This is to certify that the

thesis entitled

GEOMORPHIC HISTORY OF SAN DUNES
AT PETOSKEY STATE PARK, PETOSKEY MICHIGAN

presented by

Xiomara Cordoba Lepczyk

has been accepted towards fulfillment
of the requirements for

M.S. degree in Geography

Major professor

Date June 19, 2001

GEOMORPHIC HISTORY OF SAND DUNES AT PETOSKEY STATE PARK, PETOSKEY, MICHIGAN

By

Xiomara Cordoba Lepczyk

AN ABSTRACT OF A THESIS

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

MASTER OF SCIENCE

Department of Geography

2001

Professor Alan F. Arbogast

ABSTRACT

GEOMORPHIC HISTORY OF PETOSKEY STATE PARK DUNE FIELD, PETOSKEY, MICHIGAN.

By

Xiomara D. Cordoba

The goals of this study were to investigate the geomorphic evolution of the dune field at Petoskey State Park and to make recommendations for park management based on this geomorphic history. The area of study was characterized into 5 distinctive units with distinct landforms and geomorphic histories: lake terrace, massive parabolic dunes, inset dunes, linear and incipient parabolic dunes, and active dunes. Evidence from this is study suggests that these geomorphic units become progressively older towards the east. Radiometric evidence (radiocarbon and optically stimulated luminescence) suggested that these landforms developed in response to fluctuations of Lake Michigan during the middle and late Holocene after the Nipissing highstand, around 4800-4400 yrs. B.P. Results from this study indicate that, at times, several geomorphic units were active synchronously during the past 4000 yrs. Furthermore, this study suggests that foredune sequences develop landward from low sloping nearshore areas, while parabolic dune fields form landward from moderate to high sloping nearshore areas.

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

Chapter 1

INTRODUCTION

Coastal dunes are common features along Lake Michigan's shoreline and are highly sensitive to natural and anthropogenic disturbances. Studies of sand dunes along the Lake Michigan shoreline have largely concentrated on dune ecology and/or foredune development in the southern part of the basin (Cowles, 1899; Calver, 1946; Olson, 1958a, 1958b, 1958c, Thompson, 1992; Thompson and Baedke, 1995, Thompson and Baedke, 1997). However, a small number of studies have been conducted on the processes and causes associated with parabolic dune formation at scattered sites along the shoreline (Arbogast and Loope, 1999; Loope and Arbogast, 2000). These studies are important contributions to our understanding of the processes controlling cycles of eolian activity along the Lake Michigan shoreline; however, they are regional in scope and the history of individual dune fields has not been studied in detail. Within the context of previous and current research on Lake Michigan dunes, the present study focuses on the geomorphic history of a small dune field in northwestern lower Michigan.

An additional aspect of this research is associated with the utilization of dune landscapes for recreational purposes. Numerous state parks and recreation areas along Lake Michigan occur on dune fields

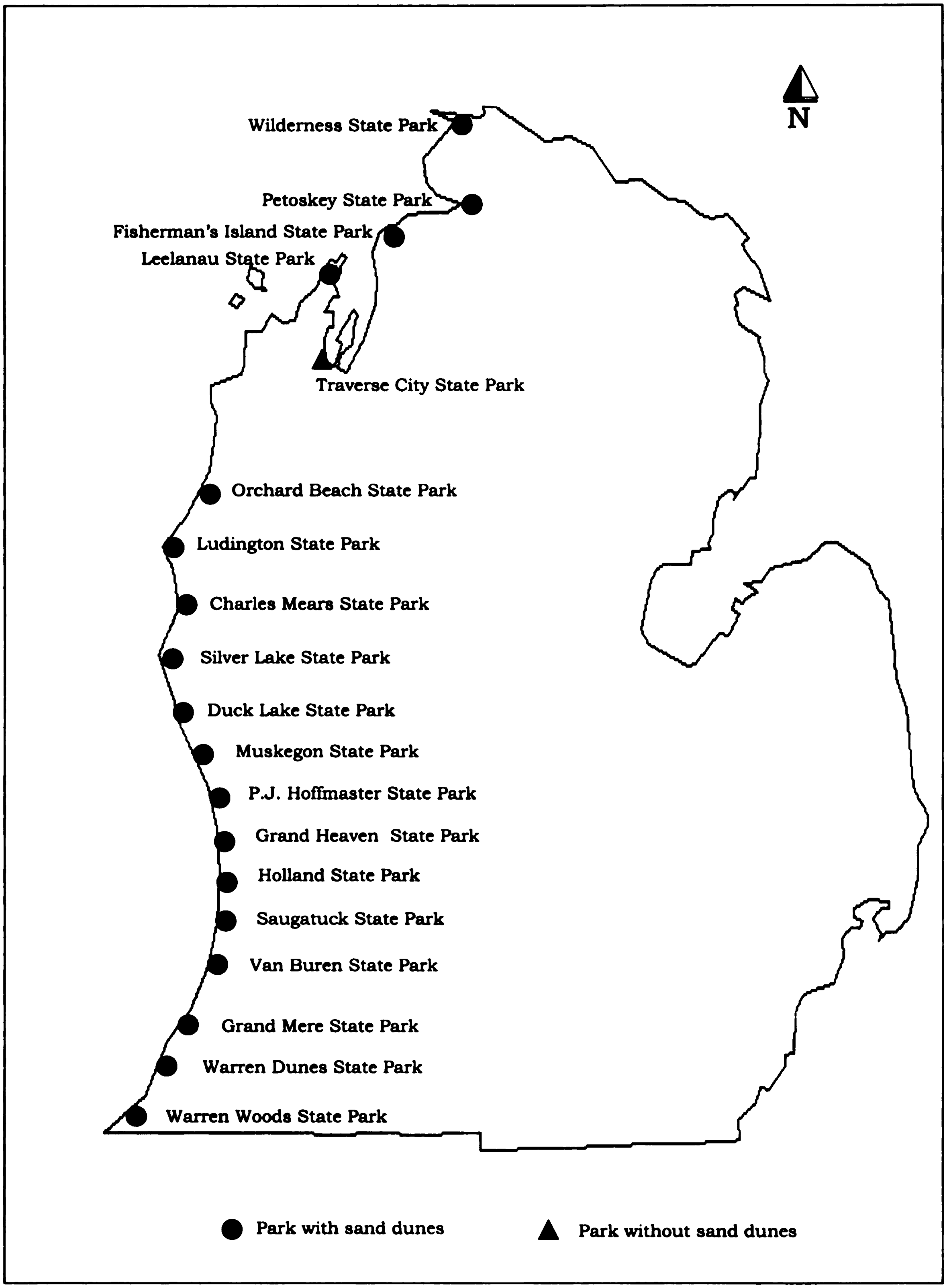

Figure 1:1. Location of state parks along the shoreline of Lake Michigan. From MDNR (2000).

(Figu

durir

with

curr

vege

othe

or ca

(Fig

cam

of N

an a

Suc

acti

of s

of s

effc

car

per

ma

eol

unc

(Figure 1:1). Michigan's residents heavily utilize these areas, especially during the summer months. Such utilization may create serious conflicts with the high sensitivity of dune landscapes. On one hand, if dunes are currently stable, heavy recreational use may destroy or fragment the vegetation cover, hence leading to reactivation of sand dunes. On the other hand, if dunes are currently active, infrastructure, such as roads or campgrounds may become progressively buried by migrating sand.

An example of such conflicting needs occurs at Petoskey State Park (Figure 1:1), where active dunes are encroaching upon a small adjacent campground. The Parks and Recreation Division of Michigan Department of Natural Resources (MDNR) regularly implements diverse measures in an attempt to control the effects of eolian activity on the campground. Such mitigation measures include planting of stabilizing dune grass on active surfaces, construction and maintenance of sand fences, placement of signs to discourage park users from climbing the dunes, and removal of sand from the Dunes Campground, among others. Despite the strong efforts by park management, eolian activity has continued and several campsites have recently been closed due to burial by sand (Tom Maclean, personal communication).

Given the existing conflicts, MDNR recently indicated a need to make management decisions that compromise human needs and active eolian processes. In order to achieve this goal, it was determined that an understanding of the history of eolian activity is required. Hence, this research comprises a comprehensive study of the Late Holocene geologic history of the park, focused on ancient and current eolian processes. In order to achieve this goal a detailed mapping of geomorphic units was conducted and radiometric dating was used to establish the chronology of eolian events.

The results from this research provide a valuable management tool for the MDNR because it outlines the nature and magnitude of dune building cycles that have occurred within the park area. Furthermore, the investigations presented here fit within the framework of recent research along the Lake Michigan shoreline, and are a significant contribution because they focus on the geomorphic history of a particular dune field. As a result, this research can be used to test recent models (Arbogast and Loope, 1999; Loope and Arbogast, 2000) about coastal dune evolution.

Chapter 2

REVIEW OF RELEVANT LITERATURE

COASTAL SAND DUNES

Understanding the nature of eolian processes is fundamental to the study of sand dunes. This section thus provides a theoretical background to the dynamics of coastal sand dunes. Carter (1988) defines sand dunes as eolian landforms that develop where the transporting competence of the wind is impaired, which implies that dune formation is usually related to the interaction of the air stream with a barrier such as vegetation.

Most coastal dune fields occur on gently sloping coastal platforms (old sea or lake terraces) created by coastal erosion, tectonic subsidence, or pre-existing river valleys (Livingstone and Warren, 1996). They are found above the high water mark on sandy beaches and occur worldwide in ocean, estuarine, and lake shorelines, but are most common on dissipative coasts with strong onshore winds and abundant sand size particles (Carter, 1988; Carter et al., 1990). The sources of sand for coastal sand dunes are varied and include sediments from rivers, coastal erosion, and offshore supplies in relatively shallow shelves (e.g. glacio-fluvial plains below base level).

The most important factors in coastal dune formation are sediment type, sediment supply, presence of winds with speed above the sediment's entrainment threshold, presence/absence of vegetation, and topographic position removed from wave activity (Carter, 1988; Sherman and Hotta, 1990). In addition to these factors, local controls include topography, nature of waves, tidal range, and base level change (Carter, 1988). Short and Hesp (1982) suggest that foredune development is most substantial on flat, dissipative shorelines where large volumes of sand are stored subaerially at low water levels. A high tidal range also contributes to the development of coastal dunes, for it exposes an extensive intertidal area, which constitutes a major source of sand. In the long term, base level changes, sediment availability, and climatic change have direct effects on the formation of coastal dunes (Carter, 1988).

Coastal dunes form from sand delivered to the shoreline by waves and currents (Carter, 1988, Easterbrook, 1990). If such sediments are subsequently dried and subaerially exposed, they become vulnerable to aerodynamic processes such as entrainment and sand transportation (Goldsmith, 1978). Carter (1988) stated that these processes are physically similar in water and in air, although water is approximately 1000 times denser than air. The interaction of the wind with the surface has a retardation effect and establishes a velocity gradient that leads to turbulence in the lower airstream and to the formation of an

a
s
i
s
b
s

t
a

aerodynamic boundary (Schenk, 1990). The velocity profile near the surface in conditions of neutral stability (when no thermal mixing occurs) is assumed to be logarithmic and depends on the elevation at which wind speed is zero (roughness length), the displacement height (aerodynamic boundary - usually assumed to be at the sand surface), and the shear stress (Sherman and Hotta, 1990).

Sand particles start to move when the shear stress exceeds a threshold value, which is equal to the forces opposing motion (gravity and cohesive agents such as moisture or chemical precipitates). As the shear velocity exceeds the threshold for motion, transport of sediment begins in the form of suspension, saltation, or traction load (or surface creep; Figure 2:1).

Suspension involves fine materials such as silt and clay. During true suspension the turbulent motion of the air transports grains for large distances (>100 km). The second wind transport mode, saltation, is probably the most important of the wind mobilizing processes in terms of the volume of sand displaced (Carter, 1988; Sherman and Hotta, 1990). Saltation is initiated by forces of aerodynamic lift due to a negative pressure on top of the sand grains caused by the Bernoulli effect (Schenk, 1990). After the grain has been lifted, it is decelerated by gravity and wind drag, which causes the aerodynamic trajectory seen in Figure 2:1. When saltating sand grains land, they dislodge further grains through ballistic impact. This impact causes the newly disloged grains to move even when shear velocity is below the initial threshold, resulting in large volumes of grains transported. The air density is increased due to the sediment load, which shifts the bed stress from the ground to the surface on top of the saltating mass, causing decay and settling. When sand grains are too large to be lifted by the wind they creep or roll intermittently along the bed as saltation of finer grains occurs. The size of the grains that are transported as surface creep at any specific time depends on the wind velocity. Surface creep is commonly initiated by impact of saltating particles.

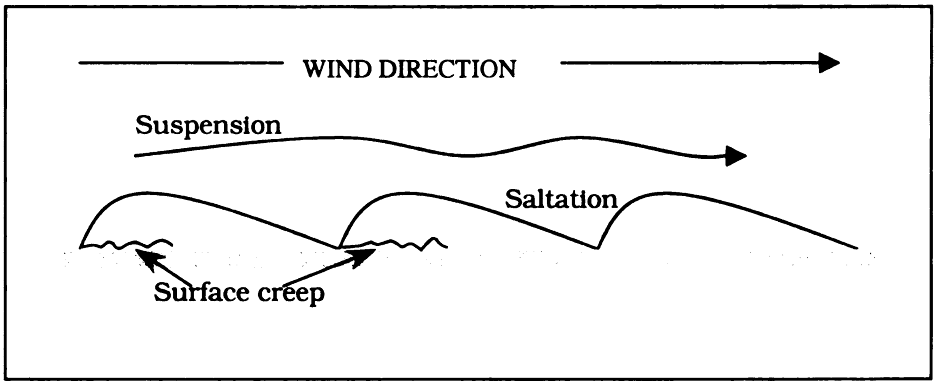

Figure 2:1. Mechanisms of sediment transport by wind (From Schenk, 1990).

