15. Detail of massive Duck Lake diamict superimposed on Leland formation deposits, Leelanau County gravel pit. Note quarter on contact (arrow).

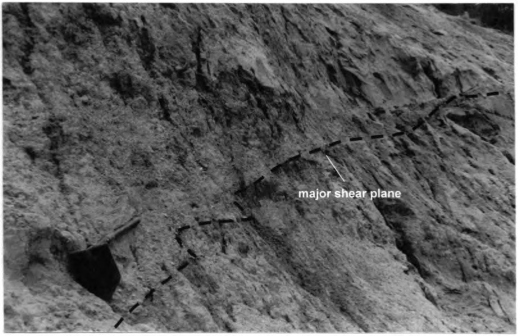

Figure 16. View upsection of a major shear plane (dashed line) in deformed Duck Lake diamict shown slightly offset by post depositional slumping at Pyramid Point. Shovel for scale is about 1 m in length.

Figure 17. View of the bluff at Pyramid Point from Lake Michigan. Direction of view is to the south.

Figure 18. Closer view of glaciolacustrine sediments at Pyramid Point. The general direction of flow was from left to right. Shovel is about 1 m in length.

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that are about 1 m wide upwards and taper to a point downwards. These horizons are infrequent, occur within mostly deformed areas of the unit, range from 0.6 to 2 m long, and generally coarsen upwards.

The Pyramid Point deposit is associated with the high ridges (Figures 5, 12, 17) that occur within the Lakeshore. The dimensions of these ridges are 0.3 to 1 km wide, 50-120 m high, and 3-12 km long and the orientation generally trends northwest to southeast. A well log (s&g#4) and frequent exposures at the Lakeshore show that its thickness ranges from 45 to 120 m.

The best exposure of the Pyramid Point deposit is the bluff at Pyramid Point (Figures 17, 18). There, about 56 to 120 m of the deposit is exposed and directly overlies Duck Lake diamict (Figure 19). Its contact with the diamict is unconformable and in general dips to the southeast at about 20 to 35°.

Primary structural features at Pyramid Point include trough cross-beds, rippling, erosional contacts and scoured bases, small-scale (<2 m) intraformational faulted blocks, conjugate faulting, and jointing. Mean direction of sediment transport associated with ripples, cross beds (Figure 18), and aligned rock and diamict clasts is from north to south.

Pearl Lake deposit. The "Pearl Lake deposit" consists of thick-bedded, planar-to-cross bedded, moderate-to-well sorted, pebbly-to-cobbly, clast-supported gravel with medium-to-coarse sand beds. The gravel is pebble and cobble-rich and consists of mostly subangular-to-rounded carbonates, crystallines, and infrequent pink-to-red sandstones. The sand consists mostly of rounded-to-subrounded quartz. Cross-stratification is common as are cross-bedding, truncated beds, imbrica

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imbricated pebbles and cobbles, and fining-upward sequences. Other characteristics include: frequent, 0.4 to 2 m thick and 1.3 to 2.5 m long, planar bedding with discontinuous rock clast horizons; infrequent, weakly developed troughs and channel deposits; frequent, weakly-developed clast imbrication; infrequent, 0.2 to 4 cm thick and 3 to 25 cm long silt and fine sand beds; and frequent, subangular-to-subrounded boulders 15 to 45 cm in diameter.

In some areas, the deposit contains more well-sorted fine-to-medium sand layers 2 to 15 cm thick and 0.5 to 2 m long and whitish fine sand lenses 1 to 3 cm thick and 3 to 8 cm long. In other places, the deposit is more cobbly and contains frequent boulders whereas other areas has fewer cobbles and fewer boulders with an overall increase in coarse sand layers that range from 0.15 to 1.8 m thick.

The Pearl Lake deposit extent is about 130 km^2 and is associated with a high flat plain that slopes to the southeast at about 6 m/km, or 0.6%. Where the deposit is partially collapsed, elongate northwest-southeast trending lakes mostly fill the lows. The highest elevation of the deposit occurs near Bow Lake at 295 m and the lowest elevation occurs to the south around Pearl Lake at about 265 m. A well log (b#3) and frequent exposures at the Lakeshore show that it's thickness ranges from 60 to 105 m.

The type section for the Pearl Lake deposit is located at the Kasson Sand and Gravel Mine, (NW¼,Sec.16,T.28N,R.13W; Figures 20, 21), a large active gravel pit located about 2 km south of Bow Lake. At the pit, about 17 m of the

Figure 19. Stratigraphic sections at Pyramid Point: A-E. Refer to Figure 28 for facies descriptions.

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SAND

LITHOFACIES

DIAMICT	D	**SAND**	S
Dmm	matrix-supported, massive	Sm	massive
Dms	" " , stratified	Sr	rippled cross-stratified
Dmg	" " , normally-graded	Sh	horizontally laminated
Dcs	clast-supported, stratified	St	trough cross bedded
MUD	F	**GRAVEL**	G
Fm	massive	Gm	massive
Fl	laminated	Gms	massive, sand supported
Fld	laminated silty-clay/dropstones	Gc	clast-supported

(modified from Eyles et al., 1983)

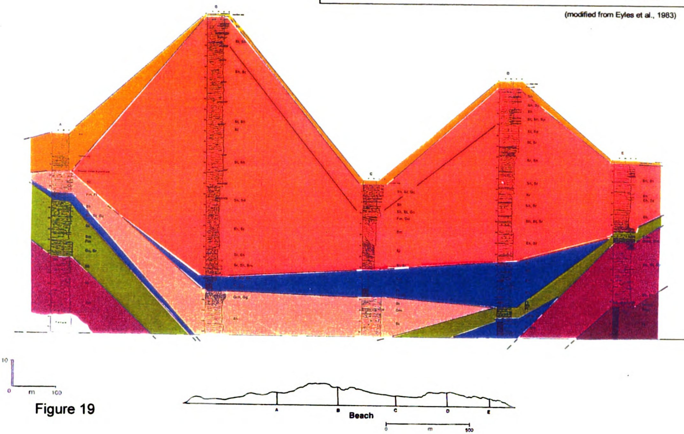


Figure 19

deposit is exposed and shows a general fining-upwards as well as imbricated rock clasts (Figure 21) with long axes that dip towards the north-northwest.

Sand and gravel-undifferentiated. This deposit consists of weakly-stratified, planar-to-crossbedded, poor-to-moderately sorted, pebbly-to-cobbly, matrix-supported, medium-to-coarse sand and gravel. The sand consists mostly of rounded-to-subrounded quartz whereas the gravel consists mostly of subangular-to-subrounded carbonates and some crystalline and pink-to-red sandstones. The cross-stratification and cross-bedding often includes foresets. Also common are truncated beds, imbricated pebbles and cobbles, and fining-upward sequences. Individual bed thickness ranges from 1.0 to 35 cm. Other characteristics are: infrequent, laminated, clay layers 0.1 to 2 cm thick; infrequent, 0.4 to 3 cm thick and 2 to 7 cm long, gray-to-red clay and silt lenses; infrequent, 0.2 to 1 cm thick and 3 to 10 cm long, gray-to-light tan silt and sand lenses; frequent, 1 to 4 cm thick, massive-to-laminated, silt beds; frequent, massive, fine-to-medium sand; frequent, 1 to 7 cm thick and 4 to 30 cm long, gravel lenses; and infrequent boulders 15 to 40 cm in diameter.

In some areas, the undifferentiated sand and gravel deposit contains more fine sand and includes 1 to 3 cm thick and 0.5 to 2 m long, rippled sand and wavy, laminated muds. In other places, the deposit is more pebbly and cobbly and includes beds of well-sorted sand, pebbles, and cobbles 1 to 15 cm thick. Frequent exposures at the Lakeshore show that it is mostly underlain by the Duck Lake diamict, the Pyramid Point and the Pearl Lake deposits, and is 5 to 29 m thick.

Figure 20. Thick (~2 m) sets of stratified, coarse-grained Pearl Lake glaciofluvial sediments at the Kasson Sand and Gravel Mine, Leelanau County. The view is to the north.

Figure 21. Closer view of Pearl Lake glaciofluvial sediments at the Kasson Sand and Gravel Mine. Well-developed clast imbrication indicates that the general current flow was southward.

The type section for the undifferentiated sand and gravel deposit is at the NPS gravel pit located just off Indian Hill Road about 5.5 km south of Empire (NE¼,Sec.7,T.27N,R.14W.). There, about 1.5 m of the deposit is exposed and consists mostly of stratified, coarse sand with gravel, pebbles, and cobbles. Also present are laminated silt and fine sand beds 0.3 to 4 cm thick, as well as clast-supported pebbles and cobbles 5 to 30 cm thick. The long axes of imbricated rock clasts within the deposit dip to the northwest.

Honor deposit. The "Honor deposit" consists of thin-to-thick bedded, planar-to-cross-stratified, poor-to-moderately sorted, pebbly-to-cobbly, matrix-supported, fine-to-medium sand. The sand consists chiefly of rounded-to-subrounded quartz whereas the pebbles and cobbles consists of mostly carbonates, some crystallines, and some shales and mudstones. Cross-bedded pebble and gravel layers are common and include cross-sets, truncated beds, and fining-upward sequences. In some locations, moderately-dipping (≤20°) foresets are common. Other characteristics include: infrequent, 1 to 4 cm thick and 3 to 8 cm long, whitish-tan, well-sorted, fine sand lenses; frequent, 0.3 to 5 cm thick and 0.4 to 2 m long, tan, laminated, silt beds; frequent, 0.3 to 5 cm thick and 2 to 12 cm long, tan, silt lenses; and frequent, 6 to 25 cm thick and 0.4 to 3 m long, rock clast horizons.

The Honor deposit in the Lakeshore area is associated with flat surfaces with an areal extent of about 1 to 2.2 km^2 and is found near to or within lowlands and large stream channels. Elevations of these surfaces vary but most fall within a range of 210 to 235 m. Deposit thickness generally ranges from 10 to 40 m.

Figure 22. The gravel pit of the lower delta northwest of Honor, Benzie County.

Figure 23. Detail of laminated sediments in the Honor delta complex. Pencil is about 10 cm in length.

The best exposure of the Honor deposit is located in the Benzie County Road Commission gravel pit (NE¼,Sec.8,T.26N.,R.14W.; Figures 22, 23) located near the town of Honor in Benzie County. There, about 8 m of the deposit is exposed and consists of mostly stratified sand and gravel (Figure 22). Most of the bedding is planar or has a low-to-moderate (5-20°) southwesterly dip. In some areas, the deposit consists of more finely laminated silt and fine sand (Figure 23) with frequent 1 to 3 cm thick and 2 to 7 cm long, white to-tan, fine sand lenses.

Stream terrace deposit. This deposit consists chiefly of thin-to-thick bedded, poorly-sorted, pebbly-to-cobbly, matrix-supported gravel. The gravel is pebble and cobble-rich and consists mostly of subangular-to-subrounded carbonates and some crystallines whereas the matrix consists chiefly of medium-to-coarse sand that is mostly subangular-to-subrounded quartz. Cross-bedded gravel layers are common and sometimes include troughs, foresets, and imbricated pebbles and cobbles. Other characteristics are: infrequent, 0.2 to 3 cm thick and 0.5 to12 cm long silt and fine sand lenses; frequent, 4 to 25 cm thick and 2 to 35 cm long, massive, fine-to-medium sand; frequent, 2 to 13 cm thick and 5 to 40 cm long, gravel lenses; and infrequent boulders 15 to 40 cm in diameter.

The stream terrace deposit is associated mostly with a pair of terraces in an abandoned channel located east of Empire. A photograph (Figure 24) taken from Norcronk Road shows one of the well-developed terraces that has a surface elevation of 225 m (a.m.s.l.), or an elevation of about 5 m above the channel floor. Deposit thickness generally ranges from 3 to 7 m.

The only exposure of the deposit occurs at the NPS gravel pit (inactive) located near M-22 just south of Empire in Benzie County about 300 m east of the intersection of Aral Hills and Aral Road (NW½,Sec.6,T.27N.,R.14W.). At the pit, about 2 m of the deposit is exposed. The long axis of imbricated rock clasts within the deposit dip to the north-northwest.

Glen Lake deposit. The "Glen Lake deposit" consists chiefly of stratified, moderate-to-well sorted, fine sand with beds of red-to-red brown, laminated silt and clay. The sand consists mostly of rounded-to-subrounded quartz and the thickness of individual sand beds range from 0.4 to 27 cm. The silt and clay beds range from 0.1 to 3 cm thick and 0.3 to 4 m long. Other characteristics include: frequent, 0.2 to 6 cm thick and 0.5 to 1 m long, rock clast horizons; infrequent, 0.5 to 1.2 m thick and 2 to 8 m long, massive-to-weakly stratified, blue-gray, fossiliferous, marl beds; and frequent, 0.3 to 1.7 m thick and 1.2 to 12 m long, tabular, rippled, fine-to-medium sands.

In some areas, the deposit includes fossiliferous sands and marl with frequent mollusc shell fragments. In other areas, the deposit has more fine sand than clasts, whereas locally it is more gravel and pebble-rich. A well log (s&g#2) and frequent exposures show that the deposit ranges from 0.04 to 8 m thick.

The Glen Lake deposit is extensive and is usually associated with low relief areas at the Lakeshore. The best exposure of the deposit is located at the easternmost end of Pyramid Point bluff (NW¼,Sec.32,T.30N.,R.13W.; Figures 17, 19) near Hidden Lake. Here, about 1.2 m of the deposit is exposed and shows mostly 1.2 m thick and 6 m long, blue, marl beds overlying stratified, red-brown, clay beds that range from 15 to 50 cm thick. Overlying the marl are massive-to-parallely-laminated, mollusc-bearing, cross-bedded sands about 4 m thick.

Sleeping Bear deposit. The "Sleeping Bear deposit" consists of mostly light-tan-to-orange, massive-to-laminated, well-sorted, fine sand. Cross-stratification is common as are truncated beds. Other characteristics are: frequent, planar-bedding; infrequent, 0.5 to 3 cm thick and 0.3 to 2 m long, clast horizons; infrequent, 0.2 to 7 cm thick, dark-brown-to-black, continuous, organic-rich layers, infrequent, 0.1 to 3 cm thick and 2 to 6 cm long, organic-rich lenses, and infrequent, 3 to 15 cm thick zones of disturbance and mixing by plants and animals. Overall thickness of the unit ranges from 2 to 40 m.

The best exposure of the deposit is the Sleeping Bear Dune, which is part of the Sleeping Bear complex (Figure 25) located near Sleeping Bear Point. Here, about 15 to 32 m of the deposit is exposed and consists of mostly massive, fine sand floored by pebble and cobble-rich sand. In other areas, the deposit is superimposed on continuous, organic-rich layers 0.3 to 5 cm thick as well as rootlets, tree, plant, and animal remains.