The presence of obstacles, such as vegetation and pre-existing topography, favors deposition of wind-transported sediments in coastal environments because it perturbs the flow of wind, creating a sheltered area. Therefore, the pattern of wind flow over dunes depends on their three dimensional shape: (1) flow over linear, continuous dunes is subject to little compression, and (2) sharp crest lines cause acceleration of the wind velocity (Carter, 1988). Once established, a dune field may distort the pattern of wind to the point that shoreline parallel winds are common in areas where large foredunes occur, leading to significant volumes of sediment moving along the beach (Carter, 1988).

The processes of entrainment and transport of sand by wind are affected by external factors such as moisture content and presence of chemical precipitates in the sand, which have a bonding effect that raises the shear stress required to mobilize the sand. This makes wet core dunes difficult to move and establishes a basis for the common belief that massive coastal dune formation occurs during low base levels, when the water table is low as well. However, wet core dunes may move, and their mobility depends on temperature, albedo, solar inclination, and slope aspect, especially within irregular topography (Hyde and Wasson, 1983). David (1978) proposed that the presence of moisture is an important factor in the development of parabolic dunes, an aspect that will be discussed later in this chapter.

Textural variations along the beach also affect the transport of sediments by wind. Pavements, for example, increase the roughness of the surface, make saltation easier (because grains bounce further), and seal underlying sediments from further deflation (Carter, 1988). Grain shape and density also affect transportability since angular grains are harder to entrain, but move farther once airborne, while denser grains are harder to move than lighter ones.

The presence of vegetation also constitutes an important factor in the formation of coastal sand dunes. According to Sherman and Hotta (1990) most coastal dunes form in the presence of vegetation because the interaction between the wind and vegetation causes an abrupt gradient in the surface shear stress that leads to sedimentation. The most rapid sedimentation occurs on the downwind side of the vegetation. Moreover, a dense vegetation cover creates a false ground surface, elevating the aerodynamic boundary and leaving a layer of still air below it that is ideal for sedimentation. This elevated aerodynamic boundary, known as *zero plane displacement*, is located at about two-thirds of the vegetation height (Carter, 1988). Additionally, the morphology of dunes is strongly controlled by vegetation. According to Short and Hesp (1982), linear dunes may form when the vegetation cover is approximately 90-100%, while other forms (such as barchans or sand waves) form when vegetation is sparse (0-20% vegetation cover).

Changes in base level, whether eustatic or local, are an important factor in shoreline behavior and development of coastal dunes. Dunes in northwestern Europe, for example, formed during the Late Pleistocene and early Holocene, when sea level was about 120 m below its present level (Carter, 1988). During periods of base level rise large amounts of unconsolidated sediments are available for eolian erosion and extensive

r
s
b
f
l
c
c
c

r
c

l
(
c
r
e
a
r
p
(

a

mobile dunes may form, especially if the sediment supply is high and sand is not vegetated. If the newly formed foredunes are soon colonized by vegetation and new material is constantly reworked, one single linear foredune will probably form (Carter et al., 1990). During a drop in base level sediment from the shelf is fed into the wave base, which may result on formation of extensive beaches and dune ridge plains over thousands of years. A prograded series of dunes may result of this process if the dunes are stabilized by vegetation (Carter, 1990).

Coastal sand dunes can be classified according to their morphology or their genetic mechanisms (Goldsmith, 1978). The most commonly used terminology is that based on dune morphology (McKee, 1979; Figure 2:2). Six types of dunes exist under this classification: barchan, parabolic, transverse, longitudinal, dome, and blowout dunes (Goldsmith, 1978; Schenk, 1990). Barchan, barchanoid, and transverse dunes have only one slip face and are formed under unidirectional wind regimes. Dome dunes do not have slip faces and are considered an embryonic type of dune. Transverse dunes generally have two slip faces and are indicative of a bi-directional wind regime. Star dunes have multiple slip faces and form in areas with complex wind regimes. Parabolic dunes are crescent shaped with the slip face convex downwind (Easterbrook, 1999).

In coastal areas where vegetation exits, dunes are mostly blowout and parabolic. Blowouts are erosional features that can be described as

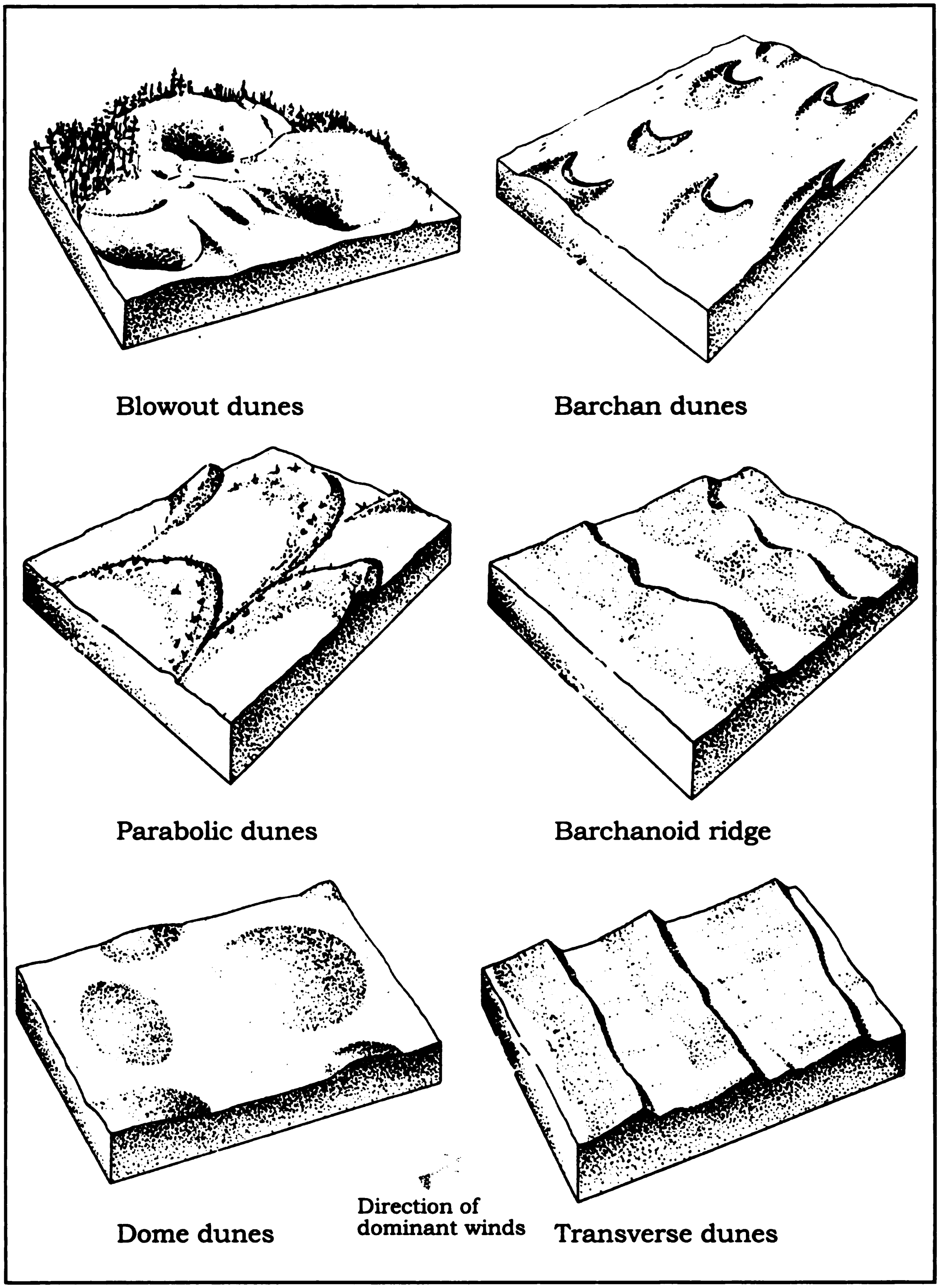

Figure 2:2. Dune forms produced by unidirectional wind regimes. (From McKee, 1979)

hollows, depressions, troughs, or swales within dune complexes. The most common way in which blowouts initiate is associated with acceleration of the wind velocity due to several possible factors: (1) shoreline erosion or washover during storms, (2) loss of vegetation cover at a particular location because of animals or particularly strong winds, and (3) human activities inducing erosion (Carter, 1988; Livingstone and Warren, 1996). Blowouts are usually formed in areas where large linear dunes stabilized by vegetation exist after a change in climate or a decrease in sand suply (especially in coastal situations).

The initial development of blowouts may be in the form of non-depositional areas or gaps in incipient foredunes, and occurs preferentially on the higher parts of the dunes due to their vulnerability to desiccation and disturbance. In some areas, blowouts may be indicative of a negative sand budget on the dune (Livingstone and Warren, 1996).

Carter (1990) identified two types of blowouts: saucer, and trough. Saucer blowouts are "shallow, ovoid, dish shaped hollows with a steep marginal rim and a flat convex downwind depositional lobe" (Carter, 1990). Trough blowouts are deep, narrow, steep sided, have more pronounced downwind depositional lobes, and mark deflation basins. Trough blowouts commonly evolve into parabolic dunes.

Once initiated, blowouts evolve through a series of mechanisms. Erosion of the blowout surface may lower the axis of the blowout and crea

winc

velo

max

wall

and

the

free

ava

blo

tra

blo

bas

Fu

an

an

wi

di

bl

(C

bl

create a funneling effect that in turn increases the rate of erosion. The wind flow is accelerated in blowouts and jet flow is common. The highest velocity occurs at the axis of the deflation basin and shear stress is maximum at the base of the trough and along the steepest part of the walls. Additionally, increased instability of its walls leads to slumping and avalanching. Erosion of the deflation basin removes support from the walls of the blowout and instability occurs (Carter, 1988).

The sidewalls of blowouts are composite slopes. The upper part is a free face in the vegetation root zone. Underlying the root zone lies a loose avalanche incline. Sediments accumulate in this zone before reaching the blowout floor, where it is reworked by the wind. Sediment can also be transported along the walls by oblique ripple migration. The growth of blowouts in coastal dunes may be limited by a lag concentrate at the base of the blowout or by erosion down to the water table (Carter, 1988). Furthermore, the widening of the blowout may limit the funneling effect and consequently, wind erosion. At this time, vegetation may colonize and stabilize the blowout (Livingstone and Warren, 1996).

Blowout orientation is a commonly used indicator of the prevailing wind directions since the long axis of blowouts is usually parallel to the direction of the prevailing winds (Jugerius et al., 1991). However, blowout orientation may also be controlled by pre-existing topography (Carter, 1990). Jugerius et al. (1991) studied the development of blowouts in coastal sand dunes in the Netherlands and found that wind velocities between 6.25 and 12.5 m/s are the most erosive. An experiment conducted by Jugerius et al. (1991) indicated that wind velocities above 12.5 m/s shift large volumes of sand but do not necessarily result in net erosion within the blowout. Jugerius et al. (1991) used these findings to conclude that the orientation and rate of erosion of blowouts is determined by the direction and magnitude of the most frequent winds rather than by infrequent, strong winds.

Parabolic dunes are also important in coastal regions. They are U or V shaped and their arms point upwind. They commonly occur in deserts, in vegetated cold climates like the central and midwestern United States, coastal areas, and stabilized dune fields (Livingstone and Warren, 1996). They can be classified according to their length: width ratios: less than 0.4 are lunate, 0.4-1.0 are hemicyclic, 1.0-3.0 are lobate, and greater than 3.0 are elongate (Pye, 1990). The most accepted theory of parabolic dune formation is that they evolve from blowouts: the dune grows as sediment is supplied from underlying sediments; eventually erosion ceases, the advance of the apex stops and the dune is colonized by vegetation. A parabolic dune will form if the dune movement is enough to separate it from the original ridge (Livingstone and Warren, 1996). Vegetation has a very important role in parabolic dune formation since it protects the vegetated arms from erosion and perpetuates the dune's shape. David (1978) proposed a different theory of parabolic dune formation that involved the existence of a wet core dune. According to

D
in
s
w
e
in

I

t
V
s
I
i
t
e
b

s
T
F
a
y

David (1978) the presence of ground moisture at the base of the dune increases the shear stress of the sand and causes an increase in the shear strength of the local air stream, promoting erosion on the windward side of the dune. However, David's (1978) model fails to explain how the parabolic form is initiated as well as the reasons for irregular distribution of moisture along linear ridges.

LATE QUATERNARY GEOLOGY

The present landscape in northwestern lower Michigan is largely the result of geomorphic processes active during and since the Late Wisconsin glaciation. The Quaternary glaciations in the Midwest were strongly influenced by the pre-existing topography of the proto-St. Lawrence drainage, resulting in an intricate lobate configuration of the ice margin (Mickelson et al., 1983). Evidently, as glaciers advanced through the Great Lakes region, ice preferentially eroded along the pre-existing valleys underlain by shale, creating the basins that would later become the Great Lakes (Farrand and Eschman, 1974).

In addition to the formation of the Great Lakes, the Laurentide ice sheet deposited thick deposits of glacial drift (Rieck and Winters, 1993). The earliest recorded glacial deposits in Michigan are found near Grand Rapids, where the John Ball Park till occurs (Eschman, 1985). Zumberge and Benninghoff (1969) dated these beds and reported an age of 51,000 yrs BP. Nevertheless, the record of Early Wisconsin and pre-Wisconsin glaciations is not clear in most parts of Michigan because deposits of these events, if preserved, underlie the thick cover of middle and late Wisconsin till (Eschman, 1985; Rieck and Winters, 1993). According to Rieck and Winters (1993) the thickness of glacial drift in lower Michigan may reach over 285 m, with the area around Petoskey State Park having a drift thickness of approximately 100 m.

The chronology of the Late Wisconsin glaciation (Table 2:1) has been determined mainly on the basis of radiocarbon dates and stratigraphic relations among different till units (Eschman, 1985; Farrand and Eschman, 1974). Within the Late Wisconsin several major fluctuations of the ice margin have been recognized (Table 2:1). However, the geomorphology of the area of study is the result of the Greatlakean advance, middle and late Holocene lake level fluctuations, and eolian processes (Farrand and Bell, 1982).

Table 2:1. General chronology of the Late Wisconsin glaciation as it applies to the Petoskey area. (Modified from Dreimanis and Karrow, 1972).

Holocene	
	10,000 BP
Greatlakean (Valderan) Substage	11,850 BP
Twocreekan Interstadial	
Port Huron Substage	13,000 BP
Pre - Port Huron	

After retreating from the Port Huron moraine, lake levels dropped significantly in the Michigan basin due to the opening of an eastward outlet during the Twocreekan interstadial. The next glacial advance, the Greatlakean, overran two plant communities, the Two Creeks forest and the Cheboygan bryophyte bed. Both of these localities have been radiocarbon dated (Farrand, 1969; Miller and Benninghoff, 1969, Larson et al., 1994) and used to establish the chronology of events between the Port Huron and Greatlakean advances.

The Greatlakean advance (formerly referred to as Valders advance), occurred about 11,850 years. ago (Larson et al., 1994) and is characterized by a red, clayey till (Farrand and Eschman, 1974). It was first reported in lower Michigan by Leverett and Taylor (1915), but the term Greatlakean was not introduced until Evenson et al. (1976) studied the red tills in Wisconsin and northern lower Michigan.

Melhorn (1954), who mapped the distribution of the red tills in lower Michigan, conducted the first study on the characteristics and extent of these tills. In Emmet County, Melhorn (1954) described a complex of morainal hills that extend from Little Traverse Bay to the Straits of Mackinaw. The moraines reach altitudes between 216 and 335 m with the highest parts occurring towards the west. These moraines were interpreted by Leverett and Taylor (1915) as an interlobate track between the Huron and Michigan lobes formed after the Port Huron moraine, but probably synchronous with the Manistee moraine. The best exposures of the red till in lower Michigan occur in Emmet county according to Melhorn (1954). He described the till there as a "tough, sandy, stony clay till ranging from light pink to medium red in color" (Melhorn, 1954; page 64). The percentage of clay in the till and its color vary spatially. The color of the till changes from pink to reddish-brown to light reddish-brown to light brown in a southward and eastward direction (Schaetzl, 2001). Several authors have used the distinctive color of the till to identify the limits of this advance. However, its lateral extension is still under debate because additional chronological control is needed (Schaetzl, 2001).

After the retreat of the Greatlakean ice, lower Michigan became ice-free and a sequence of postglacial lakes developed (Evenson et al., 1976). Glacial Lake Algonquin was the first of these lakes, forming as ice retreated and the Straits of Mackinaw opened (Martin, 1957; Eschman and Karrow, 1985). At this time the Huron and Michigan basins became confluent. Karrow et al. (1975) dated sediments in the Lake Michigan basin and concluded that the Main Algonquin phase probably occurred between 11,200 and 10,400 BP. As deglaciation opened new outlets the level of Glacial Lake Algonquin decreased and a series of progresively lower lakes developed (Hansel et al., 1985; Schaetzl et al., 2001)). Much controversy exists about how many identifiable lake levels occurred during the Holocene between Lake Algonquin and Lake Nipissing. In a general sense, the post Algonquin lakes have been classified into the Upper and the lower Group. Around Petoskey State Park, four shorelines corresponding to these levels have been reported: the highest Algonquin beach (213 m), the lowest beach of the Upper Group (204 m), Main Battlefield (203 m), and Fort Brady (199 m) (Leveret and Taylor, 1915). Eschman and Karrow (1985) proposed that the post-Algonquin lakes were probably short lived and that the whole sequence only took a few hundred years.

When an outlet opened at North Bay, Ontario, due to deglaciation, lake levels dropped significantly in the Michigan basin. This level, named the Chippewa low by Leveret and Taylor (1915), was extremely low (approximately 61 m below the present level) and initiated around 10,000 BP (Eschman and Karrow, 1985; Hansel et al., 1985). Subsequently, lake levels rose during the Nipissing transgression in the middle Holocene. This transgression coincided with the abandonment of the North Bay outlet and the end of the middle Holocene Hypsithermal (the transition from a dry, warm climate to cooler, moister conditions; Hansel et al., 1985).

Lake Nipissing drained initially through three outlets: North Bay, Chicago, and Port Huron, but the North Bay outlet was soon abandoned (Hough, 1958). At this time, the Michigan, Huron and Superior basins were confluent. Radiometric dates indicate two distinctive Nipissing peaks: Nipissing I at 4,500 yrs. BP, reached an altitude of 183 m, and the Nipissing II, which reached a maximum level of 180 m about 4,000 yrs. BP (Hansel et al., 1985). According to Larsen (1985) the Chicago outlet was abandoned at the end of the Nipissing II phase and the Saint

Cl

Hu

Ho

lev

ele

ar

la

Su

ha

ge

th

le

oc

be

la

Th

flu

Th

as

le

Claire River at Port Huron became the main outlet for the Superior, Huron and Michigan basins around 3,400 yrs. BP. For the Petoskey area Hough (1958) reported and elevation of 186 m for the Nipissing terrace.

Incision of the Port Huron outlet caused lake levels to drop to the level of Lake Algoma. The terrace from Lake Algoma occurs at an elevation of approximately 183 m around Petoskey State Park (Leveret and Taylor, 1915). Continued incision of the Port Huron outlet caused lake levels to drop to their present levels of Lakes Michigan, Huron, and Superior (Hansel et al., 1985). However, fluctuations of the lake levels have continued through the middle and late Holocene (Figure 2:3).

Larsen (1985) interpreted lake level history based on geomorphology, stratigraphy, and radiocarbon dates. His study shows that fluctuations of 200-300 years duration around the modern mean level have occurred in the past 3,000 years. Distinct high lake levels occurred about 1,600-1,200 BP, 950-750, 450 yrs. BP, and at the beginning of the 18^{th} century (Larsen, 1985).

Studies of beach ridges in the northern and southern shorelines of lake Michigan by Thompson (1992), Thompson and Baedke (1995), and Thompson and Beadke (1997) suggest that three quasi-periodic lake level fluctuations occur in Lake Michigan every 600-550, 150, and 30 years. The 600-yr. fluctuation, which is in the order of 1.8-3.7 m, is usually associated with progradation of the shoreline due to the slow rate of lake level change. The 150-yr. fluctuation (0.8-0.9 m) is associated with

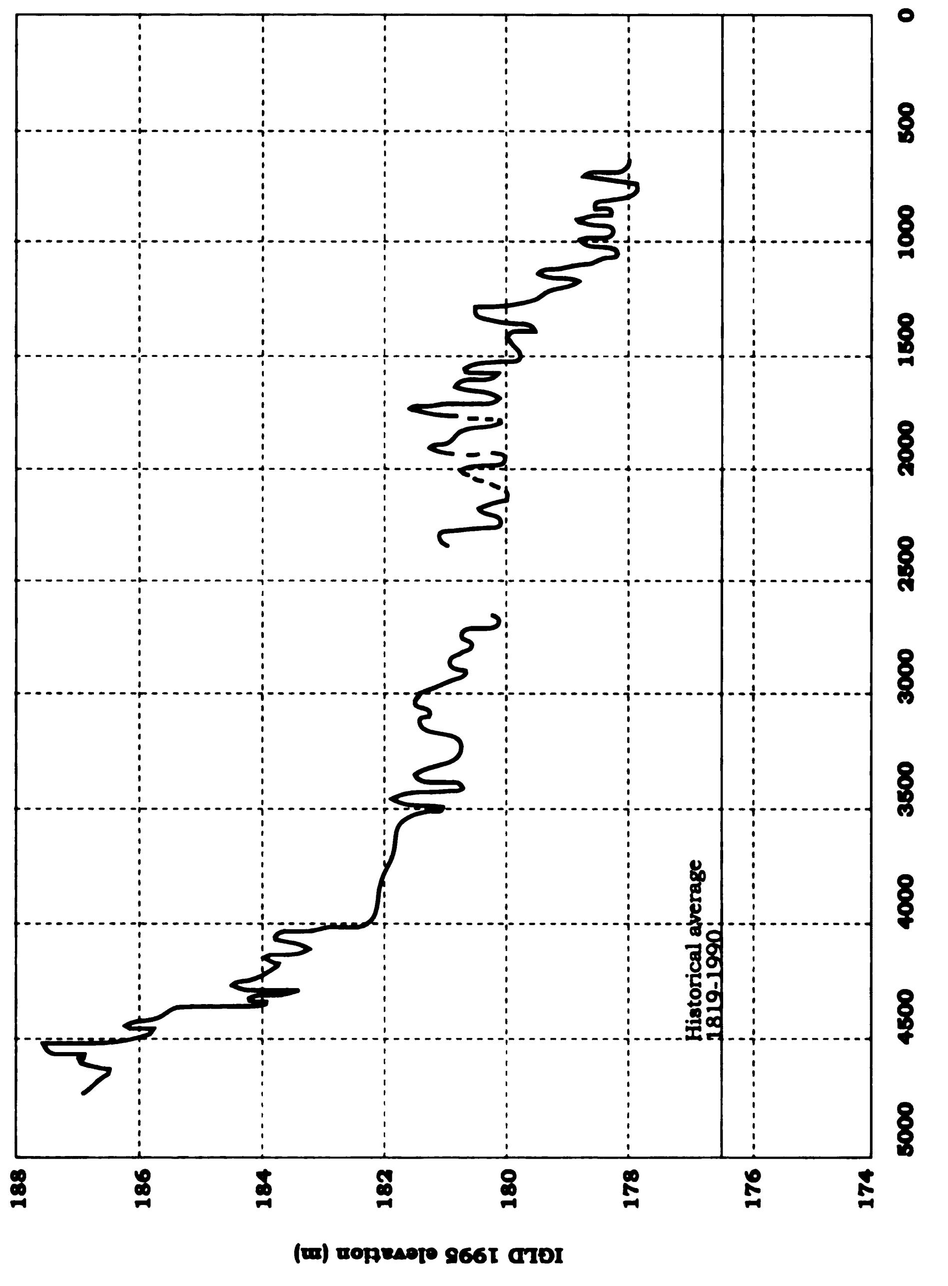

Figure 2:3. Relative late Holocene lake level for Sturgeon Bay, approximately 30 km north from Petoskey State Park. (From Thompson and Baedke, 1997)

parabolic dune building (Loope and Arbogast, 2000). The 30-yr. fluctuation (0.5-0.6 m) has the fastest rate of change and is associated with aggradation and progradation of the shoreline (Thompson and Baedke, 1995).

Lichter (1995) studied a beach ridge succession in Sturgeon Bay, MI. In that study he found a relationship between beach ridge formation and regional drought in the 5 younger beach ridges in the Bay. His chronology shows that at least 108 beach ridges formed in the Bay during the past 3500 yrs and that the rate of formation of beach ridges (average of one every 32 years) has been relatively uniform during this period of time.

The causes of late Holocene lake level fluctuations have been linked with changes regional climatic conditions (Hansel et al., 1985; Hansel and Mickelson, 1988; Larsen, 1985). The effect of climatic changes in lake levels is mainly attributed to changes in the effective precipitation entering the lakes (Hansel and Mickelson, 1988). Larsen (1985) compared the scale of changes in climate and lake levels during the past 2,000 years in the Lake Michigan basin and concluded that a relationship between these two factors exists: low levels occur during warm, dry periods, while high levels occur during cool, moist periods. However, other factors control post-glacial lake levels as well, including: (1) opening and closing of outlets during the early Holocene, (2) downcutting of outlets, (3) changes in the volume of water entering the basin, and (4) differential isostatic uplift (Hansel et al., 1985). Dune formation initiated as lake levels dropped and large volumes of sandy material became available for eolian transport along the coast of Lake Michigan (Hansel et al., 1985).

LAKE MICHIGAN SAND DUNES

Coastal sand dunes occur in numerous locations along the eastern and southern shorelines of Lake Michigan (Figure 2:4). This section focuses on results from previous studies conducted on Lake Michigan sand dunes.

Early studies of Lake Michigan dunes, conducted during the late 18th and early 19th centuries, were qualitative in nature and addressed issues related with dune ecology (Cowles, 1899; Calver, 1946; Olson, 1958a, b, c). Olson (1958a, b) conducted one of the most influential studies on vegetation successions on Lake Michigan dunes. According to Olson (1958a, 1958b) dune successions comprise pioneer, intermediate, and mature stages, with narrow ephemeral ecotones. Basinward dunes support pioneer vegetation that can survive constant sand input, while inland dunes usually support perennial vegetation, have some soil development, and receive little or no sand input. In some areas, whereprogradation consists of shoreline parallel bands (or foredunes), the succession shifts forward as progradation occurs.