Beach/dune complex. This deposit consists of tan-to-reddish brown, massive-to-laminated, moderate-to-well sorted, fine-to-medium, sand deposits with gravel and pebble layers. The sand consists chiefly of subrounded-to-subangular, quartz sand whereas the gravel and pebbles are mostly carbonates and crystallines. Primary structures range from horizontal parallel laminations to low-angle and high-angle parallel laminations. Other characteristics include:

Figure 24. A ~5-meter high terrace in an abandoned channel near Empire. Direction of flow was from left to right. View is towards the southwest.

Figure 25. A large blowout surrounded by stabilized dunes perched atop Sleeping Bear bluffs. Note possible Algonquin-age terrace (arrow) developed near the southern end of South Manitou Island.

frequent, 0.4 to 6 cm thick and 0.2 to 1 m long gravel and pebble horizons; frequent, coarse, sand beds 5 to 20 cm thick; frequent, 0.2 to 3 cm thick and 4 to 25 cm long, reddish-brown-to-black laminae of heavy minerals; and infrequent, 0.4 to 3 cm thick and 0.6 to 5 m long, dark-brown-to-black, organic-rich layers mixed with fine sand.

Beach/dune deposits are associated with distinctive, arcuate, ridge-like landforms (Figure 12) 1.3 to 20 m high and 0.2 to 5 km long and are generally located within enclosed semi-protected embayments. The deposit oftentimes is superimposed on Glen Lake deposits but may also be found superimposed on the Pyramid Point deposit and on the undifferentiated sand and gravel deposit. Deposit thickness ranges from 1 to 20 m.

The best exposure of the deposit is near Glen Haven just north of M109 (S½,Sec.20.,T.29N.,R.14W.; about 0.3 km west of the D.H. Day Campground). There, about 3 m of mostly tan-to-light-orange, massive, medium sand is exposed and includes a continuous, brown-black, planar, organic layer about 0.5 to 6 cm thick. The uppermost 0.3 m is heavily bioturbated and has a 10 cm thick dark-brown soil layer that is partially covered with dune grasses and ferns.

Fan deposit. This deposit is composed of mostly massive-to-weakly-stratified, planar-to-crossbedded, poorly-sorted, pebbly-to-cobbly, matrix-supported gravel. The gravel consists chiefly of subrounded to-subangular carbonates, crystallines, and red sandstones 0.5 to 40 cm in diameter. The matrix consists mostly of medium-to-coarse, rounded-to-subrounded, quartz sand. Foresets are common as are truncated beds and imbricated pebbles and cobbles. Other characteristics are: infrequent, gray-red, silt laminae 0.15 to 2 m long and lenses 2 to 8 cm thick and 5 to 22 cm long; frequent, 5 to 15 cm thick and 0.2 to 7 m long, clast horizons; and infrequent boulders 15 to 40 cm in diameter.

In some areas, usually at or near the fan toe, the deposit contains more silt and fine sand whereas in other places it includes 0.4 to 3 cm thick and 1 to 2 m long, wavy-to-laminated, mud horizons. In other places, usually at or near the fan head, it is more pebbly and cobbly and includes beds of well-sorted sand 4 to 20 cm thick. Frequent exposures show that most of the fan deposits are 3 to 12 m thick.

Fans at the Lakeshore are common (Figure 12) but most are less than 1 km^2 in area. The best exposure of a fan deposit is located near the southwestern area of Little Traverse Lake (W½,Sec.14,T.29N.,R.13W.) and consists of about 2 m of fine-to-medium sand interfingering with 0.4 to 3 cm thick and 2 to 5 cm long, silt laminations and massive, red, clay horizons located near the fan toe or margin that is mostly superimposed on beach/dune and Glen Lake deposits. Locally, the upper part of the fan deposit is more pebbly and cobbly and includes 1 to 4 cm thick and 3 to 8 cm long, gravel lenses. Foresets dipping 40-70° are also common on the north and south ends of the fan.

<u>Modern delta and terrace deposit</u>. This deposit consists of complexes of thin bedded, planar-to-cross-stratified, moderate-to-well sorted, gravel-to-pebbly, matrix-supported, fine-to-medium sand. The sand consists chiefly of rounded-to-subrounded quartz whereas the pebbles consist of mostly carbonates, some crystallines, and some shales. Cross-bedded pebble and gravel layers are

common and include cross-sets, truncated beds, and fining-upward and fining-away sequences with topsets, foresets, and bottomsets. Other characteristics include: frequent, 0.2 to 3 cm thick and 1 to 3 cm long, clay laminae; frequent, 0.1 to 0.4 cm thick and 0.7 to 3 m long clay and silt lenses; frequent, stratified, laminated-to-thinly bedded silt and fine sand layers; infrequent, 1 to 3 cm thick and 4 to 15 cm long, clast horizons; and frequent, 2 to 8 cm thick and 0.2 to 6 m long, organic-rich layers. Locally, the unit is massive and chaotic. Unit thickness ranges from 1 to 3 m.

Modern delta and terrace complexes are located near bodies of water around the Lakeshore area. The best example of one of these complexes is located where the Platte River empties into Platte Lake near Honor (Figure 12). Here, the river, assisted by its north branch, has formed a delta nearly 1.2 km^2 in area and mostly covered by swamp and marsh deposits.

<u>Marsh and swamp deposit</u>. This deposit consists mostly of dark-gray-to-red-brown, well-sorted, laminated, fine-to-medium silt mixed with fine-to-medium sand. The silt and sand consists of mostly rounded-to-subrounded quartz. Individual bed thickness ranges from 0.5 to 3 cm. Other characteristics include: frequent, laminated, red-to-red-brown, clay beds 0.4 to 7 cm thick; frequent, dark-brown-to-black, thin-to-thick, peat deposits along with decaying plant and animal matter mixed with silt and sand; and infrequent, 0.6 to 3 cm thick and 2 to 18 cm long gravel and pebble lenses.

In some areas, the deposit contains more coarse sand as well as 2 to 6 cm thick and 0.4 to 5 m long, wavy-to-laminated muds. Locally, it is massive and chaotic. Deposit thickness generally ranges from 0.2 to 3 m.

Exposures of the deposit are common around the Lakeshore (Figure 12) and are generally associated with inland bodies of water located within protected embayments and lowlands. The best exposures occur in the vicinity of Otter Creek, located in Benzie County (S½,Sec.12,T.27N.,R.15W.) where Otter Creek flows northward from Otter Lake. Here, about 1 m of the deposit is exposed and consists of mostly peat and fine sand interfingering with tan-to-red-brown clay and silt. Locally, it contains rock clast horizons 4 to 17 cm thick and 0.3 to 2 m long. Some localized bodies of standing water are also common.

Modern beach deposit. This deposit consists mostly of tan-to-light brown, massive-to-laminated, moderate-to-well sorted, medium-to-coarse sand with gravel and pebble layers. The sand consists of subrounded-to-subangular, quartz sand whereas the gravel and pebbles consist of mostly subrounded-to-subangular carbonates and crystallines. Primary structural features include horizontal parallel lamination to low-angle and high-angle parallel lamination. Other characteristics of the deposit are: frequent, 0.2 to 3 cm thick and 4 to 25 cm long, reddish-brown-to-black laminations of heavy minerals; infrequent, 0.4 to 3 cm thick and 0.6 to 5 m long, dark-brown-to-black, organic-rich layers mixed with silt and fine sand; frequent, 0.4 to 6 cm thick and 0.2 to 1 m long gravel and pebble horizons; infrequent, subrounded to subangular cobbles 4 to 12 cm in diameter; and infrequent, subrounded to subangular boulders up to 2 m in diameter.

The modern beach deposit is associated with proximal sediments along the Lakeshore margin (Figure 26). It is superimposed mostly on older beach/dune complexes and Glen Lake deposits or on eroded surfaces. Locally, it is mixed with slumped material from bluffs. Deposit thickness ranges from 0.2 to 2 m.

The best exposure of the deposit is near the outlet of North Bar Lake, located about 4 km northeast of Empire. Here, about 0.6 to 1.5 m of it is exposed and shows mostly well-sorted, thick-bedded sand with a discontinuous 0.5 to 4 cm thick and 0.3 to 1.2 m long, gravel horizon occurring near the base of the exposure. Also common are 0.2 to 1 cm thick and 5 to 14 cm long, black-to-red, planar-to-wavy, fine laminations of heavy minerals.

Geomorphology

Drumlins. An elevated plain (Figure 5), 235 to 290 m in elevation and covering about 96 km^2, is located around Lime Lake, Little Traverse Lake, Duck Lake, and Lake Leelanau. The plain includes numerous, well-developed drumlins and drumlinoidal hills and has a gently undulating topography. Most of the drumlins are oriented roughly north-south but locally trend northeast-southwest. They are commonly elongated, low, and relatively mounded with smooth surfaces and range from 6 to 42 m high, 18 to 250 m wide, and 0.2 to 1 km long. Length-to-width-to-height ratios range from 0.1:0.4:2 to 0.5:6:17 (Figure 12, or see the

Good Harbor Bay U.S.G.S. 7.5" quadrangle). The drumlins vary in shape from oval to linear with the largest drumlins and drumlinoids exhibiting fluted slopes to the south and blunt noses to the north. They are generally separated from one another by swales and depressions 10 to 500 m wide, some of which are filled with standing water, swamps, and bogs.

Ridges and kames. High flat ridges with long axes that are in large part transverse to the margin of Lake Michigan form the most prominent landforms at the Lakeshore. The ridges include (all ridge names listed here are unofficial) Sleeping Bear, Empire, Point Betsie (or "Point Betsie Moraine", or "PBM"), Alligator Hill, Miller Hill, and Pyramid Point (Figure 17), many of which intersect the Lake Michigan shoreline to form steep bluffs (Figures 17, 26).

The Sleeping Bear ridge (Figure 5) is truncated at the Lake Michigan shoreline and forms a bluff nearly 100 m high (Figure 26) that towers over Lake Michigan. Along the shore, the ridge is generally capped by both active and partially vegetated, stabilized dunes. The Sleeping Bear ridge is oriented roughly northwest-southeast, is about 3 km wide and 8 km long, and has a flattened surface that slopes gently to the southeast.

Directly south of the Sleeping Bear ridge is Empire ridge (Figure 5). It consists of two to three rounded hills and also forms a bluff (Figure 26) that stands about 65 m above the surface of Lake Michigan. Like the Sleeping Bear ridge, Empire ridge is locally capped by dunes. About 14 km south of the Empire ridge is the PBM (Figure 5) which is about 85 m high and has a relatively hummocky surface. This ridge is the widest and longest and is 2 to 3 km wide and about 12 km long.

To the east of Sleeping Bear ridge is Alligator Hill (Figure 5) which stands about 50 m above Glen Lake (Figure 26). Like the PBM, it is oriented northwest-southeast but it is smaller and only 0.3 to 2 km wide and about 4 km long. Its surface consists of small, sharp peaks and crests that resembles an alligator's back, henceforth its name.

About 4 km east of Alligator Hill is a large ridge complex named collectively Miller Hill (Figure 5). It is 80 to 129 m high and separates the Glen Lake lowland from the Bass/School Lakes lowland. Unlike the adjacent ridges, Miller Hill is oriented roughly north-south and consists of mainly three interconnected ridges 1 to 2 km wide and 2 to 8 km long and two proximal isolated ridges 0.6 to 1 km wide and 1.5 to 3 km long that are separated by lowlands, sags, and channels. The surface of Miller Hill is mostly hummocky with wide, flat sections near its southern end as well as localized, incised channels about 5 to 23 m deep and oriented roughly east-west (Figure 5).

About 8 km northeast of Glen Lake and directly north of Miller Hill is Pyramid Point (Figure 17) which forms a bluff about 2.2 km long and stands about 80 m above the surface of Lake Michigan. Its surface is mostly hummocky and is locally trenched by several shallow channels 2 to 6 m deep and oriented roughly northwest-to-southeast. The ridge itself is also oriented roughly northwest-to-southeast and is about 2 km wide towards the northwest and tapers to a point toward the southeast (Figure 5).

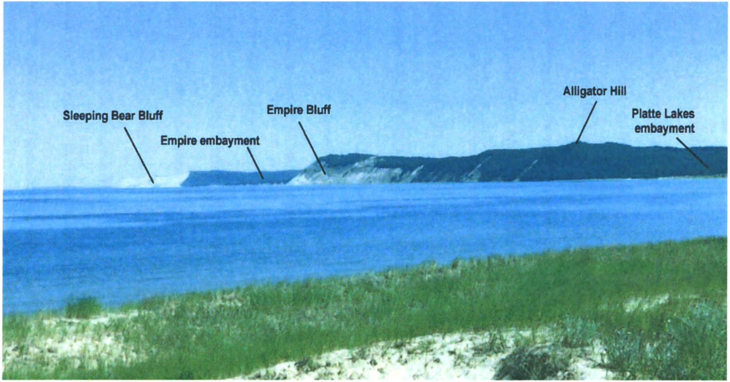


Figure 26. View of the bluffs occurring near Empire. The view is towards the northeast.

Also within the Lakeshore are several smaller ridges, many of which are less than 1 km wide and less than 2 km long. These form either sharp ridge crests or gently undulating hills oriented roughly northwest-southeast.

Kames and kamic hills. A north-south trending line of kames and kamic hills or mounds occur between Little Traverse Lake and Lake Leelanau. The most prominent of these is Sugarloaf Mountain, which has an elevation of about 323 m and is about 74 m of relief. Directly south of it is a series of progressively smaller, unnamed kames 30 to 55 m high while to the north of it in Good Harbor Bay is a submerged kame with a relief of about 22 m. This kame may have been partially exposed when lake levels were lower.

Lowlands, embayments, and channels. The Lakeshore also includes several wide, shallow, bay-shaped lowlands proximal to Lake Michigan (Figures 5, 12, 26). They are generally flat with elevations that range from 178 to 204 m a.s.l (Table 3) and are widest near the shoreline and narrowest from it. They are often partially occupied by lakes and swamps.

The Glen Lake lowland (Figure 5) is a semi-circular-to-rectangular-shaped depression about 6 km wide and 4 km long and is partially surrounded by the ridges of Sleeping Bear, Alligator Hill, and Miller Hill (Figures 5, 12). The lowland is partially filled by Glen Lake (Figure 5).

About 4 km northeast of Glen Lake is the Bass/School Lakes lowland. This is an irregularly-shaped depression about 1 km wide and 4 km long. It is bordered by Miller Hill to the west and a high drumlinized ridge to the east

(Figure 5). It is partially filled by Bass Lake and School Lake located north and south from each other, respectively. About 3 km east of Bass Lake is the Little Traverse/Lime Lakes lowland (Figure 5). This is a tongue-shaped depression that is 2 to 5 km wide and about 7 km long that lies between a high drumlinized ridge to the west and a high drumlinized plain to the east (Figure 5). The Little Traverse/Lime Lakes lowland is partially filled by Little Traverse Lake and Lime Lake that are located north and south from each other, respectively.

About 3 km south of the Glen Lake lowland is the Empire lowland (Figures 5, 26), which is an arcuate depression nearly 5 km wide that separates Sleeping Bear and Empire ridges. About 4 km south of the Empire lowland is the Platte Lakes lowland. It is a large arcuate depression about 6 km wide that separates Empire ridge to the north from PBM to the south. The lowland is partially occupied by Little Platte Lake and Platte Lake, located north and south from each other, respectively, and is divided by a low-relief linear ridge that separates the two lakes (Figure 5).

To the south of the Platte Lakes lowland is the Crystal Lake lowland. This is a rectangular depression 3 km wide and 7 km long and is partially filled by Crystal Lake (Figure 5). It is bordered to the north by the PBM and to the south by the hills and ridges that surround Benzonia. It is also separated from Lake Michigan to the west by dunes, spits, and bars.

Empire channel. Located between the Glen Lakes and Platte Lakes lowlands is a singular, north-south trending channel (Figures 12, 24, or see the Empire U.S.G.S. 7.5" quadrangle) with a floor (Figure 24) that ranges in elevation from

219 to 228 m, and consists of three large prominent loops or meanders. The meanders have a wavelength of 2 to 4 km and amplitude of about 3.5 km. The channel floor is 15-20 m deep. It is relatively flat, featureless, boulder-free, and is otherwise unremarkable except for a single, linear, elliptically-shaped sag roughly 12 m deep, 120 m wide, and 1 km long located around the center of the channel (SW¼,Sec.28/SE¼,Sec.29/NE¼,Sec.32,T.28N.,R.14W.). The channel also includes several smaller elliptical and oval-shaped depressions 6 to 9 m deep, 35 to 200 m wide, and 250 to 600 m long located further to the south.

Bluffs and scarps. Nearly 70 km of Lakeshore bluffs (Figures 5, 12, 17, 26) are exposed along the Lake Michigan shoreline. Dunes cap many of the bluffs. Most of the bluffs (Figures 17, 26) rise 50 to 120 m above the lake with slope gradients of 36 to 57° and are in varying stages of stability. Erosion and mass-wasting have modified the bluff faces along the Lakeshore with periodic slumps and landslides, especially at Sleeping Bear Point, Pyramid Point (Figure 17), and occasionally at the west-facing bluffs of North and South Manitou Islands. In general, those bluffs along the Lakeshore margin that are the most stable and less prone to mass-wasting are either protected to a degree from the elements by facing away from direct northeasterly storm wind and wave activity or are relatively stabilized due to their compositional nature.

Scarps. Elevations for well-developed scarps were noted from various locations (Table 3) around the Lakeshore area and its vicinity from Lake Leelanau south to the Upper/Lower Herring Lakes (Figure 5).

The scarp with the highest elevation lies at 225 m and occurs along the northeast-facing slope of Pyramid Point ridge and on west, north, and northeast-facing slopes of North Manitou Island. The next lower scarp elevation occurs at 213 m and is found only along some of the slopes at North Manitou Island below the 225 level. Below this elevation is a series of closely-spaced scarps that range from 200 to 203 m in elevation and are found on several westward-facing ridges and slopes near Duck Lake, Pyramid Point, Port Oneida, and North Manitou Island. Below this is another series of scarps that are mostly around 196 to 199 m in elevation and along both the westward-facing slopes of North and South Manitou Islands and the westward-facing ridges near Frankfort and Elberta.

<u>Dunes and beaches.</u> Dunes at the Lakeshore take the form of either perched dunes on bluffs and ridges (Figure 25) or lower-lying coastal beach/dune complexes usually located in lowlands and depressions. The perched dunes include Michigan's most famous dune, the Sleeping Bear Dune, a 32 m thick, rapidly eroding sand dune that is located at the edge of Sleeping Bear bluff (Figure 26). Sleeping Bear Dune and the other similar perched dunes along the Lakeshore margin are mostly parabolic in shape (Waterman, 1926; Snyder, 1985) and are best-developed on the steep bluffs 80 to 143 m high that face the strong northeasterly winds blowing off Lake Michigan. The perched dunes are locally both active and inactive, and in places are stabilized by dune grasses, thistles, ferns, and other vegetation. Locally, large blowouts occur (Figure 25), some over 100 m in diameter and 16 m deep. In some cases, "lag gravel" of

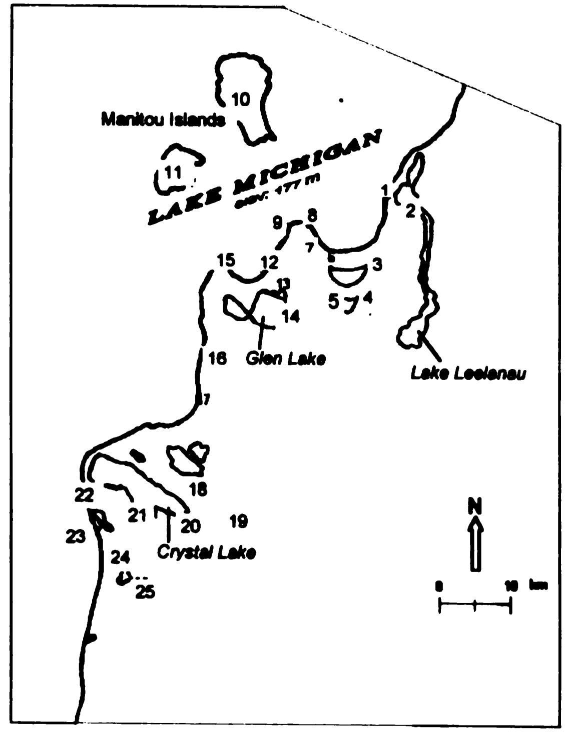
pebbles are left behind or exposed at the base of some of the deeper blowouts (Figure 25) while in others, wind-polished boulders, some more than 1 m in diameter, lie within them.

Beach/dunes at the Lakeshore are mostly arcuate ridges of sand (Figure 12), 1.3 to 20 m high and 0.2 to 5 km long, that mostly parallel the shoreline and are generally located within protected embayments. The elevations for these ridges range from 181 m to 189 m (Table 3). In general, beach/dune ridges conform to the present shoreline configuration. However, with increasing distance from the shoreline margin, some ridges trend either roughly normal or oblique to it.

Table 3. Scarp elevations at the Lakeshore and vicinity (in meters).
Note map (inset) for survey site locations.

SITE#	LOCATION	SCARP ELEVATIONS - m - (a.m.s.l.) Lowest>>>>>>>>>>Highest					
1	Duck Lake	196	201				
2	Lake Leelanau	205					
3	Little Traverse Lk.	179	180	186	189		
4	Lime Lake	189	191	200			
5	School Lake	200	205	222			
6	Shalda Creek	180					
7	Shell Lake	180	185				
8	Pyramid Point	200	203	225			
9	Port Oneida	185	200				
10	N. Manitou Is.	185	189	200	203	213	225
11	S. Manitou Is.	179	183	185	197		
12	Glen Arbor	180	190	217			
13	Glen Lake	185	190	195	215	223	
14	Brooks Lake	192	220	225			
15	Glen Haven	185	202	225			
16	Empire	190	191	203	225		
17	Otter Creek	182	185	190			
18	Honor	187	207	210	213	230	235
19	Platte River	200	220				
20	Beulah	189	192	223	225	235	
21	Onkeonwe Bch.	186					
22	Frankfort	197	221				
23	Elberta	190	199				
24	Lower Herring Lk.	181	185				
25	Upper Herring Lk.	183	184	197			

Note: scarp elevations as shown are correct to within ± 5 m.



SECTION III: GLACIAL AND POST-GLACIAL HISTORY

Deglaciation History

The Leland formation is the oldest surficial geologic unit at the Lakeshore and is interpreted to represent a complex sequence of glaciofluvial and glaciolacustrine sediments deposited at, or near, an ice margin. It consists mostly of three different facies: 1) a reddish-brown stony diamict facies, 2) a stratified sand facies, and 3) a blue-gray silt and clay facies with intercalated sand and gravel.

The origin of the reddish stony diamict facies (as shown in well logs b#2; s&g#3) is unclear but may represent debris flows and/or rainout till deposited in a subaqueous setting. Evidence to support this interpretation includes massive, structureless silt and clay that is described in well logs as highly heterogeneous and includes several interbedded "yellow" sand layers. These layers could represent finely-textured stringers in the diamict produced by mobility of sediment-laden fluids along marginal shear planes (Evenson et al., 1977; Boulton, 1979; Huddart, 1983).

The stratified sand facies (Figure 13) most likely represents glacial outwash of subaerial and subaqueous origin. Evidence in support of a subaerial origin includes laterally-continuous, coarsening-upward sequences of stratified, moderately-sorted, fine-to-medium sand to mixed sand and gravel (Gustavson, 1974; Boothroyd and Ashley, 1975; Blewett, 1990). Evidence in support of a subaqueous origin includes parallel-bedding, rippling, cross-bedding, truncated beds (Osborne, 1950; Shaw and Archer, 1978; Clifton and Dingler, 1984; Rovey II and Borucki, 1995), and rock clast imbrication (Eyles and Eyles, 1983). The imbrication of clasts indicates that the dominant flow direction was generally towards the south.

The blue-gray clay and silt facies (Figure 14) is most likely of low energy origin with probably nearly all of it deposited distal to the ice margin. Evidence to support this interpretation includes distinctly laminated (varves?) bottomsets, silty ripple trains and sandy horizons that show distinct signs of bottom current activity (Rust and Romanelli, 1973; Ashley, 1975; Karrow et al., 1975; Rovey II and Borucki, 1995), isolated clasts that are suggestive of dropstones (Eyles and Eyles, 1983), frequent silt and clay lenses (Eyles and Eyles, 1983), and primary sedimentary structures that include rippling (Shaw and Archer, 1978; Clifton and Dingler, 1984). The intercalated sand and gravel beds are interpreted either as a series of subaqueous debris flow, subaqueous fan, or rain-out deposits formed either in local proglacial lakes or in a single lake that was likely impounded by ice subject to sudden episodes of large volumes of water released periodically (Eyles and Eyles, 1983). In addition, the presence of well-sorted fine-grained material (Figure 14) suggests that deposition was subaqueous (Dreimanis and Goldthwait, 1973; Ashley, 1975; Huddart, 1983; Rovey II and Borucki, 1995).

Although the age of the Leland formation is uncertain, it clearly predates the other mapped units at the Lakeshore. Evidence in support of this is the fact that it directly underlies the Duck Lake diamict (Figure 13) which has been interpreted to represent an ice advance (Monaghan et al., 1986; Monaghan and Hansel, 1990). Therefore, the sand most likely mostly formed during an earlier ice margin retreat.

The Duck Lake diamict is represented by two facies at the Lakeshore exposed at the Leelanau County gravel pit (Figures 15,16) and at Pyramid Point (Figure 16) and most likely represents subglacial till. Evidence in support of this interpretation includes the fact that the diamict is generally massive, relatively structureless, poorly-sorted, and contains frequent flattened, striated, and elongated clasts and clast pavements (Boulton, 1968; 1979). The unconformable contact between the diamict and the underlying Leland formation (Figure 15), as well as downthrusted diamict wedges into the Leland formation, also supports this interpretation (Boulton, 1979; Acomb et al., 1982; Dreimanis, 1989). At the pit, the red-brown till thrusted over the more massive till that forms the core of the Duck Lake diamict may also represent a separate basal till formed during a subsequent ice readvance (Boulton, 1968; Lawson, 1979; Dreimanis, 1989). The drumlinized topography developed on the diamict surface would likewise suggest it was deposited as basal till when ice flowed in a southerly direction (Leverett and Taylor, 1915; Melhorn, 1954).

At Pyramid Point (Figure 16), the Duck Lake diamict probably represents deformation till. Evidence in support of this interpretation includes the fact that the diamict shows evidence of compressional folds, major shear planes (Figure 16), matrix-supported, rock clast lenses, inclusion of clasts with a preferred N-S orientation, and overconsolidation (Boulton, 1968; 1979; Dreimanis, 1989). The overconsolidation was likely produced by the heavy overburden of the ice while the folds, shear planes, rock clast lenses, and oriented clasts were likely produced by the stresses applied by the moving ice or compaction due to its weight or both (Boulton, 1968; 1979; Lawson, 1979; Dreimanis, 1989).

The age of the Duck Lake diamict is uncertain, but it is probably Port Huron or younger. Evidence that it is no older than Port Huron is based on the fact that it can be traced no further south than the Manistee (inner Port Huron) moraine as defined by Melhorn (1954). However, since the Greatlakean ice advance (Figure 2) may also have extended as far south as the Manistee moraine (Evenson, 1973; Hansel et al., 1985; Taylor, 1990), it is possible that all or part of the Duck Lake diamict was deposited by this advance. On the other hand, exposures and one well log (s&g#4) show that the Duck Lake diamict is locally overlain by the sand that comprises the Pyramid Point deposit, hence it must be younger than the Pyramid Point deposit.

The Pyramid Point deposit (Figures 17-19) includes ice-proximal, medial, and distal facies and probably represents a series of localized subaqueous fan deposits.

The ice-proximal facies probably represents high-energy subaqueous gravity deposits formed at the ice margin near vents and/or tunnel-mouths. Evidence in support of this interpretation includes stratified, moderately-sorted gravel intercalated with fine sand (Ashley, 1975; Eyles and Eyles, 1983; Rovey II and Borucki, 1995) while primary sedimentary structures that include cross-bedding, truncated surfaces, and troughs suggest upper flow regimes probably due to high discharge events occurring at the ice margin (Morgan, 1970; Rust and Romanelli, 1973; Boothroyd and Ashley, 1975; Wright, 1977; Blewett, 1990). Cross-bedding also suggests that the general flow direction was to the south.

The medial facies probably represents moderate-to-low energy subaqueous gravity flow deposits accumulating near the ice margin. Evidence in support of this interpretation are stratified, laterally continuous, coarsening-upward sequences of sand (Figure 18) that are locally intercalated with clay, silt, and gravel (Ashley, 1975; Karrow et al., 1975; Shaw and Archer, 1978; Eyles and Eyles, 1983). These sequences include what appear to be dropstones as well as primary sedimentary structures such as cross-bedding (Figure 18), which indicates that the flow direction was generally to the south, parallel bedding (Figure 19), frequent troughs and rippling, truncated beds, and rock clast imbrication (Rust and Romanelli, 1973; Boothroyd and Ashley, 1975). The distal facies was probably deposited within static water distal to the ice margin. Evidence in support of this interpretation includes the fact that it consists of massive-to-laminated silt and clay layers (Rust and Romanelli, 1973) that include what appear to be dropstones (Eyles and Eyles, 1983).

Pyramid Point deposits were most likely limited to restricted, or semi-enclosed, deep basins probably related to water-filled longitudinal crevasses developed along the active ice margin. Evidence in support of this interpretation includes layer-cake stratigraphy (Flint, 1928; Osborne, 1950; Klajnert, 1966; Huddart, 1983) and morphological evidence that includes long linear ridges with truncated flanks that roughly parallel the known southeasterly ice flow direction

(Leverett and Taylor, 1915; Calver, 1947; Klajnert, 1966; Huddart, 1983). The above evidence supports the theory that the ridges are representative of sediment containment by walls of ice during ridge construction (Flint, 1928; Huddart, 1983). The age of the Pyramid Point deposit is uncertain but its stratigraphic position above the Duck Lake diamict suggest it must have formed during either the Port Huron or Greatlakean stades.