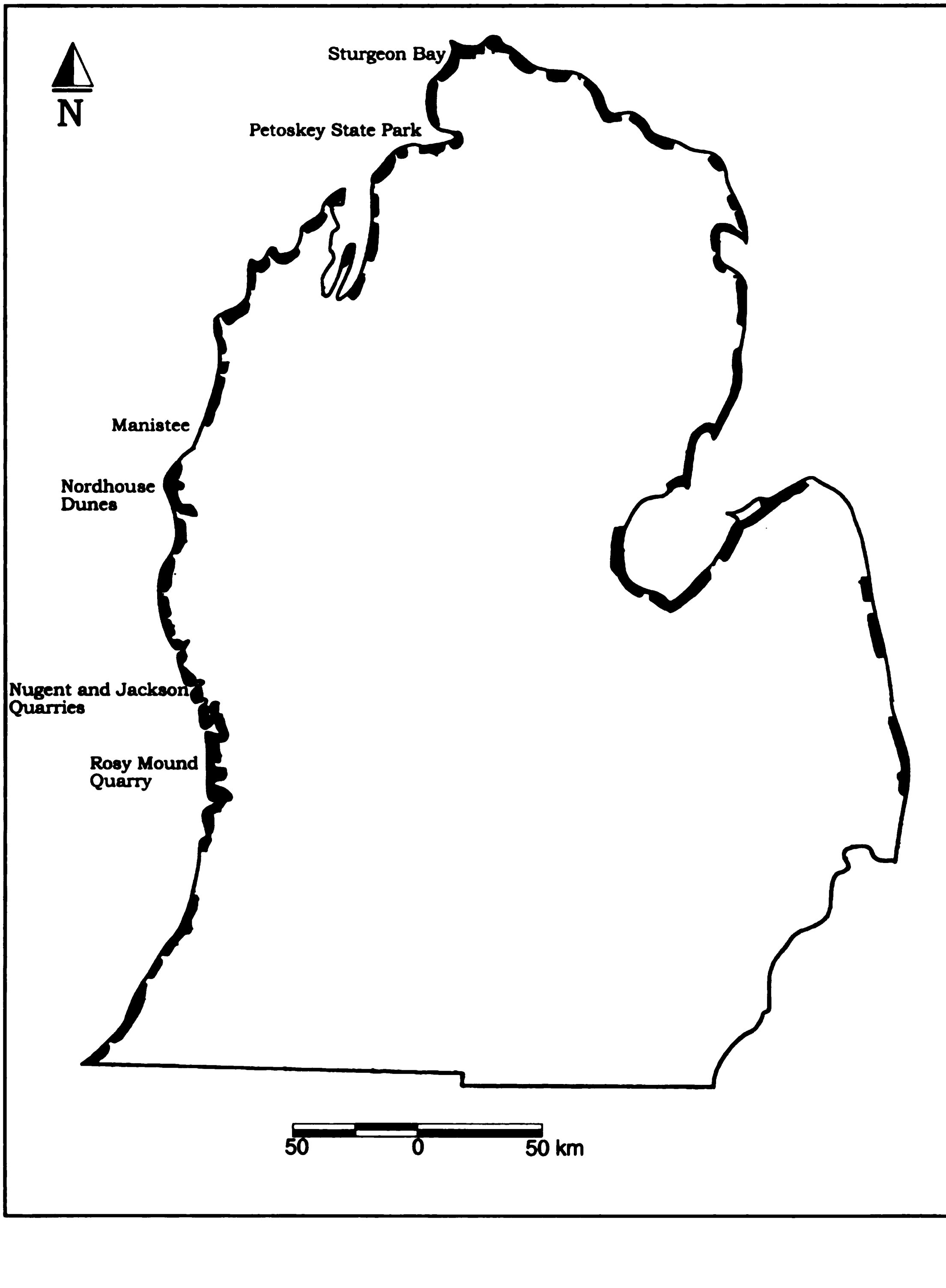

Figure 2:4. Location of coastal dunes in lower Michigan. (From Arbogast and Loope, 1999).

Olson (1958a, 1958b) described the succession of vegetation in southern Lake Michigan, where sand stabilization starts with american beach grass (*Ammophila breviligulata*) and sand reed grass (*Calamovifa longifolia*). Beach grass predominates in areas where rates of sand deposition are high, while sand reed grass tolerates lower sedimentation rates and eroding environments (Olson, 1958a; 1958b). When aggradation rates are lower than a few centimeters per year, sand cherry (*Prunus pumila*), choke cherry (*Prunus virniana*), grape (*Vitis spp.*), red ozier dogwood (*Cornus stonifera*) and bluestem bunchgrass (*Andropogon scoparius seplentrionalis*) slowly replace beach grasses. White and jack pines (*Pinus strobus* and *P. banksiana*) may colonize dunes within a couple of years of dune stabilization but only predominant on surfaces 90-600 years old. Hardwood communities replace these species within 100 years and dominate the subsequent composition of the forest (Olson, 1958b).

Lichter (1997, 1998) conducted a later study on the vegetation succession on Lake Michigan dunes. Lichter (1998) emphasized the role of vegetation in dune development and stated that plants are important in the formation and secondary development of coastal dunes because the type and extent of vegetation cover affect the susceptibility of dunes to erosion. Litcher (1998) studied the succession of vegetation in sand dunes in Sturgeon Bay. His surveys concentrated on dunes younger than 500 years because vegetation and ecosystem changes are more pronounced there. He found that in the younger beach ridge (< 40 years old) the vegetation assemblage was dominated by beach grass (*Ammophila breviligulata*), although other dune building species such as willow shrubs (*Salix spp.*) and sand cherry (*Prunus pumila*) also occurred. On older dunes (between 55 to 175 years) the percentage of beach grass decreased and was replaced by evergreen shrubs such as juniper (*Jiniperus communis*), bearberry (*Arctostaphylos uva-ursi*), and minor bluestem bunchgrass (*Schizachyrium scoparium*). This assemblage was followed by the development of a mixed pine forest after 225 to 440 years This mixed pine forest, composed of white pine (*Pinus strobus*), red pine (*Pinus resinosa*), white spruce (*Picea glauca*), balsam fir (*Abies balsamea*), white cedar (*Thuja occidentalis*), and paper birch (*Betula papyrifera*), developed between 225 and 440 years of dune stabilization.

In addition to ecological studies on coastal dunes, more recent research has concentrated on the factors that control shoreline behavior and dune evolution. In the Great Lakes three factors are believed to control the behavior of shorelines: wave climate, sediment supply, and lake level fluctuations (Thompson and Baedke, 1997). Sand dunes along Lake Michigan generally develop in locations protected from intense wave activity where sediment budget is positive. In northwestern lower Michigan, local sediment sources are unconsolidated glacial sediments, which are abundant throughout the region (Thompson and Baedke, 1995). Due to wave refraction, wave energy is concentrated on headlands, causing erosion. The sediments are subsequently transported along the shoreline by longshore currents and littoral drift (Goldsmith, 1978; Easterbrook, 1999). Wave energy in bays and embayments is dissipated, making them preferred areas of deposition (Thompson and Baedke, 1995). The resulting positive sediment budget allows the formation of sand dune complexes in many of these locations along the shoreline of Lake Michigan.

Numerous studies have been conducted on the evolution of small linear dune fields and their relationship to lake level fluctuations (Cowles, 1899; Olson, 1958a, b, c; Thompson, 1992; Lichter, 1995; Thompson and Baedke, 1995; Petty et al., 1996; Lichter, 1997 Thompson and Baedke, 1997). The most influential of the early studies on sand dune formation along Lake Michigan was that conducted by Olson (1958a, b, c), who focused on the development of shore - parallel ridges and their relation to lake level oscillations. According to Olson (1958c) these ridges are initiated as offshore bars that migrate inland during lake level fall. Once the bars are exposed subaerially, colonization by vegetation (especially dune building species) might occur and dune construction initiates. Olson's (1958a, b, c) works set the grounds for later research by developing the initial hypothesis that foredunes form in response to lake level fluctuations.

Later research along the Lake Michigan shoreline, conducted during the 1990s, concentrated on testing and further developing Olson's

(1958c) model (Thompson, 1992; Lichter, 1995; Thompson and Baedke, 1995; Petty et al., 1996; Lichter, 1997; Thompson and Baedke, 1997). Although these studies were related to foredune development, their emphasis was on the beach ridges that underlie the dunes. The ultimate goal of these studies has been to reconstruct Late Holocene lake level fluctuations in Lake Michigan.

Lichter (1995) studied a series of beach ridges in Sturgeon Bay, Michigan, to establish the periodicity of dune formation along Lake Michigan's shoreline and its relationship with Lake Michigan levels. Lichter (1995) dated roots and rhizomes from the base of beach ridges using accelerator mass spectrometry (AMS) and found that the currently active beach ridge (the one closest to the shoreline) formed during "a precipitous" recession of Lake Michigan level associated with a regional drought in the 1960s.

A series of influential studies conducted by Thompson (1992), Thompson and Baedke (1995), and Thompson and Baedke (1997) focus on the development of beach ridges in the Lake Michigan basin. The underlying assumption of Thompson and Beadke's (1992, 1995, 1997) work is that lake level fluctuations control vertical aggradation in coastal eolian systems. Based on this assumption, they linked processes of foredune formation with changes in Lake Michigan levels.

Thompson and Baedke (1995) clearly presented their conceptual model using the Curray (1964) diagram of shoreline behavior to explain these relationships (Figure 2:5). Their model assumes that abundant sediment supply and a location protected from storms exists. Foredune development initiates with early lake level rise, when the rate of change is slow. At this time, the shoreline experiences a depositional transgression if sediment supply remains constant. Erosion becomes the dominant process on the shoreline as the rate of lake level rise increases. Due to increased shoreline erosion a large volume of sediment becomes available and a phase of aggradation follows. At this time, the rate of transgression would probably decrease due to buffering of erosion by deposition and a nearshore berm and beach ridge could develop. The magnitude of the berm formed at this time depends on the occurrence of a high stillstand before lake level falls again.

When lake level drops, a depositional regression occurs and progradation of the shoreline takes place. At this time, pioneer vegetation may colonize the berm (Thompson and Baedke, 1995). Vegetation acts as a trap for eolian sediments and aggradation occurs creating a dune cap on the ridge (or foredune). A series of closely spaced beach ridges forms if sediment supply is high during lake level fall. On the other hand, a rise in lake level would cause erosion of one or more beach ridges. In this situation sediment eroded from the beach ridges becomes available for eolian transport and renewed aggradation of the closest inland dune would occur (Thompson and Baedke, 1995). The result of this process would be widely spaced beach ridges and potential localized truncation of the beach ridge progression by parabolic dunes (such as some areas of Baileys Harbor embayment, WI, and Sturgeon Bay, MI; Thompson and Baedke, 1997).

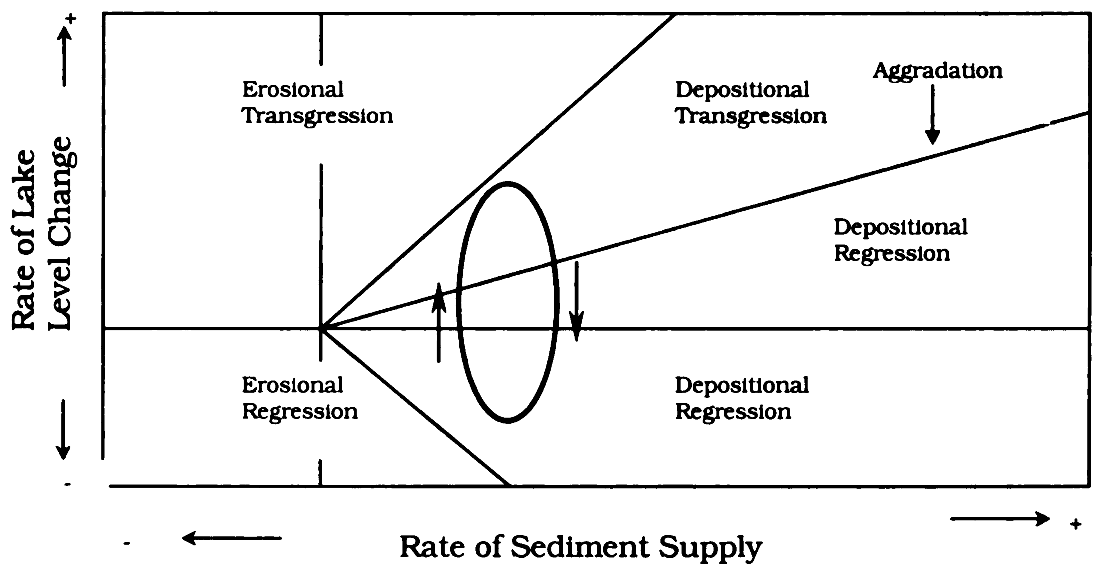

Figure 2:5. Diagram of shoreline behavior during fluctuating lake level. Oval represents development of a single beach ridge. (From Thompson and Baedke, 1995).

Thompson and Baedke's (1995, 1997) results agree with the studies conducted by Olson (1958c), who proposed that foredunes develop as a response to falling lake levels. Thompson and Baedke's (1995, 1997) model stands undisputed up to the present time. However, Delcourt et al. (1996) and Petty et al. (1996), who studied the formation of beach ridges in the southeastern Upper Peninsula of Michigan, argue that the ridges initially form as offshore bars that move landward as lake level falls, a conclusion that agrees with Olson's (1958c) model.

The most significant contribution of the recent studies conducted on strand-plains along the Lake Michigan shorelines was a conceptual model for the development of beach ridges and the overlying linear dunes (Olson, 1958c; Thompson, 1992; Lichter, 1995; Thompson and Baedke, 1995; Petty et al., 1996; Lichter, 1997, Thompson and Baedke, 1997). In summary, this conceptual model suggests that beach ridges and their eolian caps develop in response to lake level fluctuations: (1) beach ridges form during highstands, and (2) their eolian caps form during low lake levels, when wide beaches are exposed to eolian processes.

Although numerous studies have focused on dune ecology and foredune development, few studies have addressed the evolution of parabolic dune fields along the Lake Michigan shoreline. In addition, most of the studies that had been conducted were qualitative in nature. Scott (1942) referred to parabolic dunes as *secondary or deformed dunes* and stated that their development was associated with blowout formation. Furthermore, Scott interpreted the presence of parabolic dune fields in northern lower Michigan as indicative of lake level highstands. Dorr and Eschman (1970) referred to parabolic dunes as *high dunes* and suggested that coastal dunes along Lake Michigan formed during lake levels higher than the present, mostly during the Nipissing highstand, a conclusion that agrees with the hypothesis proposed by Scott (1942).

A later study conducted by Buckler (1979) on Lake Michigan sand dunes suggested that dunes along the Lake Michigan shoreline formed as the lake dropped from its glacial levels. According to Buckler (1979) the largest parabolic dunes along Lake Michigan formed during the regression from the Nipissing level, 3000-4000 years ago. At this time, unconsolidated Pleistocene sediments were theoretically exposed on wide beaches and reworked by winds, favoring vertical aggradation of massive parabolic dunes, in a fashion similar to Olson's (1958c) foredune model (Dorr and Eschman, 1970; Buckler, 1979). A more recent study conducted by Litcher (1995) in the Wilderness State Park, on Sturgeon Bay, suggested that parabolic dunes within the park developed during high lake levels, when shoreline erosion is intensive on headlands and large volumes of sediments become available on bays and embayments.