The Pearl Lake deposit (Figures 20, 21) is of glaciofluvial origin. Evidence in support of this interpretation includes laterally continuous, fining-upward sequences of planar-bedded, well-sorted gravel and coarse sand that suggest an ice-proximal depositional environment (Boothroyd and Ashley, 1975; Blewett, 1990) and primary sedimentary structures (Figure 19) that include troughs, channels, truncated beds, and rock clast imbrication (Gustavson, 1974; Boothroyd and Ashley, 1975). The clast imbrication also suggests a southerly flow direction (Eyles and Eyles, 1983). Morphological evidence also indicates that the outwash of Pearl Lake deposit forms a sandur with a surface characterized by an elevated, extensive, mostly well-drained plain that is partially collapsed and filled by kettle lakes (Boothroyd and Ashley, 1975; Blewett, 1990).

A well-defined ice-contact margin occurs near the Glen Lake, Empire, and Platte Lakes lowlands (Figures 5, 12) and represents the proximal margin of the deposit. Near Glen Lake, the margin rises 66 m above the surrounding landscape and slopes northward at about 27° while near Empire and Honor, it slopes westward at about 22°.

The thickness of the Pearl Lake deposit suggests that large volumes of sediment-charged meltwater exited the ice front and quickly built up the outwash. The absence of fines in the upper part of the outwash (Figure 21) indicates that streams were highly efficient in winnowing out suspended sediment and the general fining-upwards characteristic suggests a steady decrease in flow velocity accompanied by an increase in the amount of suspended sediment (Wright, 1977; Blewett and Winters, 1995).

The age of the Pearl Lake deposit is uncertain but since it locally buries sediments associated with the Duck Lake diamict and Pyramid Point deposit, it is probably related either to the Port Huron or Greatlakean ice sheet and was most likely formed when the ice margin stood near the southern periphery of Glen Lake and near Empire and Honor.

The undifferentiated sand and gravel deposit is interpreted to be of glaciofluvial and/or glaciolacustrine origin and deposited locally against the proximal edge of the Pearl Lake deposit. Evidence in support of a glaciofluvial origin is weakly-stratified, moderately-sorted, coarse sand and gravel (Boothroyd and Ashley, 1975; Blewett, 1990) and primary sedimentary structures that include planar-bedding, cross-bedding, with occasional foresets, truncated beds, and rock-clast imbrication (Gustavson, 1974; Boothroyd and Ashley, 1975). Rock-clast imbrication indicates that the main direction of flow was to the southeast. Evidence in support of a glaciolacustrine origin includes the interbedding of laminated clay and silt layers (Ashley, 1975; Shaw and Archer, 1978), clay and silt lenses (Eyles and Eyles, 1983; Rovey II and Borucki, 1995), ripples, and wavy, laminated muds suggestive of active bottom currents that reworked and shaped the lake sediments (Morgan, 1970; Rovey II and Borucki, 1995).

The age of the undifferentiated sand and gravel deposit is uncertain but because it locally buries sediments associated with the Duck Lake diamict, Pyramid Point deposit, and Pearl Lake deposit, it is probably related either to the Port Huron or Greatlakean stades and deposited during the retreat of the ice margin from the edge of the Pearl Lake deposit.

The Honor deposit (Figures 22, 23) is interpreted as a delta complex. Evidence in support of this interpretation includes planar-to-cross-beds, moderately-dipping foresets (Gilbert, 1890; Morgan, 1970; Wright, 1977), and truncated surfaces. Foresets suggest that the delta prograded out into a lake and truncated surfaces suggest that the lake level fluctuated. The morphological evidence in support of a delta includes a relatively planar surface, sharp slopes along its margin (Gilbert, 1890; Wright, 1977), its position at the former outlet of the Platte River (Figure 12), and the fact that it is located along the margin of a paleolake (Calver, 1947; Taylor, 1990) that existed within the Platte Lakes lowland (Figure 4).

The age of the Honor deposit is uncertain, but because it locally rests against Port Huron or Greatlakean Pyramid Point and Pearl Lake deposits, it must be post-Port Huron in age.

Sediments associated with the stream terrace deposit are interpreted as fluvial in origin. Evidence in support of this interpretation includes such primary structures as crossbedding, troughs, foresets, and imbricated pebbles and cobbles (Rust and Romanelli, 1973; Boothroyd and Ashley, 1975; Lawson, 1982). The imbricated pebbles and cobbles suggest that the general flow direction was towards the south. In addition, the fact that the terraces are localized, restricted to a broad channel (Figure 24), and lack a kettled surface are good indicators that they are stream terraces.

The age of the stream and terrace deposit is uncertain, but it must be post-Port Huron because it rests on undifferentiated sand and gravel previously shown to be either Port Huron or Greatlakean in age. In addition, since it is restricted to the floor of a meltwater channel (Figure 24) formed during deglaciation, initial deposition and construction of the terraces must have begun shortly after the ice margin retreated from the head of the undifferentiated sand and gravel deposit to allow glacial melt waters to carve out the channel. Deposition and build-up of the terraces likely continued until flow through the channel ceased.

Sediments associated with the Glen Lake deposit are interpreted as lacustrine/glaciolacustrine in origin and were probably deposited proximal and distal to the ice margin. Evidence in support of a lacustrine origin includes the presence of fossiliferous, stratified, well-sorted, fine sand (Ashley, 1975; Shaw and Archer, 1978; Eyles and Eyles, 1983; Clifton and Dingler, 1984; Rovey II and Borucki, 1995), laminated silt and clay [(possibly varves?) (Ashley, 1975; Shaw and Archer, 1978)], and beds of bluish, fossiliferous marl (Shaw and Archer, 1978). Evidence in support of a glaciolacustrine origin proximal to the ice margin includes the presence of isolated clasts that are most likely dropstones (Eyles and Eyles, 1983), while evidence in support of a glaciolacustrine origin distal to the ice margin may include tabular, rippled, sand beds (Clifton and Dingler, 1984) as well as the laminated silt and clay previously described. Morphological evidence in support of a lacustrine/glaciolacustrine origin includes the fact that the deposit forms a poorly drained, broad, flat plain throughout the lowlands and is locally associated with elevated beach and delta deposits. The extent of the Glen Lake deposit would suggest that the lake into which it was deposited extended well into the Platte, Empire, and Glen Lakes lowlands (Figures 5, 12).

The age of the Glen Lake deposit is uncertain, but because it locally rests on the undifferentiated sand and gravel deposit previously shown to be associated either with Port Huron or Greatlakean ice, initial deposition of the Glen Lake deposit must have begun shortly after the ice margin withdrew north and westward from the head of the Pearl Lake deposit to allow glacial lake waters to flood the lowlands. Deposition continued as long as the lowlands remained flooded and ceased when lake levels fell or the land surface rose above lake level due to isostatic rebound. The Glen Lake deposit is time-transgressive and is at least as old as the Honor delta complex.

Sediments associated with the Sleeping Bear deposit (Figure 25) are of eolian origin. Evidence in support of this interpretation are well-draining, massive-to-laminated, well-sorted, fine sand (Bagnold, 1936) and primary structures that include cross-stratification and truncated beds (McKee, 1979). The morphological evidence in support of an eolian origin includes the fact that the Sleeping Bear deposit is associated with dunes that are perched on high bluffs along the shore well above any known lake level. The source of the sand that produces the dunes is mostly from the exposed bluffs (Waterman, 1926; Dow, 1938; Snyder, 1985). Dow (1938) and Snyder (1985) have also suggested that nearby modern beaches could also provide a secondary sand source.

The age of the Sleeping Bear deposit is uncertain, but because it rests on Pyramid Point deposits shown previously to be Port Huron or Greatlakean in age, deposition probably began soon after lake waters invaded the lowlands following ice margin retreat (Dow, 1938; Snyder, 1985). Snyder (1985) states that the age of the Sleeping Bear deposit is about 3,000 yrs B.P. based on ^{14}C dating of paleosols on which the deposit is superimposed. However, he mentions that the majority of eolian activity and dune buildup generally began about 1,000 yrs B.P (1985).

Sediments associated with beach/dune complexes are of littoral and eolian origin and were deposited proximal to the lake margin. Evidence in support of this interpretation are well-draining, massive-to-laminated, well-sorted, medium sand locally mixed with gravel and pebbles (Shepard and LaFond, 1940; Komar, 1971; DuBois, 1978), frequent occurrences of dark, fine, heavy minerals which suggest nearshore deposition (DuBois, 1978), and occasional buried, organic-rich layers that suggest partial lake level stability followed by dune migration as lake levels shifted (Kelley and Farrand, 1967). The morphological evidence in support of a littoral origin includes the fact that the deposits form arcuate-shaped, well-draining ridges that conform roughly to the present shoreline configuration and are mostly located within enclosed, semi-protected lowlands (Calver, 1947).

The age of the beach/dune complex is uncertain, but because it rests on Glen Lake deposits shown previously to be younger than the undifferentiated sand and gravel deposit, deposition must have begun shortly following lake regression out of the lowlands.

Sediments associated with fan deposits are interpreted to be mostly alluvial in origin. Evidence in support of this interpretation includes well-drained, weakly stratified gravel interbedded with wavy, laminated mud and well-sorted sand, and the fact that the coarsest sediments (pebbles and cobbles) usually occurs near the head of the deposit and grades into finer sediments downslope of the deposit (Boothroyd and Ashley, 1975). The morphological evidence includes low, triangular-shaped deposits at the mouth of valleys which issues from upland areas (Ryder, 1971; Boothroyd and Ashley, 1975), and multiple stream incisions along the surfaces of the deposit (Boothroyd and Ashley, 1975).

The age of the fan deposit is unclear, but construction probably began not long after deglaciation and continues today. The fact that many fan sediments overlie undifferentiated sand and gravel and Glen Lake deposits as well as on some beach/dune deposits supports this conclusion.

Sediments associated with modern delta and terrace deposits are interpreted to represent alluvial complexes. Evidence in support of this interpretation include planar-to-cross stratified clay, silt, and fine sand deposits with steeply-dipping foresets which include some gravel accumulations in the foreset beds (Gilbert, 1890; Wright, 1977). The morphological evidence in support of this interpretation includes planar surfaces that extend like arrowheads into inland lakes from stream outlets (Wright, 1977). In some cases, small spits extend laterally along the delta front.

The age of the modern delta deposit is uncertain, but deposition and build-up probably began shortly after the inland lakes at the Lakeshore stabilized to assume their present outline. The fact that they mostly overlie Glen Lake deposits, extend into the lakes, and are partially covered with marsh and swamp deposits on stabilized surfaces supports this interpretation.

Sediments associated with the marsh and swamp deposit are interpreted to be lacustrine in origin. Evidence in support of this interpretation includes poorly-drained, well-sorted, stratified accumulations of mud, fine sand, and peat (Swift, 1975). The morphological evidence includes the fact that many of these deposits occupy low areas associated with Glen Lake deposits as well as the fact that these deposits often occur around the margins of some inland lakes (Swift, 1975; DuBois, 1978).

The age of the marsh and swamp deposit is uncertain but because they overlie Glen Lake deposits and partially cover modern delta sediments they probably formed soon after lake regression out of the lowlands and inland lake levels stabilized.

Sediments associated with the modern beach deposit are of littoral origin and formed mostly by wave action and partly by eolian processes. Evidence in support of this interpretation include well-draining, massive-to-laminated, well-sorted sand with some gravel and pebble accumulations (Shepard and LaFond, 1940; Komar, 1971; DuBois, 1978) and the fact that these deposits rim most of the lakeshore with the slip face sloping lakeward (Komar, 1971).

The modern beach deposit is the youngest unit at the Lakeshore and it is likely that initial deposition began in the late Holocene and is continuing today. The fact that the deposit overlies Glen Lake deposits, older beach and dune deposits, marsh and swamp deposits, or eroded surfaces, supports this conclusion.

Extent and tills of the Port Huron and Greatlakean ice advances

Ice advances of the LML during the Port Huron Substage occurred at least twice between 13,300 and 12,200 years B.P. (Farrand and Eschman, 1974; Fullerton, 1980; Hansel et al., 1985; Blewett and Winters, 1995). The first advance formed the outer Port Huron moraine in Michigan (Figures 1-3) between 13,300 and 13,000 yrs B.P. (Farrand and Eschman, 1974; Hansel et al., 1985; Blewett, 1990) while the second advance formed the Manistee moraine/inner Port Huron moraine (Figures 1-3) between 13,000 and 12,200 yrs B.P. (Evenson, 1973; Farrand and Eschman, 1974; Eschman, 1985; Hansel et al., 1985; Taylor, 1990; Blewett and Winters, 1995).

According to Leverett and Taylor (1915), Bergquist (1954), Melhorn (1954, pg. 32), and Monaghan et al. (1990, pg. 47) the Port Huron advance is marked by a red-brown clay drift. Bergquist (1954) distinguished two tills near the Lakeshore and described the lowest as a highly-compacted brown till deposited during the Cary ice advance. Melhorn also stated that he found multiple tills underlying the drumlins around Leelanau County near the Lakeshore and suggested that the oldest till was deposited during the Cary and Port Huron ice advances. However, he states that most of the till underlying the drumlin surfaces is composed of a red, clay-rich till that he interprets as Valders. Later, Monaghan et al. (1990) noted several tills occurring within the drumlins around Leelanau County and found that the clay mineralogy of the lowermost till is similar to Ozaukee, Haven, and Valders tills from eastern Wisconsin deposited during the Port Huron ice advance (Figure 3).

As previously discussed, Leland formation likely formed during the retreat of the ice margin from the outer Port Huron moraine and may include till or diamict related to the ice advance that formed the outer Port Huron moraine (Figures 1-3). On the other hand, the Duck Lake diamict was probably deposited by Port Huron ice that advanced to the position of the inner Port Huron moraine (Figures 1-3). The diamict has similar color and sedimentological characteristics to other Port Huron tills observed by Melhorn along the Lake Michigan basin up to the position of the inner Port Huron moraine (Figure 4).

The extent of the second Port Huron ice advance is also marked at the Lakeshore by several large mega crevasse-fill, kame, and outwash deposits (Figure 12) that signify an active ice-margin which produced enormous volumes of glacially-derived sediment via complex subglacial conduits, vents, as well as englacial and supraglacial drainages.

The assignment of a Port Huron age to the Duck Lake diamict modifies Melhorn's hypothesis (1954) that the till coring the drumlins (*Duck Lake diamict*), as well as the pink, sandy-silty till (Taylor, 1990) underlying the crevasse-fill (*Pyramid Point deposit*) and outwash (*Pearl Lake deposit*) deposits at the Lakeshore, was produced by Valders (Greatlakean) ice (Figure 2). The Duck Lake diamict is probably stratigraphically correlatable to the Port Huron till of Leverett and Taylor (1915) and perhaps even to the Haven and Valders Member of the Kewaunee Formation of eastern Wisconsin (Evenson, 1973; Hansel et al., 1985; Peterson, 1986; Monaghan et al., 1990; Rovey II and Borucki, 1995). However, this linkage is tenuous and further work is necessary before any stratigraphic relationships can be properly determined.