As a result of these studies two hypotheses developed about the formation of parabolic dune fields, which have opposing views: (1) Buckler (1970) hypothesized that they formed during the regression from the Nipissing highstand, and (2) Litcher's (1995) proposal that parabolic dunes developed during high lake levels. However, such hypotheses have never been rigorously tested in a chronostratigraphic framework and the relationships between lake level fluctuations and lake terrace dune development are unclear.

Recent studies by Arbogast and Loope (1999) and Loope and Arbogast (2000) focused on testing the opposing hypotheses of Lake Michigan parabolic dune formation (Buckler, 1970; Litcher, 1995) by studying lake terrace dunes along eastern Lake Michigan. More specifically, Arbogast and Loope (1999) tested the hypothesis that massive parabolic dune formation along the shoreline occurred during the Nipissing highstand (e.g. Dorr and Eschman, 1970), while Loope and Arbogast (2000) rigorously examined the relation between episodes of parabolic dune aggradation and fluctuating lake levels in the Lake Michigan basin.

In their initial study, Arbogast and Loope (1999) studied buried soils at the base of parabolic dunes at four sites (Nordhouse dunes, Nugent, Jackson, Rosy Mound quarries; Figure 2:4) on the east shore of Lake Michigan and established the chronology for the onset of parabolic dune formation. At these sites, active parabolic dunes overlie sandy lacustrine deposits. Based on radiocarbon dates and stratigraphic evidence (elevation of lake terrace), Arbogast and Loope (1999) concluded that the basal lacustrine unit probably accumulated during the Glenwood (14,100-12,700 yrs. B.P) or Calumet (12,700-11,000 yrs. B.P.; Leveret and Taylor, 1915; Hansel and Mickelson, 1988) highstands of Lake Michigan. Once lake levels dropped to the Chippewa level (between 10,000 and 7,500 yrs. B.P.), these lake terraces were subaerially exposed and soil development occurred. Eolian sand accumulation initiated at their sites as follows: (1) during the Nippising highstand at the Nordhouse dunes, (2) during a highstand between 4300 and 3900 cal. yrs. BP at the Nugent and Jackson quarries, and (3) during the Algoma highstand at the Rosy Mound Quarry.

Based on radiocarbon and stratigraphic evidence Arbogast and Loope (1999) concluded that the lakeward face of lake terraces are destabilized due to wave erosion as lake level rises. Erosion of the lake terrace provides sands that can be subsequently transported inland by wind, causing burial of soils and/or aggradation of pre-existing dunes. During lowstands the surface of the dunes is stabilized by vegetation and soil development occurs (Arbogast and Loope, 1999). The results by Arbogast and Loope (1999) agree with studies by Lichter (1995), that indicate that parabolic dunes in Sturgeon Bay formed during major lake level transgressions around 2375 yrs. BP and 3200 yrs. BP.

In a later study, Loope and Arbogast (2000) refined their model and tested several assumptions about the timing of development of parabolic dunes along Lake Michigan. Based on radiocarbon dates from 75 buried soils in active dune fields they determined that, contrary to the common assumption that dune landscapes in the eastern shore of Lake Michigan are of Nipissing age, these landscapes are primarily post-Nipissing landforms, with the majority of dune complexes being ≤ 1500 cal yr. BP. The most important contribution of this study was the hypothesis that that blowout development and parabolic dune aggradation occur in response to the 150 yr. lake level highstands identified by Thompson and Baedke (1997; Arbogast and Loope, 2000).

In summary, previous research suggests that coastal dunes begin to form during lowstands, when a wide beach is exposed to eolian action and sands are transported shoreward by the wind and subsequently trapped by plants, developing embryonic dunes. These embryonic dunes may coalesce into a continuous ridge (a foredune) above the upper reach of the waves (Litcher, 1998). During episodes of high lake level, when wave erosion is enhanced, the existence and extent of vegetation cover "determines whether dunes retain their linear nature or develop a blowout that would eventually form parabolic dunes" (Pye, 1990). Parabolic dunes, therefore, form during major lake level transgressions, which means they are indicative of more extreme wave and wind erosion than are beach ridges (Arbogast and Loope, 1999; Loope and Arbogast, 2000).

Chapter 3

STUDY AREA

Petoskey (pronounced Pet-ōh-skee) State Park is a small (2 km^2) park in Emmet County, along the east coast of Lake Michigan in northwestern lower Michigan (Figure 3:1). The park is located 5 km northeast of the city of Petoskey and 8 km southeast of the city of Harbor Springs. The study area is bounded by the highway M-131 to the east and south, Little Traverse Bay to the west, and an urban development to the north. In this part of Michigan sand dunes extend along the coast of Lake Michigan from Petoskey to the Straits of Mackinaw (Farrand and Bell, 1982). This thesis focuses specifically on the dunes within Petoskey State Park, coving an area of approximately 2 km^2.

In Petoskey State Park, active and inactive sand dunes occur up to 600 m. from the Lake Michigan shoreline. The largest dunes are generally farthest from the shore (Figure 3:2). These inland dunes are parabolic and vary in height between 30 and 40 m. The size of the dunes decreases with distance to the shoreline, with the smallest dune represented by a nearshore foredune that is about 2 m in height.

CLIMATE

The climate of the study area is strongly influenced by the proximity of Lake Michigan and the predominant westerly winds that

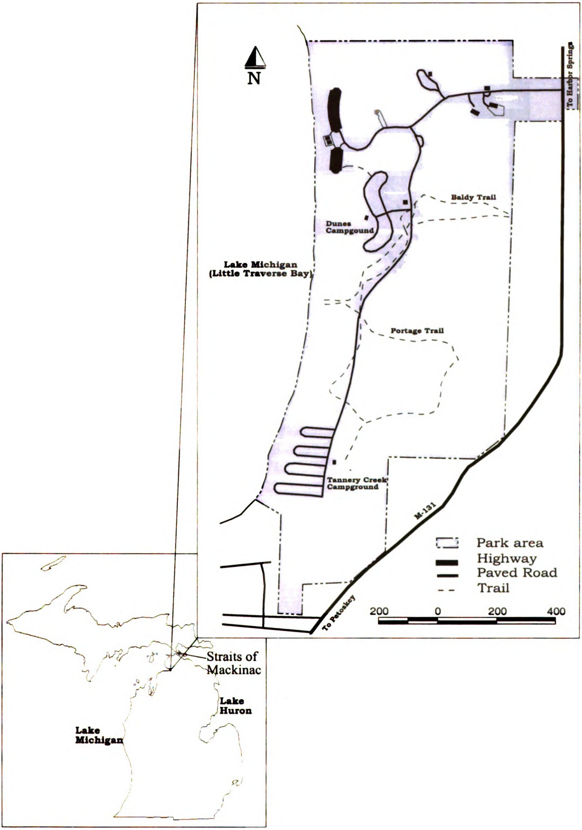

Figure 3:1. Location of Petoskey State Park, Michigan.

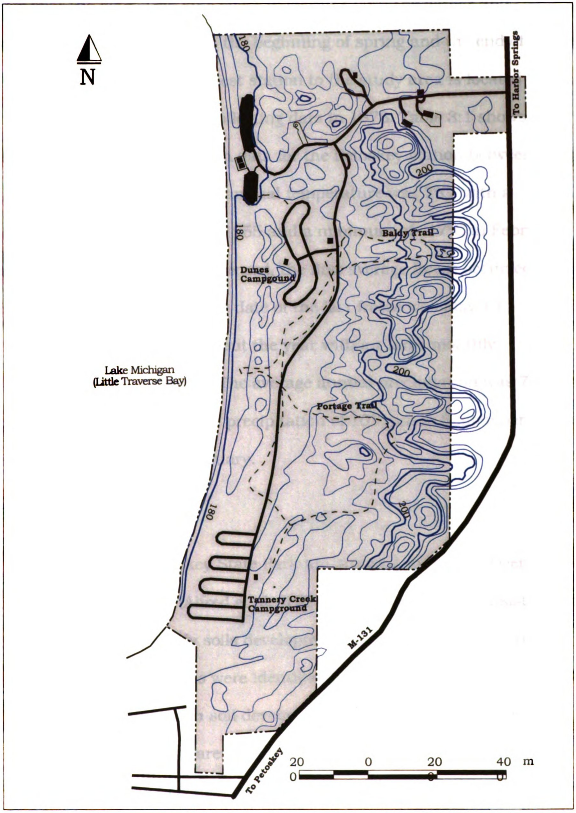

Figure 3:2. Topography of Petoskey State Park. Countours every 5 m. (From Harbor Springs Quadrangle, 1983).

flow over the lake. This lake effect moderates both the winter and summer seasons and delays the beginning of spring and the end of fall (Alfred et al., 1973). The closest station to the study area is located in the city of Petoskey and began collecting data in 1878. Table 3:1 shows temperature and precipitation data for the Petoskey station between 1952 and 1980. The average annual temperature was 7°C, with a maximum of 36°C in July of 1955 and a minimum of −17°C in February of 1979. Based on the same record, the first average freeze occurred on October 16, while the average date of the last freeze was May 10. Precipitation occurs throughout the year without large monthly variations month to month. The average annual precipitation was 770 mm, with maximum average precipitation of 101 mm in September and a minimum of 35 mm in February.

SOILS

Soils within Petoskey State Park are grouped within the Deer Park-Dune land association (Alfred et al., 1973). This association consists of deep, well drained, sandy soils developed on lake beaches, lake terraces, and dunes. Five soil series were identified by Alfred et al. (1973) within the park (Figure 3:3), with soil development increasing with distance from the shoreline. The parent materials include dune sand, sandy lake beaches, stony lake beaches, and mucky sand.

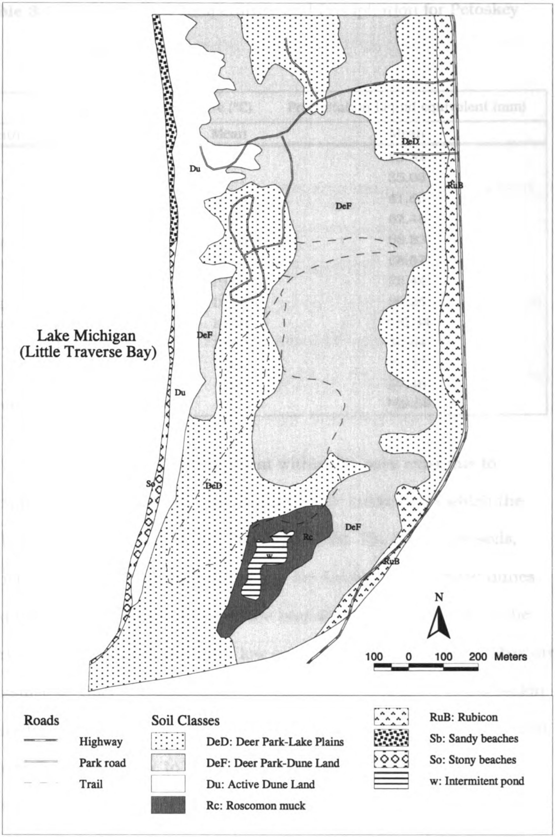

Figure 3:3. Soil Map for Petoskey State Park and sorrounding area. (Alfred et al, 1977).

Table 3:1. Mean Annual Temperature and Precipitation for Petoskey (1952-1980)

	Average Daily Temperature (°C)			Precipitation Liquid equivalent (mm)
Month	Max.	Min.	Mean	Mean
Jan.	-2.67	-9.67	-6.17	52.32
Feb.	-2.00	-11.22	-6.61	35.05
Mar.	3.33	-6.44	-1.56	41.66
Apr.	10.67	0.22	5.44	62.48
May	17.44	5.50	11.44	68.33
Jun.	22.78	11.11	16.94	68.07
Jul.	25.28	14.50	19.89	75.44
Aug.	24.83	14.22	19.50	80.26
Sep.	20.72	10.44	15.56	100.84
Oct.	14.78	5.56	10.17	56.64
Nov.	6.83	0.06	3.44	71.37
Dec.	0.28	-6.22	-3.00	57.40
Annual	11.83	2.33	7.11	769.88

Differences in soil development within the park exist due to textural differences and the age of geomorphic surfaces on which the soils are developed (Barret and Schaetzl, 1998). The Deer Park soils, which occur away from Lake Michigan, are formed on stabilized dunes and lake terraces. These soils are the best developed dune soils in the park, are well drained, and have low contents of organic matter. They are classified as Spodic Udipsamments, with A/E/Bs1/Bs2/C horizonation (Alfred et al., 1973). The dune land consists of active (unvegetated) sand dunes that occur close to the shoreline. Given the active eolian processes, soils have not developed in these areas (Alfred et al., 1973). Roscomon mucky sand occurs in the depression east of the Tannery

Creek campground. These soils develop in poorly drained, nearly level areas and are classified as mixed, frigid, Mollic Psammaquents, with A/E/Bs1/C horizonation.

VEGETATION

The vegetation history in northwestern lower Michigan has been influenced by a number of natural and human factors. Prior to settlement of the area the natural vegetation consisted of a mature conifer bog forest (Cleland, 1979). This forest, which extended east from Little Traverse Bay to Burt Lake, contained tamarack (*Larix laricina*), black spruce (*Picea mariana*), and white cedar (*Thuja occidentalis*). Other species present included red maple (*Acer rubrum*), sedge (*Carex* spp.), leatherleaf (*Chamaedaphne caliculata*), holly (*Ilex verticillata*), bog kalmia (*Kalmia polifolia*), labrador tea (*Ledum groenlandicum*), and peat moss (*sphagnum* spp.). More specifically, the vegetation in Petoskey State Park consisted mostly of pine, aspen, hemlock and fir close to Lake Michigan; cedar, tamarack, and hemlock, spur, fir, black ash, cider and white pine occurred in the eastern part of the park (Cleland, 1979).