The advance of Greatlakean ice (Valders of Melhorn) to the position of the Manistee moraine (Figures 2-4) occurred between about 11,800 and 11,200 yrs B.P. (Thwaites and Bertrand, 1957; Evenson, 1973; Schneider, 1990; Schneider and Hansel, 1990). According to Melhorn (1954), Evenson (1973), and Taylor (1990), it is marked by a thin deposit of distinctive, red, clay-rich till whereas, by comparison, the color and texture for the Port Huron ice advance is marked by a brown, sandy till (Bretz, 1951; Bergquist, 1954; Melhorn, 1954; Hough, 1955). Melhorn's limit of the Valders ice advance (1954, pg. 159, Plate 11) in northwest lower Michigan was based on his view of the extent of red-clay till. His limit would have Valders ice extending over the drumlins cored by Duck Lake diamict to the outer flanks of the Manistee moraine (inner Port Huron). This view was supported by Taylor (1990) who describes what he believes is a "Greatlakean" red sandy till occurring along the flanks of the PBM and individual ridges in the Benzie County area. In addition, Monaghan et al. (1990) analyzed the clay mineralogy of till in Leelanau County and found it to be similar to that of Two Rivers Till in eastern Wisconsin which is of Greatlakean age (Evenson, 1973).

Despite Melhorn (1954) and Taylor's (1990) arguments, the extent of their "Valders/Greatlakean" till at the Lakeshore is unsupported since there is no clear field evidence to support the opinion that Greatlakean ice ever extended along the PBM or over ridges or high plains in the Benzie County area. In essence, there is no evidence at the Lakeshore for a stratigraphic marker that shows Valders/Greatlakean-age, red, clay-rich till superimposed on Port Huron-age, brown, sandy till (Bretz, 1951, Bergquist, 1954, Melhorn, 1954, Hough, 1955, Evenson, 1973, Taylor, 1990). Furthermore, the surficial geologic map of the Lakeshore (Figure 12) shows no characteristic landforms like Greatlakean-related end, recessional, or ground moraines overlying Port Huron-related landforms (Eschman, 1985; Hansel et al., 1985; Taylor, 1990; Rovey II and Borucki, 1995). These facts suggest that Greatlakean ice either never advanced out of the Lake Michigan basin into the Lakeshore region to the position of the Manistee moraine, or it only extended as thin lobes into the lowlands for no more than 10 km and never covered the high ridges and plains.

The view stated above is in agreement with other studies in the Lake Michigan basin that suggest the Valders/Greatlakean ice was thin, occupied mainly lowlands, and was constrained by preexisting topography (Evenson, 1973; Farrand and Eschman, 1974; Hansel et al., 1985; Schneider, 1990; Larson et al., 1994). What's more, others have suggested that this advance is better represented in northern lower Michigan by erosion than by accumulations of till (Evenson, 1973; Eschman, 1985; Taylor, 1990). Finally, many workers have recognized that the red-clay lithology is not unique to the Greatlakean ice advance and may instead occur in deposits related to the Port Huron ice advance (Evenson, 1973; Farrand and Eschman, 1974; Eschman, 1985; Schneider and Hansel, 1990). Such observations probably also apply to the Lakeshore in that the high landforms there likely prevented the lateral spread of Greatlakean ice beyond the lowlands (Figure 27). If any Greatlakean till/diamict was deposited in the lowlands, it is now probably buried by subsequent lacustrine sediments associated with the Glen Lake deposit.

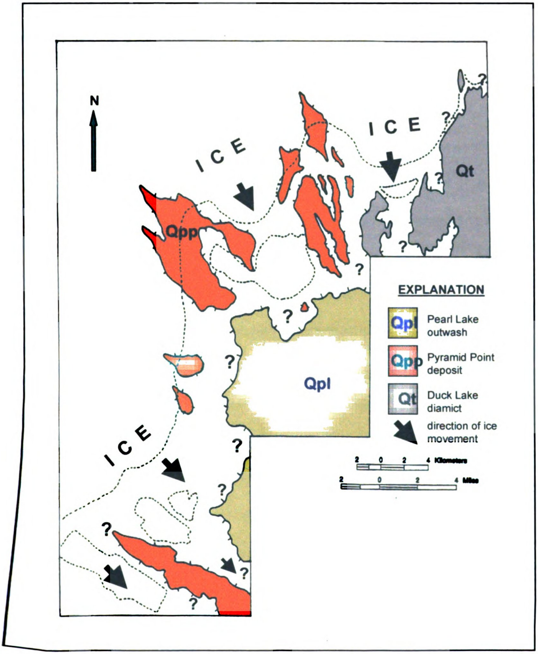


Figure 27. Hypothetical extent of Greatlakean ice conforming mainly to lowland areas along the Lakeshore (≈11,800 yrs B.P.).

Origin of the ridges at Sleeping Bear Dunes National Lakeshore

As previously discussed, sediments associated with the Pyramid Point deposit (Figures 17-19) likely developed within longitudinal fissures along an active ice-margin. These fissures deepened and widened with time to accommodate high volumes of sediment that entered via subglacial and englacial tunnels located along the ice margin. Once the ice margin retreated fully into the Lake Michigan basin, construction of the Pyramid Point deposit ceased, partial collapse redistributed some of the sediments, and water levels fell below the tops of the corresponding ridges.

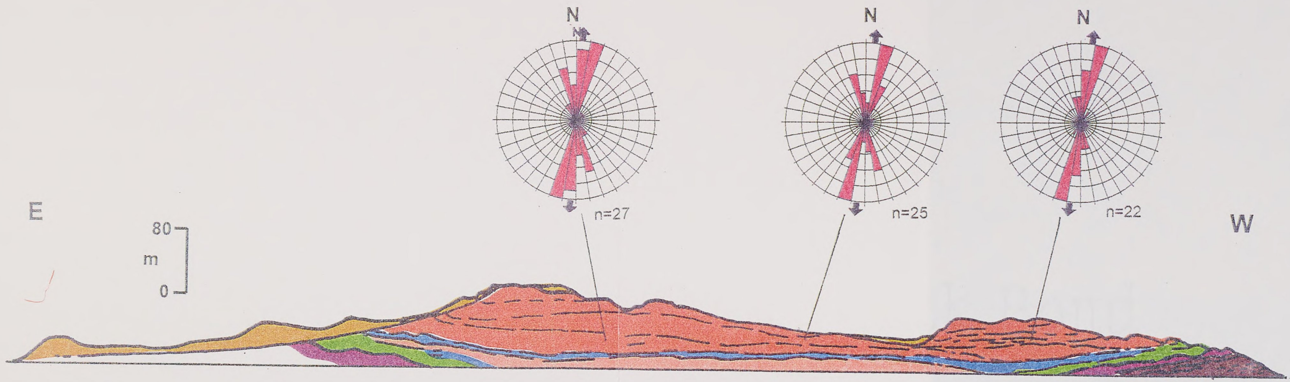
The overall heterogeneity inherent in the complex depositional environment associated with the crevasses is probably best illustrated in the bluff stratigraphy at Pyramid Point as shown in Figures 19 and 28. Here, at least three different depositional facies are present: an ice-proximal facies, a middle facies, and a distal facies. The gravels (Figure 28) represent the ice-proximal facies, where coarse detritus is ejected from the nearby ice margin and deposited almost immediately. This facies has numerous planar-convex boundaries (Figure 19) and isolated flow structures and was likely derived from vent or tunnel-mouth discharge in sediment-charged plumes, rainout from overturning and melting of icebergs, and superglacial meltwater depositing supraglacial sediment along the ice margin. Coarse sediment could also have been transported away from the ice margin into the crevasse by debris flows that were energized by meltwater pulses and deposited as massive gravel beds and lags.

The upper surface of some of the gravels appear to have been subsequently shaped and eroded by fluvial processes producing troughs and channels and are overlain by fluvial deposits with ripples and cross-sets (Figure 19) associated with the middle facies of the Pyramid Point deposit. This middle facies includes stratified sand bodies (Figure 28) that were deposited as either debris flows or sediment gravity flows. Sorting and remobilization of the sand probably also occurred subaqueously as sediment flows which were either fanned out in broad aprons forming thin sheets, or as individual massive bodies that moved as a unit and filled in holes, erosional contacts, and depressions caused by scour.

The fine-grained sediments (Figure 28) of the distal facies probably settled over the middle-facies as the ice margin backed away and produced thick accumulations of clay and silt rythmites. The distal facies includes red-brown clay and silt deposits (Figure 28) with possibly dropstones that were probably deposited during periods of slack water, or when sediment outflow was temporarily restricted.

The entire sequence probably occurred within a trough-type subaqueous system walled by ice. As the ice margin regressed, oscillatory sediment pulses were produced forming intraformational truncated surfaces that grade away from the ice margin. These surfaces were probably caused by channels that eroded into the deposit during powerful spring and summer flows. Finally, the continued regression of the ice-margin retreated from the crevasse fills and other associative deposits such as kames and caused the water-filled fissures to drain.

Figure 28. Description and distribution of the main lithologic units at Pyramid Point.



Description of Units

AEOLIAN/BEACH DUNE SAND. Locally includes peat, soils, buried soils.	SAND, massive to trough cross-bedded. Locally includes clay, silt ribbons, gravel lenses, and boulders.
SAND, medium-coarse, planar bedded with occasional troughs, ripples. Gravel-rich in upper unit.	SAND, massive to trough cross-bedded. Locally includes gravel lenses.
CLAY, SILT, and fine SAND, massive to laminated. Locally includes dropstones.	CLAY, SILT, and fine SAND, massive to laminated. Deformed in upper unit.
GRAVEL, massive, with medium-coarse sand in lower unit.	

Figure 28

A proposed conceptual model of a subaqueous crevasse-fill deposit

A model for the origin of the subaqueous crevasse-fills at the Lakeshore is given in Figure 29 and shows characteristics typical of such deposits. Tunnel-mouths ("TM") debouch sand and gravel directly into a large, open, water-filled crevasse (Figure 29: # 3) to form planar-bedded, graded sand and gravel beds that directly overlie basal till with current flow direction normal to the ice margin. Supraglacial and englacial runoff (# 6) carry loose particles into the crevasses, while bottom currents redistribute sand and gravel away from the ice, forming ripples, truncated ripples, and cross-sets, and leaving behind pebbles, cobbles, and boulders. Thick (meter-scale) massive-to-laminated deposits of fine material occur and are locally punctuated by dropstones and rain-out (# 16) from icebergs floating overhead. As the ice margin retreats, the ridge is completely abandoned and the supply of sediment is terminated. Truncated sides, partly offset by listric normal faulting and partial collapse (# 7) as the deposit expands out of the confinement of the crevasse in the ice, mark the extent of the filling inside of the crevasse.

This model is based on personal field observations and interpretations made about ridges at the Lakeshore as well as on written accounts and interpretations noted by researchers describing similar ridges in Canada (Osborne, 1950; Duckworth, 1978), Europe (Klajnert, 1966; Huddart, 1983), and the United States (Flint, 1928; Colton and Jensen, 1953; Blewett, 1990).

However, most of these ridges are not quite as large and extensive as those at the Lakeshore.

The lack of normal faulting along the margins of the bluff exposure at Pyramid Point is puzzling (Figures 17, 28). However, several simple theories may help to explain this. They include the possibility that: 1) the east and west ends of the bluff are poorly exposed due to local slumping (likely along normal faults) and talus cover which may conceal any structures; 2) evidence of faulting may have been removed by subsequent erosion by Greatlakean ice; and 3) erosion associated with post-glacial lake-level fluctuations may have undercut and eroded the flanks of the filling and removed any evidence of faulting.

Figure 29. A conceptual model and general depositional environment for mega crevasse fillings at the Lakeshore.

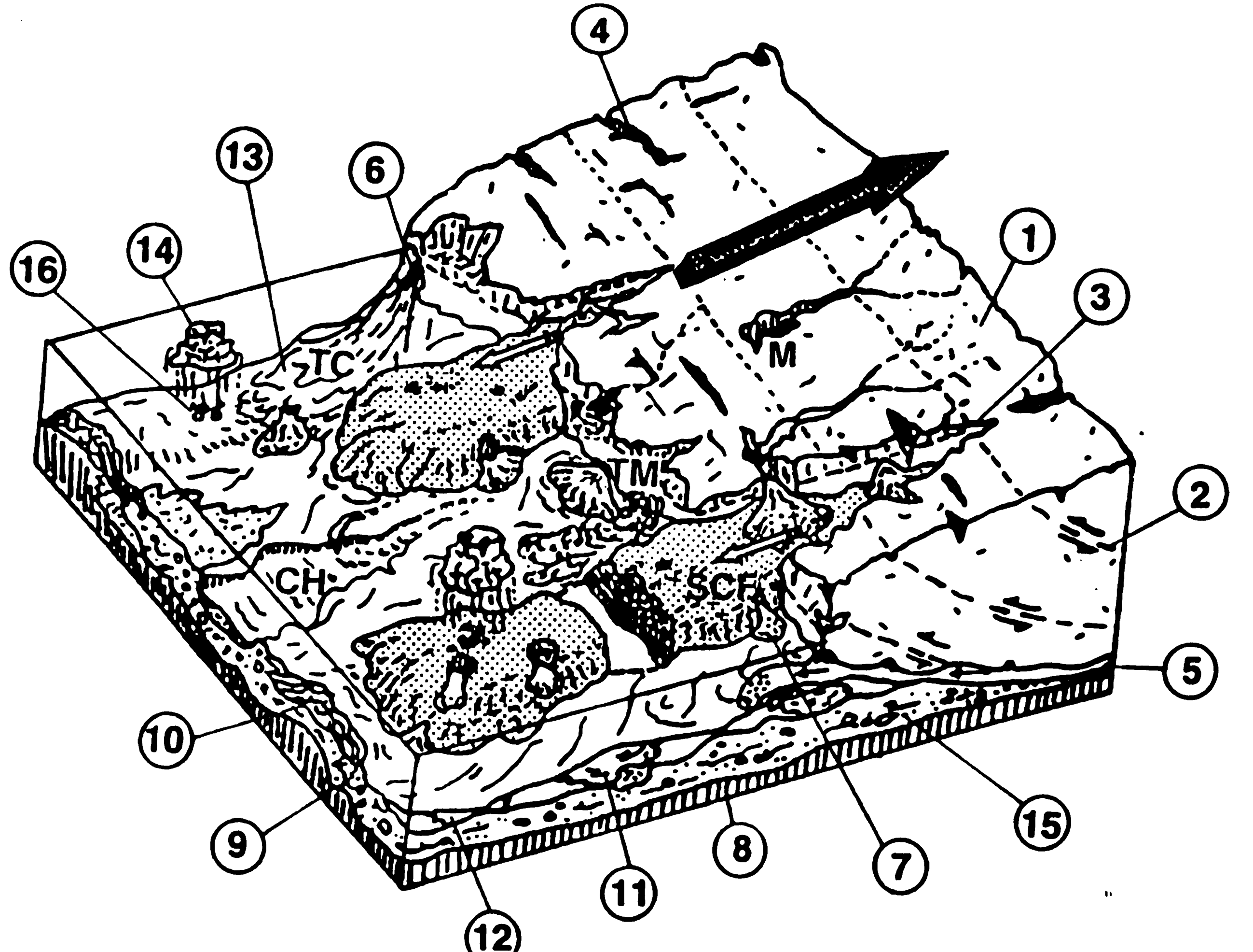


Figure 29

Subaqueous crevasse fillings formed during the retreat of a grounding ice front in a terminoglacial lake.
SCF = subaqueous crevasse fill; CH = channel; TM = tunnel mouth; TC = turbidity current; M = moulin.
1 = active ice; 2 = flow lines within ice; 3 = mega crevasses; 4 = supraglacial crevasses; 5 = subglacial till; 6 = englacial tunnel; 7 = subaqeuous slumps; 8 = substratum; 9 = buried ice; 10 = tunnel mouth deposits; 11 = diamicts; 12 = deltaic facies; 13 = subaqueous flow till; 14 = ice bergs; 15 = transgressive deposits; 16 = rain-out till

Discussion

Excellent lakeshore exposures of the Manistee moraine at the Lakeshore show distinct bedforms and structures that do not fit Leverett's (1915) classical definition of a moraine. Instead, the ridges at the Lakeshore seem to be indicative of infilling of large (*mega*) crevasses by subaqueous fan deposits, proximal to an active ice margin, as originally proposed by Calver in 1947.