During the late 1800s Europeans settled in northwestern lower Michigan (Alfred et al., 1973). After European settlement, lumbering, post logging fires, and changes in land use altered the pre-existing forest (Cleland, 1979). At the present time several forests associations exist throughout northwestern lower Michigan, consisting mainly of sugar maple (*Acer saccharum*), American beech (*Fagus grandifolia*), eastern hemlock (*Tsuga canadensis*), and yellow birch (*Betula lutea;* Cleland, 1979). Also common in this area are jack pine, scrub oak, and red pine, with tamarack and northern white cedar on wet sites (Alfred et al., 1973). A the present time the vegetation at Petoskey State Park consists primarily of mixed deciduous and coniferous forest. These forest ranges from predominantly deciduous (sugar maple, paper birch, American beech, and red pine) in the eastern parts of the park to predominantly coniferous in the eastern part (composed of cedar, white and red pines, and sugar maple saplings). The areas of active dunes are mostly unvegetated. However, scattered pioneer vegetation including juniper, bearberry, sandcherry, and beach grass is present on the active surfaces.

HISTORY OF LAND USE IN THE PARK

The history of Petoskey State Park presented in this section was summarized from a detailed document prepared by Sharkey (1972). The area that is currently occupied by the park was used by the Ottawa Indian tribe prior to European settlement in the area. The first recorded private owner of the area was an Ottawa named Paymegwau, who received the land ownership under a treaty in 1855. In 1860 the park land was sold to Julia Sheenan, who later sold it to the W.W. Rice company. In 1880 the Bay View, Little Traverse & Mackinaw Railroad was constructed. The railroad ran along the shoreline at the base of the dunes through what is now park property. This railroad was subsequently abandoned and its tracks removed in 1962. From 1886 until 1950 a leather tannery operated east of what is now the Tannery Creek campground. In 1934 the city of Petoskey purchased the land and named it the Petoskey Bathing Beach. Subsequently, the state of Michigan acquired the land in 1968 and established the park, which opened in July 1970. Currently the park is operated by the Parks and Recreation Division of the Michigan Department of Natural Resources (MDNR) and has 2 separate campgrounds, the Dunes and the Tannery Creek campgrounds.

Chapter 4

METHODS

Integral to this research project is an in depth understanding of both current and ancient eolian processes. For this reason a number of sites including active and stabilized dunes were studied. Additionally, the use of computer assisted cartography provided additional information about the changes that have occurred in the park in the past 35 years. The following section describes the methods used in this investigation.

MAPPING METHODS

Geomorphic mapping is a valuable necessary tool to interpret relict processes and reconstruct the geomorphic history of an area. For this reason, geomorphic mapping was an integral part of the study.

Initial mapping of geomorphic units at Petoskey State Park was obtained through stereoscopic interpretation of 1998 aerial photographs. In areas of low relief, where the height of the canopy impaired photo interpretation, the Emmet county soil survey (Alfred et al., 1973) was used as an aid to delineate individual units. This initial characterization was primarily based on the different types of dunes and size of the landforms observed as well as on the degree of soil development as mapped by Alfred et al. (1973). A total of 5 geomorphic units were identified using this approach.

The initial geomorphic map of the park was revised after visual interpretation of a 3D view of the park, generated in ArcInfo from a DEM. This DEM was constructed in ArcInfo after digitizing 2-ft contour lines from a park topographic map. The original map was constructed by MDNR after a topographic survey of the park 1964 by MDNR. Visual interpretation of the 3D perspective allowed more detailed delimitation of geomorphic units because it was constructed from a large-scale map and was not affected by canopy. The basic shape and location of the geomorphic units outlined during the initial mapping was preserved, but the delineation of their boundaries was refined based on this new information.

In addition to geomorphic mapping of the park, historical land cover changes between 1965 and 1998 were studied. The 1965 photographs provided a base line before the construction of the park, while the 1998 photographs allowed for analysis of the current land cover. This information was extracted from black and white panchromatic aerial photographs taken in 1938, 1965, and 1998. The 1938 and 1965 photographs were taken by the USDA (United State Department of Agriculture) and the 1998 photographs by MDNR (Michigan Department of Natural Resources). Vegetation cover for each year when air photographs were available was interpreted using a stereoscope and then the polygons were digitized into ArcInfo. The changes in total area of the different land cover units were visually interpreted.

FIELD METHODS

The early stages of this research began with a field reconnaissance of the park. During this reconnaissance several sites (Figure 4:1) were selected for further investigation. Sites were chosen to gather information about the timing of eolian activity at Petoskey State Park. At all the sites the visual characteristics of the surface soils and paleosols (when present) were described and samples were taken for absolute dating. Two methods of dating were used in this study to reconstruct the chronology of cycles of eolian activity and landscape stability: radiocarbon dating of organic materials in paleosols and optically stimulated luminescence (OSL) dating on dune sand. Eleven organic samples (sites 1-5, 6-toe, 7, 13, 20, 21, and 22, Figure 4:1) were sent to Beta Analytic Inc. in Miami, FL, for radiocarbon dating. The dates reported by the lab were later calibrated to calendar years BP using the method suggested by Stuiver and Reimer (1993). Additionally, 3 OSL samples (sites 11, 14, and 16; Figure 4:1) were sent to the Luminescence Laboratory, Institute of Geography and Earth Science, University of Wales, Aberystwyth, WK, for age determination.

The sampling strategy required investigation of three types of sites (Figure 4:2): (1) sites at the toe of stabilized dunes, (2) sites located on

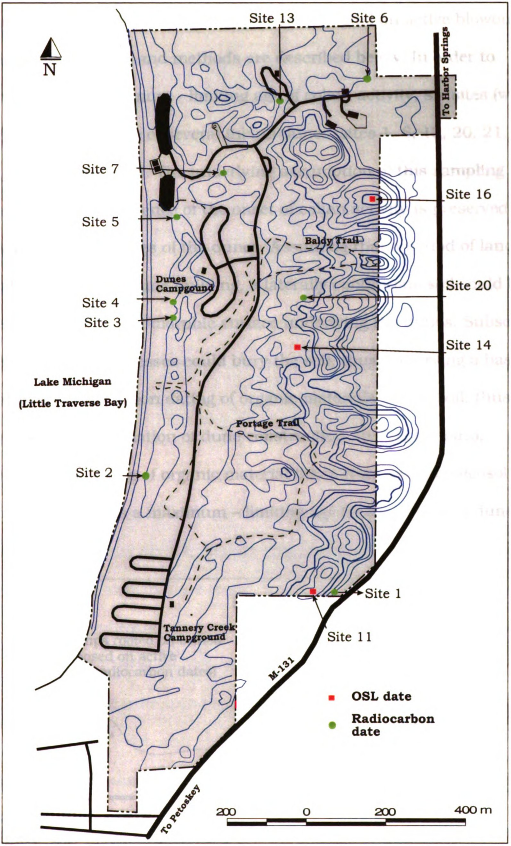

Figure 4:1. The location of study sites at Petoskey State Park.

the crest of stabilized dunes, and (3) sites located on active blowouts. The sampling rationale and methods are described below. In order to establish the maximum - limiting age of eolian activity, six sites (were examined at the toe of several inland dunes (sites 1, 6, 13, 20, 21, 22, Figures 4:1 and 4:2). The underlying assumption of this sampling strategy is that evidence of the onset of eolian activity is preserved in the basal soils at the toes of the dunes. Assuming that a period of landscape stability predated dune building, a laterally continuous soil could have developed on the geomorphic surface underlying the dunes. Subsequent dune building processes could bury the soil, thus preserving a basal paleosol. Radiocarbon dating of organic materials in this soil, thus, represents the initiation of dune construction. In this scenario, radiocarbon dating of organic material taken from such a paleosol (basal paleosol) provides a maximum – limiting age for the overlying dune.

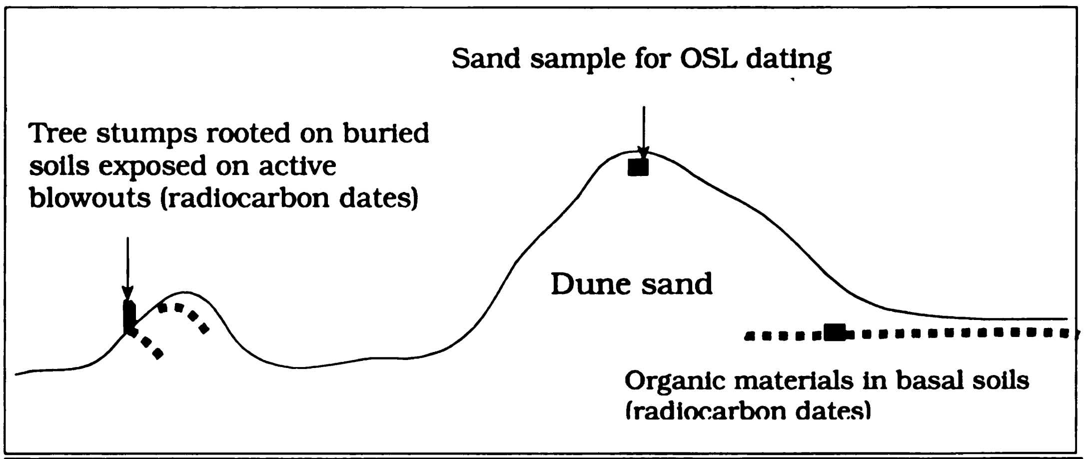

Figure 4:2. Schematic cross-section showing the types of radiometric dates used in this study.

At sites 1 and 6 (Figure 4:1) pits were hand excavated to a depth of approximately 1.5 m and samples for radiocarbon dating were taken from the A horizons of buried soils underlying dune sand. At sites 13 and 20 (Figure 4:1) the stratigraphy at the toe of the dunes was examined using a bucket auger. In order to determine the lateral continuity of the buried soils found at these sites several auger holes were excavated at approximately 2-3 m intervals. When the observed stratigraphy and soil development were laterally traceable, samples for radiocarbon dating were taken. The samples were then carefully wrapped in aluminum foil to prevent contamination from modern organic materials.

In addition to the radiocarbon dating of basal paleosols, samples for OSL dating were taken at the crests of three stabilized dunes (sites 11, 14, and 16; Figures 4:1 and 4:2). Dating dune crests to determine the minimum - limiting age of dune construction assumes that the crest is the last part of the dune to stabilize (McKee, 1979). OSL dating is used to resolve the time when sand grains were last exposed to sunlight, thus providing an estimate of the time elapsed since burial (Murray and Wintle, 2000). Therefore, OSL dating of dune crest sediments was used to establish an upper limit to eolian activity on the stabilized dunes.

At sites 11, 14, and 16 (Figure 4:1) pits were hand excavated from the crest of the dune to the depth of the C horizon (generally located at about 1-1.5 m depth) and samples were taken from the C horizon for OSL dating. These samples were light sensitive and therefore were taken in the dark to prevent exposure to sunlight. The samples were collected by inserting an open ended PVC pipe into the C horizon and subsequently sealing the pipe with duct tape. Subsequently, the OSL was measured from 180-212 μm quartz grains using the single aliquot regenerative dose method described by Murray and Wintle (2000).

Five sites (sites 2-5, and 7; Figures 4:1 and 4:2) were studied in the active blowouts to establish the chronology of recent eolian activity on the modern dunes. In these dunes the blowouts were carefully examined for buried soils within the dune sand sequence (Figure 4:2), which represent periods of past landscape stability since dune building begun. Even though buried soils were observed in most of the blowouts, only sites that displayed laterally continuous paleosols and upright trees were selected for dating. This prevented sampling trees from slump blocks, which would have provided artificially young dates. Samples of approximately 300 grams were taken from the upright trees for radiocarbon dating. The samples we carefully wrapped in aluminum foil and then placed in plastic bags to prevent contamination.

Chapter 5

RESULTS AND DISCUSSION

As indicated in the previous chapter, the purpose of this research is to determine the geomorphic history of sand dunes at Petoskey State Park. In that context, this chapter presents the results of the geomorphic investigations with emphasis on the geomorphic relations and geochronology.

LANDFORM ASSEMBLAGES

Mapping of landform assemblages at Petoskey State Park provided a generalization of the landscape that facilitated interpretation of the existing landforms. The geomorphic map of Petoskey State Park (Figure 5:1) is based on aerial photography interpretation and field observations. Five landform assemblages (or landform assemblages) were identified within the park (Figures 5:1 and 5:2): (1) Lake Terrace, (2) Parabolic Dunes, (3) Onlap dunes (4) Linear and Incipient parabolic dunes, (5) Active Dunes. In general, the largest dunes occur in the east part of the park and the size of the dunes decreases towards the shoreline. The following sections describe the geomorphic characteristics of each of these landform assemblages.

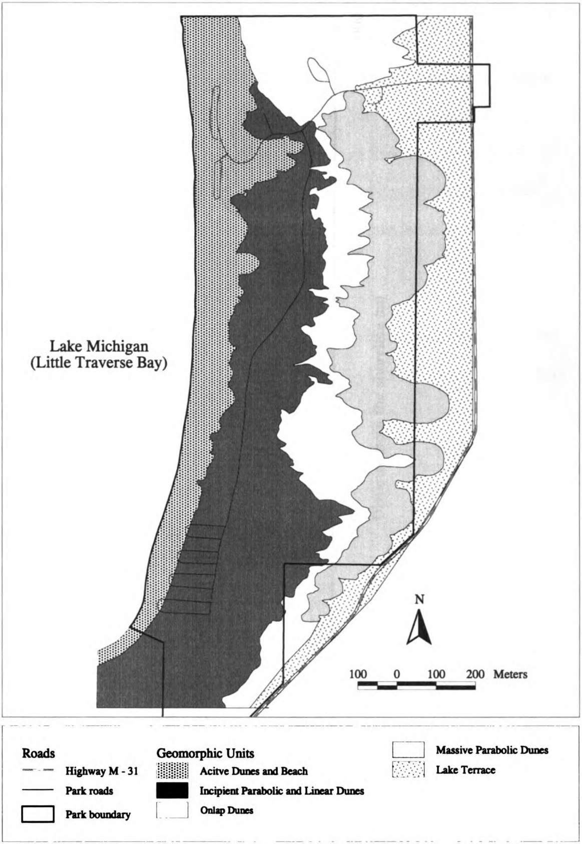

Figure 5:1. Landform Assemblages at Petoskey State Park.

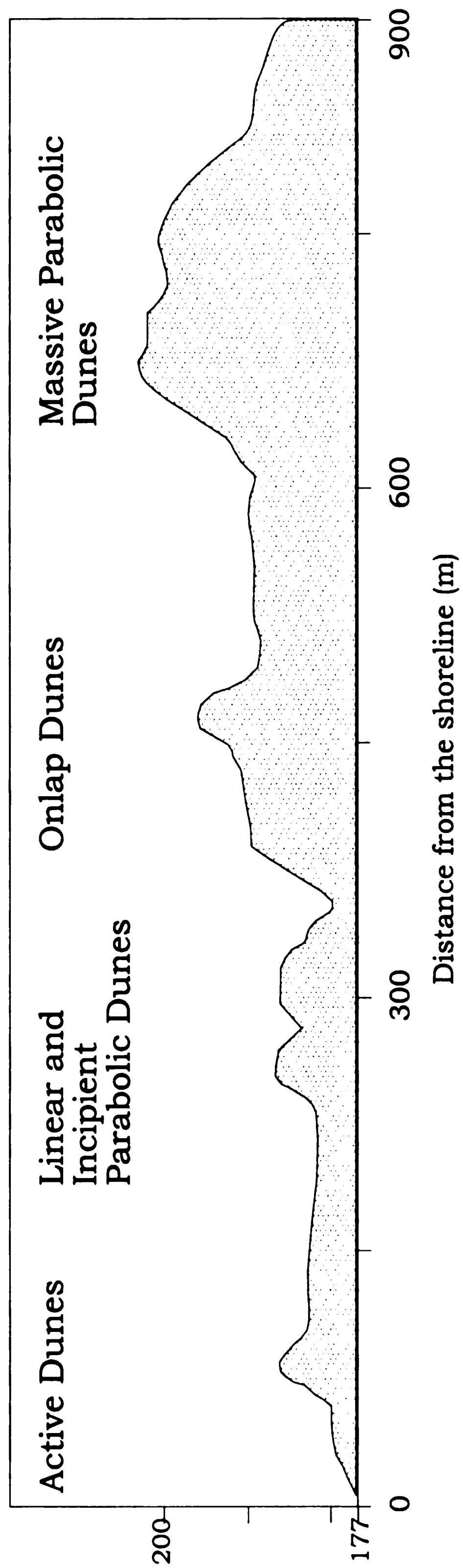

Figure 5:2. Generalized E-W cross-section of Petoskey State Park.

Lake Terrace

The lake terrace is located in the eastern part of the park (Figure 5:1). This is a low sloping surface and occurs at an elevation of approximately 186-189 m, which suggests that this terrace formed during the Nipissing I transgression of Lake Michigan (Hough, 1958). This landform is composed of beach and nearshore sand and gravel.

Massive parabolic dunes

This landform assemblage occurs along the eastern margin of the park, where it borders the lake terrace landform assemblage (Figure 5:1). At the present time the parabolic dunes assemblage is heavily forested with a mixed deciduous-coniferous forest, and, hence, stable. These dunes have their arms oriented with an azimuth of 270° (Figure 5:3), which indicates that the predominant winds during their formation were westerly. These dunes vary in height from 10 to 45 m (Table 5:1; Figure 5:3). According to Pye's (1990) classification, the dunes are lobate due to their high length to width ratio. Only one dune (labeled I on figure 5:3) was classified as hemicyclic (almost round); this dune is also one of the smallest in this assemblage (Figure 5:3). The largest dunes, which also display the highest length to width ratio, occur towards the central part of the assemblage.

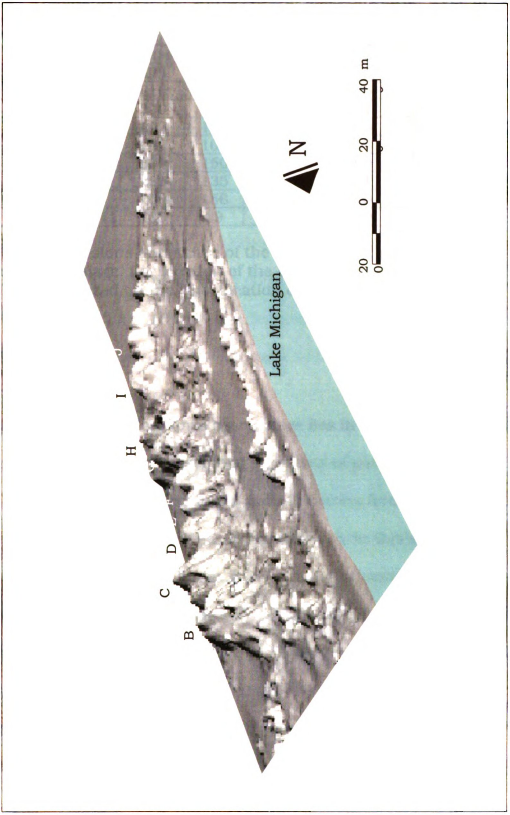

Figure 5:3. Digital elevation model of Petoskey State Park. Vertical exa-geration = 3x

Dune Id (see figure 5:3)	Sampling site	Length (m)	Width (m)	Height (m)	Length:width ratio	Type of dune
A		90	75	10.7	1.2	Lobate
B	Site 16	225	147	39.6	1.5	Lobate
C		210	81	30.5	2.6	Lobate
D		210	99	45.1	2.1	Lobate
E		192	126	34.1	1.5	Lobate
F		144	108	28.0	1.3	Lobate
G		300	150	24.4	2.0	Lobate
H		285	135	18.9	2.1	Lobate
I		150	156	28.7	1.0	Hemicyclic
J	Sites 1&11	135	69	32.9	2.0	Lobate

Table 5:1. Dimensions and shape of the dunes in the parabolic dunes landform assemblage. For location of the dunes refer to Figure 5:3. The type of dune is based on the classification of Pye (1990).

Onlap dunes

The onlap dunes landform assemblage lies in the center of the park (Figure 5:1). This landform assemblage consists of parabolic dunes, which are forested (mixed deciduous and coniferous forest) and hence stable at the present time. The easternmost dunes in this assemblage onlap the massive parabolic dunes assemblage. The onlap dunes are much smaller in size than the ones in the parabolic dunes assemblage, ranging in height between 10 and 15 m. These dunes are predominantly hemicyclic according to Pye's (1990) classification scheme, with only a few dunes being lobate. This plan view form probably suggests that the onlap dunes formed during a shorter period of time that the massive parabolic dunes assemblage (Pye, 1990).

Linear and Incipient Parabolic Dunes

This landform assemblage occurs west of the onlap dunes landform assemblage (Figure 5:1). It consists of incipient parabolic dunes, three discontinuous linear dunes, and a paleo - lake terrace. The incipient parabolic dunes reach heights of 2-3 m and are classified as hemicyclic on Pye's (1990) classification. Three linear dunes (or paleo-foredunes) were identified during fieldwork, however, due to their low relief (<2 m) only one of them can be observed on the topographic map and DEM (Figure 5:3). The paleo-lake terrace is a wide low relief surface that occurs in the western part of this assemblage at an elevation of approximately 180 – 183 m.

Due to this assemblage's low relief (Figures 5:2 and 5:3), both campgrounds are located in this assemblage, and this landform assemblage is thus heavily impacted by park visitors. Additionally, the southern part of the assemblage (east of the Tannery Creek Campground) was used in the early 1900s for a leather tannery (Sharkey, 1972) and was also greatly modified. Current evidence for this impact is the lack of vegetation, which has been unable to re-colonize due to the severe contamination of the soil. Minor areas of slumping in the dunes were observed around this area as well.

Active Dunes

The active dunes landform assemblage occurs in the westernmost part of the park, adjacent to the lake (Figure 5:1). This assemblage is of particular interest to this study from management perspective because the active eolian processes are an important issue. This landform assemblage includes a series of blowout dunes and a linear foredune, both of which parallel the current shoreline of Lake Michigan (Figures 5:1, 5:3, and 5:4). The foredune and blowout dunes occur only to the north of the Tannery Creek campground beach access. North of this location the beach is sandy, providing a sufficient sand supply for dune construction. Probably, as a result of this sand supply, the active dunes are considerably larger farther north along the assemblage. South of the Tannery Creek Campground beach access, a small escarpment that varies in height between 1 and 2 m. replaces the dunes. Along this transect, the beach is composed of gravel. The absence of dunes in this area of the park is probably related to the lack of abundant sand in the beach.

The blowout dunes are the primary site of current eolian activity in the park and are typical landforms of vegetated coastal areas. The dunes are parallel to the shoreline and range in elevation between 8 and 15 m. The windward (western) slope of the dunes faces Lake Michigan and contains numerous active blowouts. Due to their shape (deep, relatively

a

b

Figure 5:4. Photographs of the active dunes. (a) View of blowout dunes, notice the weakly developed paleosol and a person standing for scale. The warning sign on center of picture advises visitors not to climb the dunes. (b) View the foredune (center) and blowout dunes (left). Note the dense grass cover.

narrow, with steep walls), the blowouts can be classified as trough blowouts in Carter's (1990) classification. Deposition of eolian sand in these dunes occurs on their eastern slope and encroaches the Dunes Campground (Figures 5:1 and 5:5), posing a problem for park management.

The east slope of the blowout dunes is predominantly vegetated by coniferous forest, which includes cedar, white and red pines, as well as some scattered maple trees and deciduous shrubs (e.g. poison ivy). This vegetative composition, coupled with the weak development of the surface soil (A/C profile), suggests that the blowout dunes have been stable for a short period of time, probably less than 200 years (Olson, 1958a; Litcher, 1997).

The active sections of the blowout dunes have scattered pioneer vegetation, including bearberry, sand cherry, baby's breath, and dune grass, among others. The presence of sand cherry and bearberry on the blowouts indicates that the rates of sedimentation are on the order of a few centimeters per year (Olson, 1958), which suggests that the blowout dunes are sediment - starved. This creates a negative sediment budget in the blowout dunes; that is, sedimentation on the lee side of the blowout dunes at the present time occurs at the expense of their windward side. Evidently, this process would result in an increase in the extent (size and depth) of the blowouts. However, sedimentation on the lee side, as evident from the current processes of soil formation, is not sufficient to completely bury the modern soil. Hence, blowout formation and evolution cannot provide the large volumes of sand required to bury surface soils and vertically aggrade coastal dunes.

The foredune is an elongated ridge, 2 m high, that parallels the shoreline (Figure 5:4). The vegetation on the foredune consists primarily of pioneer vegetation, including beach grass, and baby's breath. Park visitors heavily utilize this ridge, which enhances the erosive processes that are naturally active.

GEOMORPHIC HISTORY OF LANDFORM ASSEMBLAGES

As mentioned previously, the geomorphic investigations conducted in the park focused on identifying periods of eolian activity and landscape stability. This section presents data from these investigations and geomorphic interpretations for each landform assemblage. A summary of the radiometric dates is presented in table 5:2.

Geomorphic History of the Massive Parabolic Dunes

In order to establish a chronological framework for the formation of this landform assemblage, three sites were examined. One of these sites was located at the toe of a dune, while the remaining two sites were located at dune crests. The information provided by these investigations allowed reconstructing the history of the massive parabolic dunes landform assemblage.

Site	Type of date	Age	Landform Assemblage
1	Radiocarbon from charcoal	4815-4425 cal yr. B.P.	Massive Parabolic Dunes
2	Radiocarbon from wood	300-0 cal yr. B.P.	Active Dunes
3	Radiocarbon from wood	270-5 cal yr. B.P.	Active Dunes
4	Radiocarbon from wood	300-0 cal yr. B.P.	Active Dunes
5	Radiocarbon from wood	470-290 cal yr. B.P.	Active Dunes
6	Radiocarbon from charcoal	1855-1625 cal yr. B.P.	Onlap Dunes
7	Radiocarbon from wood	515-315 cal yr. B.P.	Active Dunes
11	OSL	2300 - 2060yr. B.P.	Massive Parabolic Dunes
13	Radiocarbon from charcoal	2860-2740 cal yr. B.P.	Onlap Dunes
14	OSL	1820-1720 yr. B.P.	Onlap Dunes
16	OSL	1080-980 yr. B.P.	Massive Parabolic Dunes
20	Radiocarbon from charcoal	Modern	Onlap Dunes

Table 5:2. Summary of radiometric dates used in this study.

In order to establish the maximum limiting age of dune formation in this assemblage, the toe of one dune (site 1) was examined (Figures 5:5 and 5:6). Special attention was given to identifying the stratigraphic units and paleosols present in the pit because they provide evidence of periods of active sedimentation and landscape stability, respectively. Field investigations revealed that the dunes overlie lacustrine sediments that are topographically consistent with the lake terrace landform assemblage (Figure 5:6).

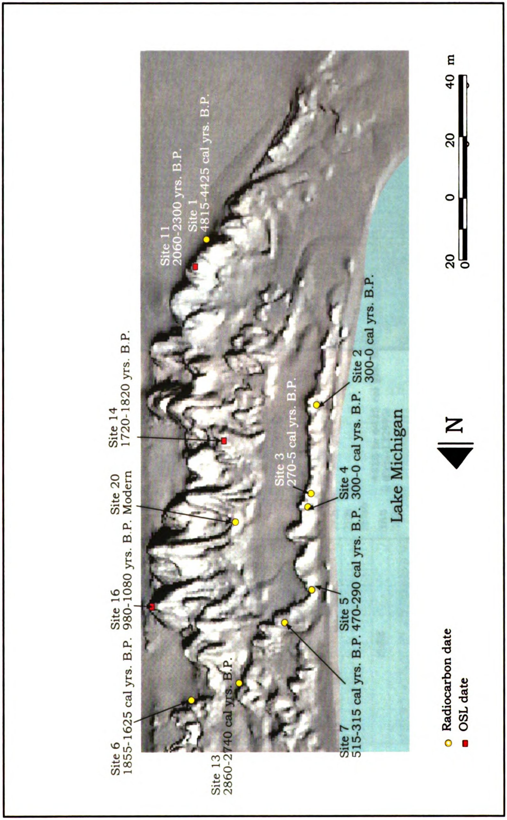

Figure 5:5. Oblique view (to east) showing location of sampling sites and radiometric dates.