The origin of the massive-to-laminated, pink-to-red, lacustrine, clay beds in the bluffs and ridges at the Lakeshore has long been controversial. Desor (1851) was the first to discuss the clay beds and observed that they are widespread and overlie older deposits. Leverett (1915) too noted the clay beds and stated correctly that they likely represent a water-laid glacial deposit formed either within the Wisconsin stage of glaciation or during a pre-Wisconsin interglacial. He also noted the high elevations of the clay which are higher than any known lake level and suggested that it was a result of subaqueous deposition between lobes of ice (pg. 315) but added that more data were needed before its origin could be determined with any accuracy. However, despite the fact that the ridges contained lacustrine sediments, he concluded that they are part of the Manistee moraine and were likely overridden by later ice.

Leverett (1915) also described the Manistee moraine as it appears within the Lakeshore as "a series of loops" and suggested that the ice that formed the moraine "...had made a readvance and had adjusted its border to these topographic features" (pg. 308). He found the inter-ridge depressions, partially occupied by Glen, Platte, and Crystal Lakes (Figure 5), blanketed with till to some degree and speculated that the till was deposited by a later ice advance following formation of the Manistee moraine (pg. 309). Clearly, the timing of the deposition of the Lakeshore ridges troubled Leverett, and he hinted that they may have formed prior to the ice advance (Figure 2) that formed the Manistee moraine (Figure 4).

Leverett and Taylor's use of the term "moraine" when describing the Manistee moraine follows accepted views at the time, when any kind of hummocky topography inferred a moraine deposit. Martin's (1955) map adhered to this definition, as did Farrand and Bell's (1982) map.

A more recent definition by Flint (1971, pg. 199) expanded on this classical approach by stating that moraines are composed chiefly of till, and are the result of direct deposition by active glacial ice. In addition, during Leverett's time, there was also greater emphasis on the importance of an advancing ice margin in producing modern glacial landscapes.

Not all early researchers accepted Leverett's general observations about the origin of the ridges at the Lakeshore. Calver (1947), working on developing the glacial and post-glacial history of the Platte and Crystal Lakes lowlands (Figure 5) in Benzie County, expressed doubt that all the ridges there (PBM and other associated linear ridges) were moraines. Instead, he suggested that "...these ridges may have formed as crevasse fillings in a stagnant block of ice that once occupied the Platte Valley." (pg. 13). However, Calver failed to provide any solid sedimentological evidence for this interpretation, nor did he suggest when the ridges formed. Nonetheless, he was particularly intrigued by the apparent linearity of the ridges and correctly observed that they were generally parallel to ice flow (northwest-southeast).

Melhorn (1954) later reported that "between Benzonia and Empire, the moraine (Manistee) is predominantly sand and there is very little till" (pg. 88). Taylor (1990) described thick accumulations of stratified sand and gravel occurring beneath the surface of the PBM and suggested that it represents an outwash deposit of Port Huron age.

To summarize, this thesis has demonstrated that the clay beds in the bluff exposures first noted by Desor (1851) were deposited in a lacustrine environment as observed by Leverett (1915) but not directly by advancing glacial ice, and that the ridges themselves are not composed of till. They were formed at a stagnant Port Huron ice margin. Therefore, they are not moraine deposits as originally suggested by Leverett (1915). Finally, they are not outwash deposits as suggested by Taylor (1990) since they are isolated, truncated ridges of lacustrine material oriented normal to the ice margin.

The major implication therefore is that the western part of the Manistee moraine (Figures 4, 5), as it occurs in the Lakeshore area as mapped (Figure 12), has been misinterpreted. Instead, the Manistee moraine at the Lakeshore was not produced by advancing Port Huron ice about 13,000 yrs B.P. but was formed by active, unstable, retreating ice probably during either the Port Huron or Two Creeks retreat sometime between 13,000 and 12,200 yrs B.P.

History of crevasse-fills

Crevasse fill deposits have been described in the literature since 1928 (Flint, 1928) and are referred to under a variety of names that include kames, kame ridges, disintegration ridges, ice-crack ridges, ice disintegration ridge networks, or ice-walled lake deposits (Flint, 1928; Calver, 1947; Colton and Jensen, 1953; Klajnert, 1966; Huddart, 1983; Blewett and Winters, 1995; Ham, 1998). Generally, the basic model is the same in most cases, where in-filling of supraglacial, englacial, or subglacial cracks, fissures, or crevasses with mostly ice-contact sediments is delivered via runoff, debris-bands, plumes, and/or drainage networks over, within, or beneath a stagnant glacier.

There have been few detailed observations made regarding mega crevasse fillings in the U.S. and Canada. Application of the term "mega-crevasse fill" for large linear ridges of the type observed at the Lakeshore is nonexistent and instead the emphasis historically has been on describing smaller landforms.

Flint (1928) was the first to mention and describe the chief sedimentological characteristics of crevasse fillings as they occur in the Connecticut area. Flint observed that these features are linear, flat-topped ridges, with the largest less than 11 m in height and 1.6 km in length, and most less than 3 to 5 m in height and about 400 m in length (pg. 414). Bedding is mostly horizontal to subhorizontal with transverse sections exhibiting lateral slump. According to Flint, these crevasse fill deposits formed within water-filled fractures or crevasses at or near the edge of a stagnant ice mass (pp. 414-415).

ice necessarily must be stagnant, Flint states (1928), otherwise movement would have disrupted these deposits.

Slightly larger crevasse fill deposits were mapped in northeastern Montana by Colton and Jensen (1953) who described essentially straight ridges of till 6 m in height, 30 to 60 m in width, and up to 3.2 km in length, with some ridges intersecting at angles ranging from 30 to 90 degrees. However, they found little evidence for subaqueous deposition, and, as a result, concluded that ice wastage was slow and produced only minimal amounts of meltwater.

Duckworth (1978) described coarse-textured, buried crevasse fillings with heights up to 20 m and with lengths up to 450 m of Port Huron age in the vicinity of Stouffville, Ontario. He describes deposition as either occurring at a tunnel mouth into standing water within a wasting, retreating ice front or the filling of linear crevasses in an ice front projecting into standing water. Both scenarios involve subaqueous deposition near a stagnating ice margin. The lack of extensive clay beds, however, seems to support the likelihood that these deposits formed by debouching of detritus via a tunnel mouth of some sort, or may even be eskers.

Klajnert (1966) described 1 to 1.5 km wide, 40 to 80 m thick, and 4 to 8 km long flat-topped ridges which comprise the Domaniewice Hills occurring in Central Poland in the vicinity of Lowicz. These deposits were previously thought to be an end moraine produced by the waning phase of the Wartian ice sheet and were reinterpreted by Klajnert to represent crevasse fillings . These ridges are roughly the size and scale of the ridges at the Lakeshore as well as similar in general sedimentology and structure. However, Klajnert (1966) interprets their formation as having occurred into "dry" crevasses by glaciofluvial deposition.

The extent of ridges (crevasse-fills?) in the Lake Michigan basin. As stated earlier, for the Lake Michigan basin, the emphasis has historically been on describing deposits and landforms associative with advancing ice margins and not stagnating ones (Leverett and Taylor, 1915; Thwaites and Bertrand, 1957; Hough, 1958). Recently, several workers have noted a marked absence of till in landforms associated with Late Wisconsin ice advances in the basin (Blewett, 1990; Clark and Rudloff, 1990; Blewett and Winters, 1995; Rovey II and Borucki, 1995; Ham, 1998). Blewett (1990) and Blewett and Winters (1990; 1995) contest that the inner and outer Port Huron moraines near Kalkaska are actually collapsed heads of glaciofluvial outwash for the nearby Mancelona and outer Port Huron outwash plains formed by melting Port Huron ice and are therefore not classic end moraines. Blewett's sedimentological work on the ridges reports very little till but mostly water-sorted, stratified sand and gravel which is interpreted as representative of major heads of outwash. Blewett and Winters (1995) also described several small, linear crevasse-fills that radiate away from the main inner Port Huron ridge like spurs, which they associated with the outwash and which therefore act as climatic markers for the transition to warmer temperatures with final deglaciation.

In the southern region of the basin along the western shore of Lake Michigan in northern Illinois, coastal bluffs showing stratified deposits which had been earlier interpreted as till sheets, were reinterpreted as subaqueous units by

Clark and Rudloff (1990). In addition, Rovey and Borucki (1995) reported on similarly stratified sediments in the St. Francis bluffs located in southeastern Wisconsin (Figure 2) and suggested that they represent subaqueous fans. Ham (1998, pg. A-113), in an abstract presented at the Geological Society of America national convention, described broad, hummocky moraines partially underlain by collapsed lacustrine sediment occurring in north-central Wisconsin. He interpreted these landforms as ice-disintegration landforms developed at the margin of the Laurentide ice-sheet about 18,000 to 15,000 yrs. B.P. and suggested that stagnation of the ice sheet persisted for several thousand years after the margin reached its maximum extent.

High transverse ridges are not unique to the Lakeshore area but are present elsewhere along the Lake Michigan basin. Analysis of topographic and 2° (4°X6°) surficial geologic maps of the Lake Michigan basin suggest that morphologically-similar ridges are widespread and seem to occur predominantly along the eastern margin of the basin. Specifically, the greatest frequency is along the modern lakeshore in western Michigan, especially north of Muskegon, whereas a few smaller subdued ridges occur in the northeast part of the basin around Charlevoix and Emmet Counties.

In the area between Frankfort/Elberta and Manistee are several ridges associated with the Manistee moraine that trend roughly east-west to northwest-southeast. There, the largest ridge, located just south of Elberta, is about 1 km wide and nearly 8 km long. To the south are several east-west trending ridges in the Big Sable/Ludington area that are associated with the Lake Border moraine and are 0.8 to 2 km wide and up to 12 km long. A single, long, broad, northeast-southwest-trending ridge, associated with the Whitehall moraine, occurs near Muskegon and is about 1 km wide and 7 km long. Just west of the Straits of Mackinac and around the Indian River lowland (Figure 2) are several ridges of various dimensions but with low relief. These low ridges mostly trend north-south or northwest-southeast and are associated with, and are oblique to, the inner Port Huron morainal belt (Blewett, 1990; Blewett and Winters, 1995).

Less distinct ridges occur along the western margin of the Lake Michigan basin and are generally more difficult to separate from adjacent morainal landform assemblages. In the Kewaunee area of eastern Wisconsin along the southern part of the Door Peninsula (about 100 km west-southwest of the Lakeshore-Figure 2) are several indistinct ridges that trend northeast-southwest and are therefore mostly normal to the western margin of the LML. In addition, several low ridges 30 to 60 m high also occur further to the south along the shoreline between Manitowoc and Sheboygan.

At this time, no workers have yet provided an explanation for the predominance of larger, more definable transverse ridges along the eastern margin of the Lake Michigan basin. However, several simple theories are possible. These include the following: 1) the LML during the Port Huron substage was thickest along its eastern margin and therefore formed the thickest accumulations of material, 2) the LML margin was "dirtier" along its eastern periphery since it had dredged up lacustrine sediments out of the basin and so provided the large volumes of material needed to build the high ridges during its melting, and 3) there were a higher occurrence of tensional fractures along the eastern ice margin than along its western margin as the LML splayed out over the Michigan mainland. In addition, these fractures were likely formed because the shear strength of the ice front was overcome due to possibly higher shear stresses (due to obstacles? greater material loads?) over on the Michigan side of the LML. This shear component probably helped to produce some of the cracks, joints, and fracture planes necessary to entrap sediment along the ice margin (Boulton, 1979). Foreland and/or subglacial material was, as a consequence, entrained or intruded into the fracture planes within it (Boulton, 1979; Lawson, 1979; Huddart, 1983). During the subsequent retreat of the ice margin, melting occurred preferentially along the fracture planes within it. As a result, these fractures widened and eventually were filled in with material (all subaqueously?) to form the ridges. One or all of these reasons may provide the answer. Even so, they are but a few out of multiple explanations. However, a proper treatment of this issue lies outside the scope of this thesis.

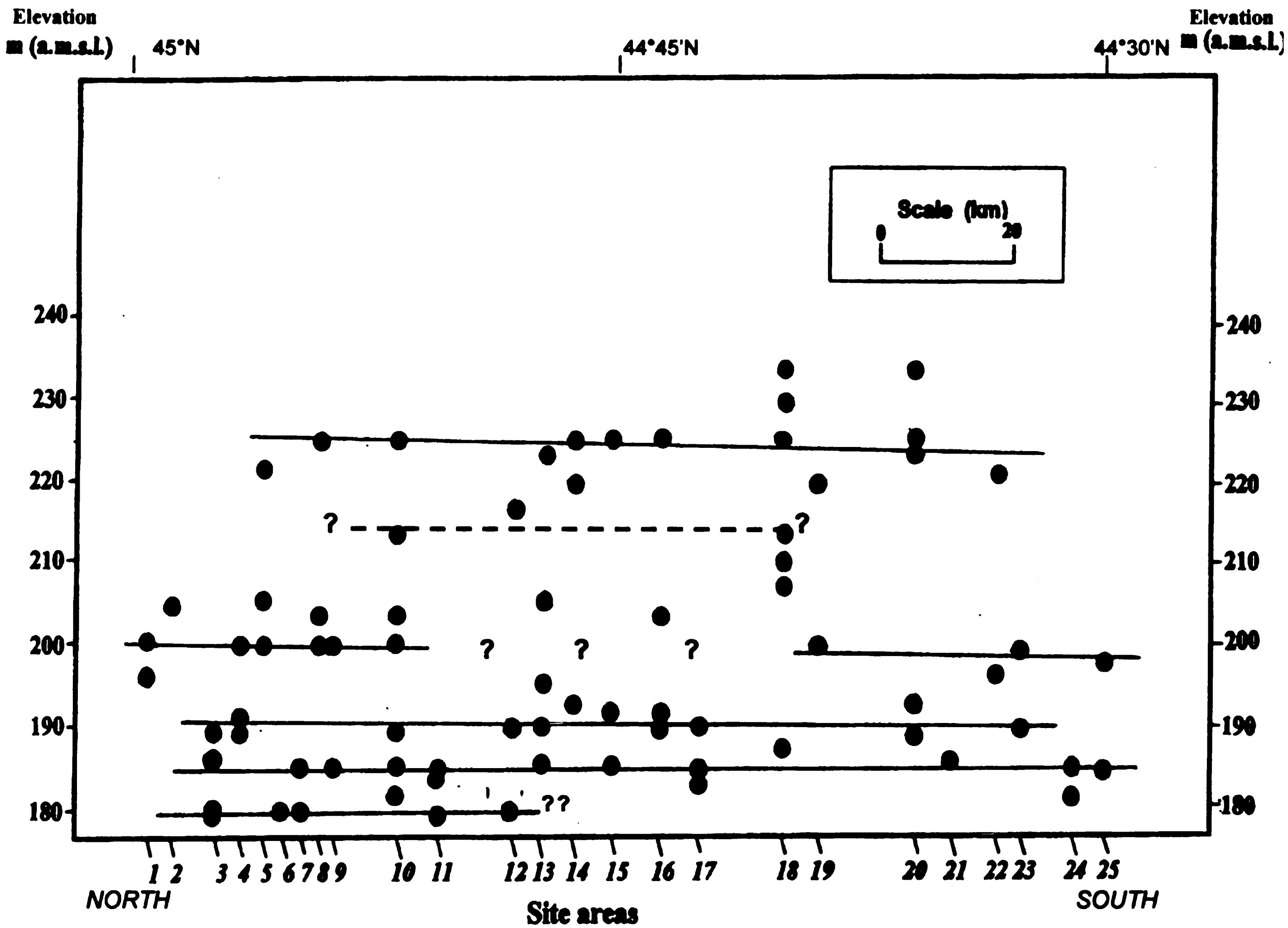
Lake Level History

A series of lake-levels at the Lakeshore (Table 3, Figure 30) are represented by morphological landforms as well as lacustrine sediments. The landforms include well-developed wave-cut scarps (Eyles and Eyles, 1983; Larsen, 1985), lacustrine plains (Bretz, 1951; Hansel et al., 1985; Taylor, 1990), erosional terraces (Leverett and Taylor, 1915; Larsen, 1985; Taylor, 1990), stream terraces (Bretz, 1963; Evenson, 1973; Larsen, 1985; Taylor, 1990), and delta surfaces (Calver, 1947; Taylor, 1990). Sedimentological evidence includes lake sediments (Evenson, 1973; Shaw and Archer, 1978; Eyles and Eyles, 1983; Hansel et al., 1985), beach/dune sand (DuBois, 1978; Clifton and Dingler, 1984; Colman et al., 1994), stream terrace deposits (Calver, 1947; Bretz, 1951; Fraser et al., 1990; Rovey II and Borucki, 1995), and deltaic sediments (Gilbert, 1890; Morgan, 1970; Wright, 1977).

The highest strandline in the study area occurs between 222 to 225 m elevation (Figure 30) and is represented by wave-cut scarps traceable for nearly 2 km on the north and east-facing slopes at Alligator Hill (Figures 5, 26) just northwest of Glen Lake, for about 1 km on the westward and northeastward slopes of North Manitou Island, for nearly 4 km around Lake Manitou at North Manitou Island, and for about 2 km on the eastward facing slopes of South Manitou Island. It is also represented by stream terraces near the west-facing slope of Miller Hill ridge (Figure 5) at about 222 m elevation, along the eastern rim of the Glen Lake lowland at about 222 to 224 m elevation and on the north facing slope of the Empire ridge in the Empire lowland at about 223 to 225 m elevation. In addition, two well-developed terraces, roughly graded to each other (a photograph of one is shown in Figure 24), also occur in the Empire channel (Table 3/Figure 30: Site 16) at about 225 m elevation and another well-developed terrace occurs near the Honor delta complex in the Platte River channel (Site 19) at about 220 m elevation. Lastly, a broad terrace occurs on the north-facing slope of the PBM (Site 18) also at about 225 m elevation.

This strandline is illustrated in Figure 30 by a roughly horizontal line that dips slightly southward and is drawn through points representing the above elevations taken around the Lakeshore. The line extends almost 100 km along the Lakeshore area, is nearly 48 m higher than the present level of Lake Michigan (177 m elevation), and its continuity and extent suggests the level of a single lake.

A lower strandline also occurred between 213 to 220 m elevation (Figure 30) and is mainly represented by poorly-developed wave-cut scarps along some of the westward-facing slopes of North Manitou Island (Site 10), Honor delta (Site 18), broad planar surfaces at about 213 to 218 m elevation, as well as by an extensive, planar surface (erosional? delta?) located along the southern side of the PBM just east of Beulah (Site 20) at about 215 to 220 m. This strandline is poorly developed at the Lakeshore and its localized nature mostly in the Platte Lake and Crystal Lake lowlands may indicate that it developed within an interconnected lake that had once existed there.



Evenson (1973)
Glenwood:
N/A (201 m?)
Calumet:
189 m
Taylor (1990)
Glenwood (II?):
235 m
Calumet:
213 m
Algonquin:
190 m
Nipissing:
183 m
Clark et al. (1994)
Glenwood:
Thin ice – 260 m
Thick ice – 390 m
Algonquin:
Thin ice – 190 m
Thick ice – 210 m
Nipissing:
Thin ice – 186 m
Thick ice – 191 m

Figure 30. Height/distance plot of lake level strandline elevations at the Lakeshore. Also shown are shoreline elevations from Evenson (1973), Taylor (1990), and Clark et al. (1994). Numbers on x-axis refer to survey sites listed in Table 3 (see map). Solid lines signify well-developed strandlines and dashed lines signify moderately to-poorly developed strandlines.

An even lower strandline occurs between 198 to 201 m elevation (Figure 30) and is represented by well-developed wave-cut scarps along some of the westward-facing slopes and ridges for 0.5 to 1 km near Frankfort and Elberta, as well as several high, flat, lacustrine plains located mostly within lowlands associated with Little Traverse Lake, Lime Lake, Glen Lake, Platte Lake, Little Platte Lake, and Crystal Lake, or near channels, and along the margins of the Manitou Islands. This strandline extends almost 100 km along the Lakeshore (Figure 30), is about 25 m higher than the present level of Lake Michigan, and its relative continuity suggests the level of a single lake or several lakes. However, the water-plane for this lake-level is not continuous along the entire Lakeshore and is either poorly developed or not present for nearly 50 km between Glen Arbor (Site 12) and Honor (Site 18). Between these two sites, scarp elevations deviated from the 198 to 201 m water-plane by about ± 5 m. One possible explanation for this gap may be that the area between Site 12 and Site 18 includes two broad lowlands (Empire and Platte Lakes lowland-Figure 5) that extend several km inland. The scouring energy from wave-action would have been diminished over this distance and therefore would have been ineffective in producing a defined scarp.

A still lower strand occurs at 189 to 192 m elevation (Figure 30) and is represented by wave-cut scarps along parts of the southern slope of South Manitou Island (Figure 25), well-developed ridges of beach/dune deposits developed upon lake-bottom deposits in the lowlands around Glen Lake, Little Traverse Lake, Platte Lake, Little Platte Lake, and Crystal Lake, and storm wave lag gravel along some of the scarps around Empire (Site 16) and Beulah (Site 20). It is also represented by lacustrine sediments (*Glen Lake deposit*) that occur in the lowlands along the Lakeshore, particularly in the Empire area (Site 16) where the deposit extends to the base of a steep, north-south trending slope just east of Empire, around the present margins of Little Glen and Glen Lakes, and along the northern boundary of the Platte Lake lowland at 191 m. In addition, it is also represented by an arcuate beach/dune ridge (*beach/dune complex*) around Little Traverse and Lime Lakes, is just over 2 km in length, overlies Glen Lake deposits, and separates the two lakes from one another. This strandline extends almost 100 km along the Lakeshore area (Figure 30), is nearly 15 m higher than the present level of Lake Michigan, and provides evidence for a lake that once occupied the Lake Michigan basin.

The next lower strand occurs at 184 to 186 m elevation (Figure 30) and is represented by an extensive lacustrine plain (*Glen Lake deposit*) on the mainland and on the Manitou Islands (Sites 10, 11), well-developed ridges of beach-dune sediments on the mainland and on the islands, and fossil-rich sands at the easternmost end of the Pyramid Point bluff. It is also represented by lacustrine deposits (*Glen Lake deposit*) in the lowlands along the Lakeshore, and is especially prominent where the lake sediments skirt along the northern margin of Alligator Hill (Site 15), along the northeastern margin of Pyramid Point near Shell Lake (Site 7), and along the flanks of ridges that occur around Onkeonwe Beach (Site 21). It is likewise represented by lacustrine benches along the eastern side of South Manitou Island (Site 11) that extend for nearly 3 km, and near the eastern margin of the Platte Lakes lowland around Honor (Site 18). Also near this elevation are two well-developed, extensive ridges of sand (beach/dune complex? spits?) about 1 to 1.5 km long that extend across isolated basins located on North Manitou Island (Site 10). Associated with these deposits at about 184 to 186 m are numerous curvilinear ridges of sand and gravel (*beach/dune complex*) which vary from 0.25 to just over 1 km in length that are superimposed on lacustrine sediments (*Glen Lake deposit*). Finally, fossiliferous sands occur at Pyramid Point near Hidden Lake (Site 7) at about 185 to 186 m.

This strandline extends almost 120 km along the Lakeshore area (Figure 30), is about 9 m higher than the present level of Lake Michigan, and the continuity of this strandline provides evidence for the minimum elevation of a lake that once occupied the Lake Michigan basin.

The lowest strand occurs at 179 to 181 m elevation (Figure 30) and is represented by beach/dune deposits and modern beach deposits that occur in the lowlands partially occupied by Glen Lake near Glen Arbor (Site 12), Little Traverse Lake (Site 3), and up to Empire near Otter Creek (Site 17). The strandline represented by beach/dune ridges is less well-developed south of Site 17 around Platte Lake, Little Platte Lake (Sites 18, 19), Crystal Lake (Sites 20, 21), and Lower Herring Lake (Site 24). The beach/dune ridges at the 179 to 181 m elevation form mostly curvilinear shapes that vary both in width and extent. However, they are usually less than 100 m wide and 0.5 km long and become more fragmented as they approach the modern shoreline, probably due to reworking by storm waves. On the other hand, the modern beach deposit extends along large areas of the shoreline at the Lakeshore. Oftentimes, the beach width is thinnest along exposed headlands such as along Pyramid Point, Sleeping Bear Point, and Empire bluff where the beach is about 1 to 6 m wide, and is widest in semi-sheltered bays along the shore such as along Good Harbor Bay, Sleeping Bear Bay, and Platte Bay where the beach is about 60 to 150 m wide.

This strandline extends along the Lakeshore for 60 to 70 km (Figure 30) and is almost 4 m higher than present. Its relative continuity provides evidence for the level of a lake that also occupied the Lake Michigan basin. However, as pointed out earlier, this strandline is not ubiquitous along the Lakeshore and is generally best represented in protected bays such as within Good Harbor Bay and Sleeping Bear Bay. One possible explanation for the gradual disappearance of the 179 to 181 m strandline south of Brooks Lake (Site 14) is likely due to reworking of nearshore deposits by storm waves along unprotected regions of the shoreline.

Discussion

Table 4 summarizes the elevations and interpretations of five paleolake strandlines recorded at the Lakeshore. The 222 to 225 m strand is believed to be a pre-Algonquin lake and probably represents a late Calumet stage of glacial Lake Chicago that probably formed at about 11,800 yrs B.P. (Hansel et al., 1985; Schneider and Hansel, 1990) as a consequence of the northward retreat of the Greatlakean ice margin. The continuity of this strand and the fact that it is the highest supports the argument that it formed in a large, open, proglacial lake and suggests that during the retreat of the Greatlakean ice margin, the Calumet stage of glacial Lake Chicago extended into the northern Lake Michigan basin.

Table 4. Lakeshore paleostrandline elevations and interpretations.

Late Pleistocene glacial lakes:
1. **222 – 225 m**: a series of levels representing the main Calumet stage of glacial Lake Chicago (to a pre-Lake Algonquin stage?).
2. **198 – 201 m**: a second Calumet stage of glacial Lake Chicago or a pre-Lake Algonquin stage.
3. **189 - 192 m**: Lake Algonquin (Main?).
Early-middle Holocene post-glacial lakes:
4. **184 - 186 m**: Lake Nipissing.
5. **179 – 181 m**: Lake Algoma.

The presence of a Calumet stage of glacial Lake Chicago at the Lakeshore significantly conflicts with interpretations made by earlier workers. For example, as shown in Figure 31, this Calumet strand or level at 222 to 225 m is nearly 32 m higher than the level suggested by Evenson (1973) and about 10 m higher than the level suggested by Taylor (1990). Evenson based his Calumet level on a flat water-plane from the Indiana-Michigan border to Muskegon, Michigan fixed at 189 m elevation. Taylor (1990) argued that the 210 to 215 m elevation of the Honor delta as well as planar surfaces located near Beulah (located only about 4 km to the southwest) were representative of the Calumet level of glacial Lake Chicago.

Also in the area around Beulah, Taylor attributes a well-developed terrace (NE¼,Sec.14,T.26N.,R.15W.) with an elevation of about 235 m on top of the PBM to the Glenwood II stage of glacial Lake Chicago. He also identifies a terrace located over on the northeast side of Crystal Lake along the south flank of the PBM (NE¼ ,Sec.15,T.26N.,R.15W.) and concluded that it too represents the Glenwood II stage of glacial Lake Chicago. Taylor (1990) correlates the highest planar surface (about 230 to 235 m elevation) of the delta at Honor (Figure 22) to the Glenwood II stage because of elevational similarities with the aforementioned terraces around Crystal Lake (pg. 108).

Taylor's (1990) Glenwood II and Calumet levels at the Lakeshore were probably produced by an interconnected terminoglacial lake that existed during the retreat of the Greatlakean ice margin. This lake, as Calver had first recognized (1947), once occupied the Platte Lake-Crystal Lake lowland, was about 40 m deep, likely fluctuated, and eventually equilibrated at 235 m in elevation (highest?) to form a strandline. Evidence in support of this interpretation includes the fact that the delta at Honor (Figures 22, 23) and terrace landforms around Crystal Lake are well preserved, show no evidence of erosion, and are localized to the lowlands (Calver, 1947). Their preservation and lack of eroded surfaces at the 235 m elevation suggest that these landforms were built mostly during the breakup of Greatlakean ice or nearly 1,000 years later than the age suggested by Taylor. Furthermore, since the landforms are localized to Platte and Crystal Lake (Figure 30, Sites 18, 20 respectively), they cannot represent the Glenwood II or Calumet stage of glacial Lake Chicago.

In addition to producing a lake in the Platte Lakes/Crystal Lakes lowlands, the Greatlakean ice-margin eventually retreated out of the lowlands along the Lakeshore and produced a series of several cascading lakes. These lakes formed a chain, as illustrated in Figure 34, from Lake Leelanau southwest to Crystal Lake over a distance of 30 km, and drained from northeast to southwest along the ice margin. Specifically, meltwater drained westward along the ice-margin from Lake Leelanau along a wide channel into the ice-blocked Little Traverse/Lime Lake lowland, then along another channel that narrowed into a southerly-bending drainageway cut between two ridges and eventually flowed into the Glen Lake lowland (Figure 34), which was also blocked by ice. Water then drained out of the lake occupying the Glen Lake lowland through a narrow gorge cut into the Sleeping Bear ridge and meandered along the margin of the ice filling the Empire lowland. This drainage eventually flowed into the large terminoglacial lake occupying the Platte Lake lowland (Figure 34).

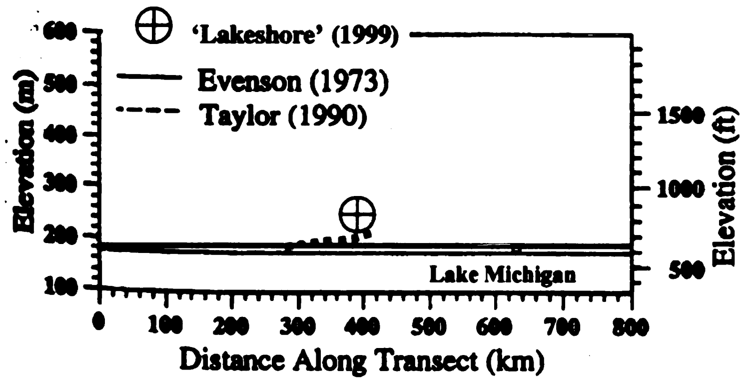


Figure31. Comparison of the Lakeshore Calumet shoreline elevation (1999) to data of Evenson (1973), and Taylor (1990).

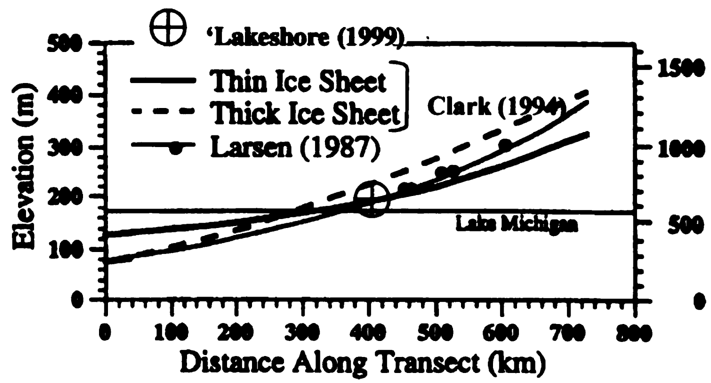


Figure 32. Comparison of the Lakeshore main Algonquin shoreline elevation (1999) to predicted data of Clark et al. (1994) and to data of Larsen (1987).

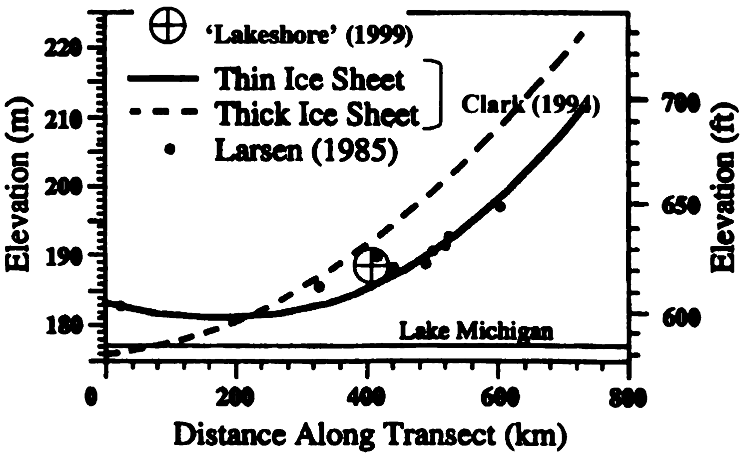
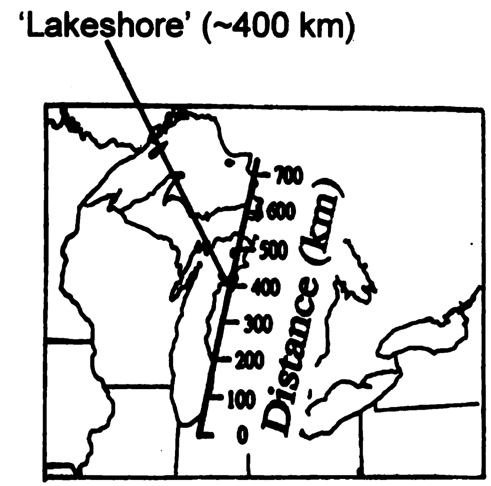


Figure 33. Comparison of the Lakeshore Nipissing shoreline elevation (1999) to predicted data of Clark et al. (1994) and to data of Larsen (1985).



The centerpiece for the ice-marginal drainage at the Lakeshore is the Empire channel (Figure 34) that once connected Glen Lake to Platte Lake. This channel is shown in Figure 34 (*stippled*). It is a well-defined, meandering, erosional landform that is entrenched mostly in collapsed outwash (*undifferentiated sand and gravel deposit*). Also within this channel lies a pair of stream terraces whose concordant elevations at about 225 m elevation (Table 3: Site 16) suggest that water flowing through the channel was at least 5 m deep. Imbricated clasts in the terrace deposit indicate that the main direction of flow was from north (Glen Lake) to south (Platte Lake).

Drainage from the Platte Lake lowland terminoglacial lake flowed through a narrow channel cut into the PBM and into the terminoglacial lake occupying the Crystal Lake lowland (Figure 34). Morphological evidence for this lake level (highest?) include Taylor's "Glenwood II-age" (1990) strandlines and terraces at about 235 m elevation (Table 3: Site 20) as well as two large, planar surfaces, also interpreted by Taylor (1990) as Glenwood II in age, located just east of Beulah.

During this period of ice-marginal drainage at the Lakeshore, the main body of the Honor delta was being developed. The Honor delta (*Honor sand and gravel deposit*) was built primarily by the Platte River as it discharged sediment-concentrated water from sources to the east into the terminoglacial lake occupying the Platte Lake lowland. The delta is slightly unusual in that sections of it consist of planar bedded to shallowly-dipping (≤ 20°) sand and gravel facies (Figure 20) with laminated-to-wavy clay beds (Figure 23). Brodzikowski and van

Loon (1991) point out that these textural changes may occur because barriers surrounding the terminoglacial lake (such as "ice masses and irregular topography"; pg. 111) can cause disturbances affecting the zonal buildup of the deltas, such as the supply of water and debris, and ultimately the water level. Consequently, factors like these would make it impossible to distinguish between topsets, foresets and bottomsets in any deltaic sequence.

Continued ice retreat out of the lowlands along with decreases in water input via ice-marginal drainage channels into the Platte Lake-Crystal Lake lowlands subsequently caused a drop in lake level and resulted in a cutoff between Platte Lake and Crystal Lake. Lastly, complete ice retreat out of the lowlands caused waters from the terminal lakes to escape and equilibrate with the 222 to 225 m Calumet level of glacial Lake Chicago.

The 198 to 201 m strand that occurs at the Lakeshore is believed to represent a pre-Algonquin lake level or possibly a low Calumet stage of glacial Lake Chicago (Table 4) formed as the Greatlakean ice margin retreated northward sometime between 11,800 yrs. B.P. and 11,200 yrs. B.P. (Hansel et al., 1985; Schneider and Hansel, 1990; Larson et al., 1994). The drop in level from about 225 m elevation to about 201 m elevation is perhaps due to the deglaciation of the Indian River lowland (Hansel et al., 1985). This interpretation is shared by Larsen (1987) and Taylor (1990) who suggest that the Calumet levels in the northern part of the Lake Michigan basin were highly unstable and short-lived and probably experienced other sharp drops just prior to the main Algonquin level.

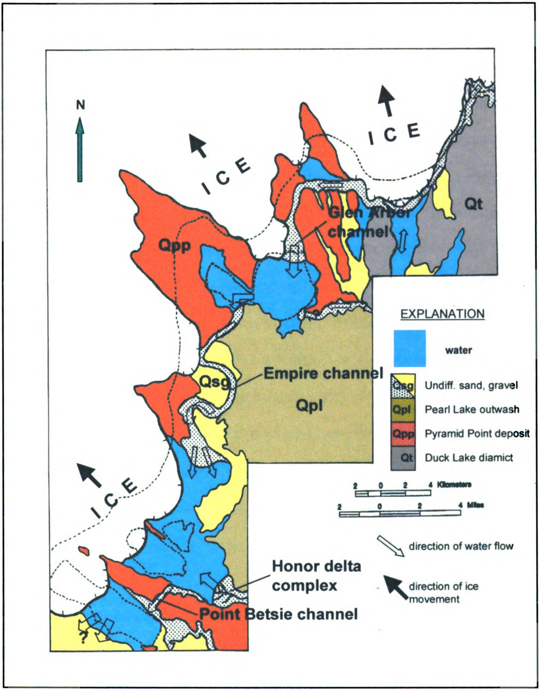


Figure 34. Mature stage of Greatlakean ice retreat/stagnation with formation of a series of large cascading lakes partially or completely ice blocked (11,800 yrs-10,700 yrs B.P.). Main channels or drainageways are shown by a stippled pattern. Arrows show probable direction of drainage.

The 189 to 192 m strand at the Lakeshore is believed to represent Main Lake Algonquin (Table 4) which formed sometime between 11,500 yrs. B.P. and 11,200 yrs. B.P. (Larsen, 1987). Main Lake Algonquin developed as a consequence of retreat of the Greatlakean ice margin north of the Straits of Mackinac (Larsen, 1987; Schneider and Hansel, 1990; Taylor, 1990; Larson et al., 1994). The 189 to 192 m strand at the Lakeshore (Figure 32) corresponds well with the Main Lake Algonquin strand suggested by Larsen (1987) and Taylor (1990), as well as that predicted by Clark et al.'s thin-ice model (1994).

The 184 to 186 m strand at the Lakeshore is believed to represent Lake Nipissing (Table 4) which formed sometime after 9,000 yrs B.P. (Larsen, 1985). Lake Nipissing formed when the outlet at North Bay, Ontario was progressively isostatically raised to an elevation above that of the southern outlet at Port Huron (Larsen, 1985), which resulted in a rise of lake level elevation close to that of earlier Lake Algonquin between 9,000 yrs and 5,000 yrs B.P. (Larsen, 1985). This event likely marked the establishment of the Nipissing Great Lakes whose level, at first, was controlled by three outlets; Chicago, Port Huron, and North Bay (Hansel et al., 1985; Larsen, 1985). By comparison, the 184 to 186 m strand at the Lakeshore (Figure 33) is within one or two meters of the Nipissing strand reported by Larsen (1987), about three meters higher than the Nipissing strand reported by Taylor (1990), and one to three meters of the Nipissing level predicted by Clark et al.'s thin-ice model (1994).

The 179 to 181 m strand at the Lakeshore is believed to represent Lake Algoma (Table 4) which formed between 4,700 to 2,500 yrs B.P. (Larsen, 1985) due to incision of the Port Huron outlet and abandonment of the outlets at Chicago and North Bay (Hansel et al., 1985). The 179 to 181 m strand is within 0.5 meter of the Algoma strand for the northern Lake Michigan basin reported by Larsen (1987). Following Lake Algoma, lake level at the Lakeshore steadily equilibrated due to incision resulting in only about a 2 m drop in the last several thousand years to form present day Lake Michigan (177 m elevation).

Correct placement of glacial and post-glacial lake levels at the Lakeshore has been difficult and inconclusive mostly due to the lack of datable material in the lacustrine deposits. Also, inaccuracies on lake level elevations are unavoidable using mostly 5 m contours drawn on 7.5" topographic maps. Such errors could be greatly reduced with the use of modern survey equipment such as a GPS receiver or a Total Station. Nonetheless, morphologic and scattered sedimentologic evidence does suggest that a series of lakes did at one time exist at the Lakeshore, that the oldest level there was at about 222 to 225 m, and likely represents the Calumet stage of glacial Lake Chicago.

CONCLUSIONS

Over the last several decades, there have been numerous studies of the glacial and post-glacial history of the Sleeping Bear Dunes National Lakeshore region. Three fundamental questions, however, remain unresolved: 1) how far did Greatlakean ice advance into the area, 2) what is the origin of the ridges at the Lakeshore, and 3) what glacial lake stage followed the retreat of the ice margin.

Detailed surficial mapping of the Lakeshore (Figure 12) shows the region to be characterized by a westward sloping drumlinized till plain partially buried by a complex of glaciolacustrine and glaciofluvial deposits, as well as by littoral and eolian sediments. Analysis of the map suggests the following:

1) The Duck Lake diamict represents the Port Huron ice advance and is probably a subglacial till. In addition, it is clear that deposition of Duck Lake diamict predates deposition of both the Pyramid Point and Pearl Lake deposits. This would imply that the Pyramid Point and Pearl Lake deposits, as well as most of the Duck Lake diamict, must have been formed prior to the Greatlakean advance and probably represent deposits associated with Port Huron-age ice.
2) There is no evidence for the Greatlakean ice advance at the Lakeshore.
3) Greatlakean ice probably invaded into the lowlands and was likely constrained by high Port Huron-age topography.

4) The ridges are most likely subaqueous crevasse fillings (*Pyramid Point deposit*) formed during the retreat/stagnation of the Port Huron ice sheet sometime during the Two Creekan Substage.
5) Paleolake-levels are represented by a series of strands: the strand at 222 to 225 m elevation probably represents the Calumet stage of glacial Lake Chicago; the strand at 198 to 201 m elevation probably represents a low Calumet stage of glacial Lake Chicago; the strand at 189 to 192 m elevation probably represents the Main Lake Algonquin stage; the strand at 184 to 186 m elevation probably represents Lake Nipissing; and the strand at 179 to 180 m elevation probably represents Lake Algoma.

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