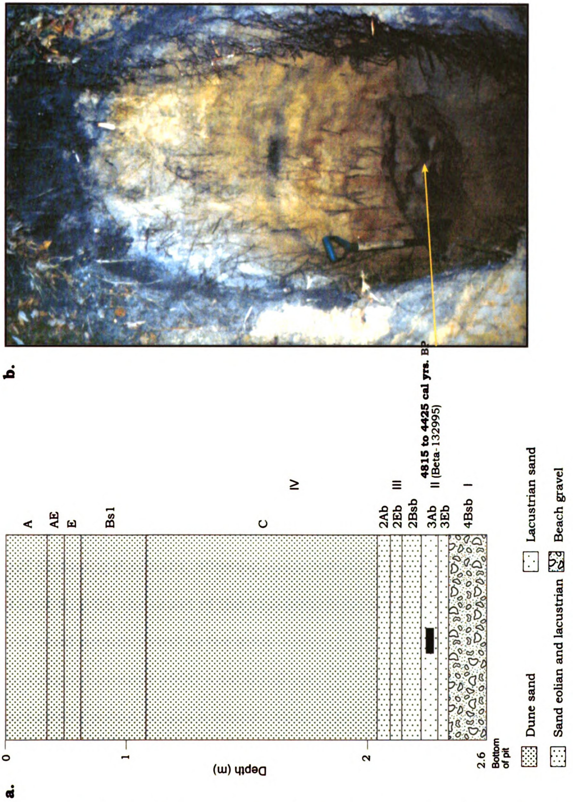

Figure 5:6. Investigations at site 1: a) Stratigraphy and soil development; b) Photography of the pit.

Site 1 was excavated at the base of one of the southernmost parabolic dunes and its stratigraphy is shown in Figure 5:6. Two clearly distinguishable paleosols and four stratigraphic units (I-IV) were observed at this site. The basal unit (Unit I) is a sandy gravel. A unit of massive sand (Unit II) approximately 20 cm thick with a moderately developed soil overlies this basal gravel. Overlying this unit is a 25 cm thick sandy unit (Unit III) with a moderately developed soil. At the top of the sequence is a massive sandy unit (Unit IV) with scattered organic material. A sample for radiocarbon dating (charcoal) was taken from the top of unit II that dated at 4620 ± 190 cal yrs. B.P. (Figure 5:6).

Based on these observations the sequence of events at site 1 was probably as follows: The gravel at the base of the excavation (Unit I) and the sand of unit II likely represent lacustrine shore-zone deposits. These deposits were presumably deposited during the early Nipissing transgression and exposed subaerially sometime between the Nipissing I and Nipissing II phases, allowing some soil formation to occur at the top of unit II. With the onset of the Nipissing II transgression, lake levels rose and sand accumulated in the area, burying the soil as deposition of unit III occurred around 4800 – 4400 cal yrs. B.P. Whether unit III is exclusively eolian or has a mixed eolian – lacustrine origin is not clear based on the available evidence. However, the complete preservation of the soil on unit III suggests that an eolian origin is more likely. After lake level dropped from the Nipissing level a soil developed on top of unit III, which was later buried by eolian sand (Unit IV).

In order to determine when eolian activity ceased in the parabolic dunes assemblage two OLS samples were collected, one each at sites 11 and 16 (Figure 5:5). These samples were taken at the crest of the dunes, based on the assumption that crests are the last portions of the dunes to stabilize (McKee, 1979). Site 11 is located on the crest of the dune which also contains site 1. This OSL sample yielded an age of 2180 ± 120 years. Site 16 is located in the north part of the assemblage and provided a date of 1030 ± 50 years.

The stratigraphic relationships and radiometric (both radiocarbon and OSL) dates clearly indicate that the massive parabolic dunes at Petoskey State Park are post-Nippissing in age. These results also suggest that eolian activity ceased in the southern part of the park approximately 1000 years earlier than in the northern part. This dichotomy may explain the more elongated shape and larger size of the dunes in the northern part of the assemblage. The reasons for such differences are impossible to explain with the available evidence but they could be related to colonization of the southern dunes by pioneer vegetation (Olson, 1958a), the formation of a protective foredune in this area (Litcher, 1995), or to the apparent increase in sand availability towards the north (as suggested by the blowout dunes).

Geomorphic History of the Onlap dunes

Four sites were investigated in the onlap dunes landform assemblage (Figure 5:5). Site 14 was located at the crest a dune, and sites 6, 13, and 20 at the toe of dunes. The stratigraphy at sites 6 and 13 is very similar (Figure 5:7), consisting of basal sandy gravel, which occurs at an elevation of approximately 186 m., overlain by medium, well-sorted sand with a moderately developed soil, capped by eolian sand. Organic samples (charcoal) were taken from both sites for radiocarbon dating (Figures 5:5 and 5:7) and yielded ages of 1740 ± 115 cal yrs. B.P. for site 6, and 2800 ± 60 cal yrs. B.P. for site 13. Based on the notion that Lake Michigan dune fields developed during a long term decrease in lake level one would expect the age of the dunes to increase landward. However, this expected trend is contradicted by the younger radiocarbon date of at site 6. Nonetheless, both of these radiocarbon samples were obtained from the A horizon of well developed soils that contained large pieces of charcoal. Hence, these dates are considered representative of the evolution of the onlap dunes landform assemblage.

The apparent disagreement between the younger date at site 6 and the older date at site 13 can be explained in terms of dune dynamics. Figure 5:5 shows that the dunes in this assemblage are randomly distributed and sometimes disconnected from each other. This distribution suggests that these dunes were transgressive (moving inland; Carter, 1990) at some time during their formation. If this

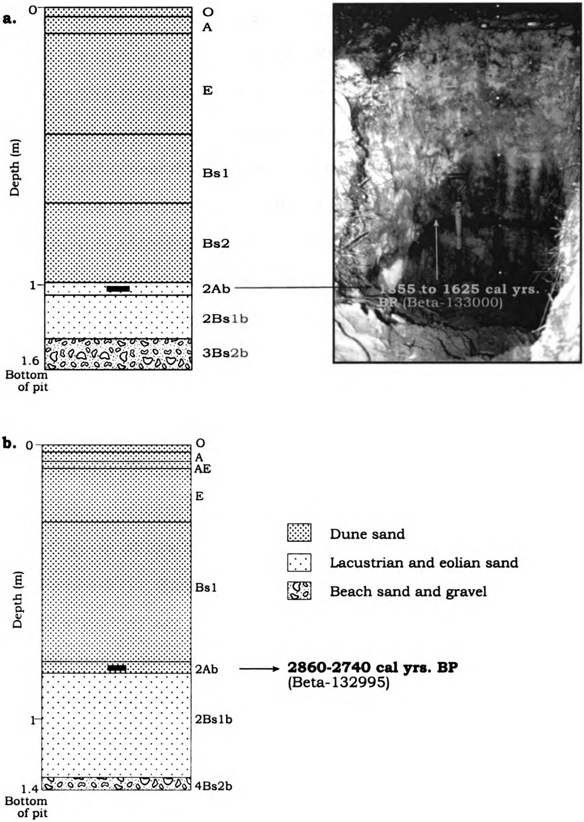

Figure 5:7. Investigations at the toe of onlap dunes: a) Site 6; b) Site 13.

hypothesis is correct, the formation of these dunes could have initiated around site 13, burying the soil about 2800-2700 cal yrs. B.P. At some time after their initial formation, the dunes might have developed blowouts, which eventually contributed to the detachment of some dunes from the initial linear dune. These dunes hypothetically would have moved inland and reached the position of site 6 sometime around 1800 – 1600 yrs. B.P. Additionally, close examination of Figure 5:5 demonstrates the absence of dunes in this assemblage north of site 13. The absence of dunes in this portion of the landform assemblage could be due to inland movement of the dunes that initially occupied that position.

Site 20 was adjacent (approximately 5 m south; Figure 5:5) to a cleared patch of forest which appeared anthropogenic in nature. This area displayed possible evidence of a controlled fire. Three borings were conducted at the toe of the dune and a buried soil was found at a depth of 40-45 cm at all locations. A sample of charcoal was taken for radiocarbon dating from the A horizon of this buried soil. This sample yielded a "modern" age. Based on the degree of soil development overlying this buried soil (A/E/Bs horizonation), it is not likely that the date represents the time of burial of the soil. Additionally, Beta Analytic reported the presence of "bomb carbon" in the sample, which means that carbon absorbed within the past 40 years was present. Furthermore, the proximity of the burned clearing of the forest suggests that the material dated probably was part of the root system of a burned tree. Therefore, this date is regarded as contaminated and is not useful to this study.

A sample for OSL dating was collected at the crest of a dune in site 14 (Figure 5:5) in order to establish the minimum limiting age for the formation of this map. This sample yielded an age of 1770 ± 50 yrs. B.P., indicating that stabilization at this site occurred around this time.

Thus, field observations and radiometric dating indicate that the sequence of events in the onlap dunes landform assemblage was probably as follows: (1) The beach gravel at the base of the sequence probably represents the shoreline zone of the Nipissing II transgression; (2) as lake levels dropped, beach sand and probably an eolian cap were emplaced upon the gravel, depositing unit II; (3) a period of landscape stability occurred and a soil developed on unit II, probably during the early Algoma transgression; (4) at the end of the Algoma transgression eolian activity was reactivated, burying the soil; (5) this period of eolian activity ceased about 1750 yrs. B.P. as indicated by an OSL date at site 14.

Geomorphic History of the Linear and Incipient Parabolic Dunes

Investigations on linear and incipient parabolic dunes concentrated on the swales between the ridges, where organic materials could hypothetically accumulate during periods of high lake levels (Thompson, 1992). In an effort to locate datable materials, several auger sites where investigated east of the Tannery Creek. No buried soils were found during these investigations. Probing revealed that the dunes are underlain by beach gravel that varies in elevation from 180 to 183 m, decreasing towards the present shoreline. According to Thompson and Baedke's (1997) lake level curve (Figure 2:3), several significant fluctuations occurred between 4000 and 1300 yrs. B.P. The first and most important of these fluctuations was the Algoma transgression, which peaked about 3500 yrs. B.P and occurs at an elevation around 183 in the study area (Leveret and Taylor, 1915).

The lakeward position of the paleo foredunes relative to the Algoma terrace at Petoskey State Park suggests that eolian activity at this landform assemblage initiated sometime after lake levels dropped from the Algoma level. A more precise determination of the age of this landform assemblage is not possible based on the evidence available for this study. However, an approximate age can be determined by comparison to Thompson and Baedke's (1997) lake level curve for Sturgeon Bay (approximately 30 km north of the park, Figure 2:3). This curve suggests that ridges at elevations between 183 – 180 m formed at Sturgeon Bay between 2300 and 1300 yrs. B.P. Due to the strong differential isostatic uplift that northern lower Michigan has undergone, these dates cannot be extrapolated directly to Petoskey State Park. However, they provide a minimum limiting age, indicating that the paleo foredunes formed at least 1300 years ago at Petoskey State Park.

The presence of scattered incipient parabolic and linear dunes across the landform assemblage suggests that the focus of eolian activity probably shifted lakeward as lake levels fluctuated. The formation of the 3 paleo foredunes probably occurred in a fashion consistent with the currently accepted models of foredune formation (Olson, 1958c, Thompson and Beadke, 1997). Minor fluctuations in lake level allowed for the formation a beach ridge during a highstand, as lake level dropped, a wide beach was exposed to eolian processes, creating the dune cap on the ridge. This process occurred at least 3 times during the formation of this assemblage, as evidenced by the 3 paleo foredunes preserved in the southern part of the assemblage.

A prominent feature associated with lake level changes during the formation of this landform assemblage is the lake terrace where the Dunes Campground lies. This terrace is a wide, low lying area at an elevation of 180 m; it is approximately 200 m wide by 800 m long and likely represents a pronounced regression of Lake Michigan in the post – Algoma period. According to Thompson and Baedke's (1997) lake level curve for Sturgeon Bay, MI (Figure 2:3), the last time lake levels were at an elevation of 180 m in the region was about 1300 yrs. B.P. Thompson and Baedke's (1997) studies also indicate that a significant fall in lake level occurred around this time, which probably explains the occurrence of the prominent lake terrace in the northwestern part of this assemblage at Petoskey State Park.

Geomorphic History of the Active Dunes

Investigations in the active dunes concentrated along the blowouts. The blowout dunes overlie beach gravel at an elevation of 180 m. Thompson and Beadke's (1997) lake level curve for Sturgeon Bay, MI (Figure 2:3) suggests that lake levels were at this elevation around 1000 years ago. Therefore, the blowout dunes probably initiated as a foredune cap on a beach ridge some time after lake levels dropped from this 180 m level. The absence of buried soils at the base of the blowout dunes suggests that dune building initiated relatively soon after or probably as lake levels dropped.

The blowout dunes were carefully examined during fieldwork for evidence of both periods of stability as well as eolian aggradation. A total of 5 sites were studied on these dunes (sites 2-5 and 7; Figure 5:5), buried soils were observed at each of these locations (Figures 5:4 and 5:8). Samples for radiocarbon dating from dead pine or cedar trees in their growth position were collected at each site.

The investigations at sites 2-5 and 7 revealed that at least 2 buried soils occur in the blowout dunes (Figure 5:8), both of which contain stumps of pine and cedar trees. According to Olson (1958) pine trees may colonize dune landscapes as soon as 60 years after stabilization, which suggests that the surfaces where they grew (now paleosols) had been stable for at least 60 years. In addition, the sampled trees appeared to be 100-150 years old at the time of death (W. Loope, personal

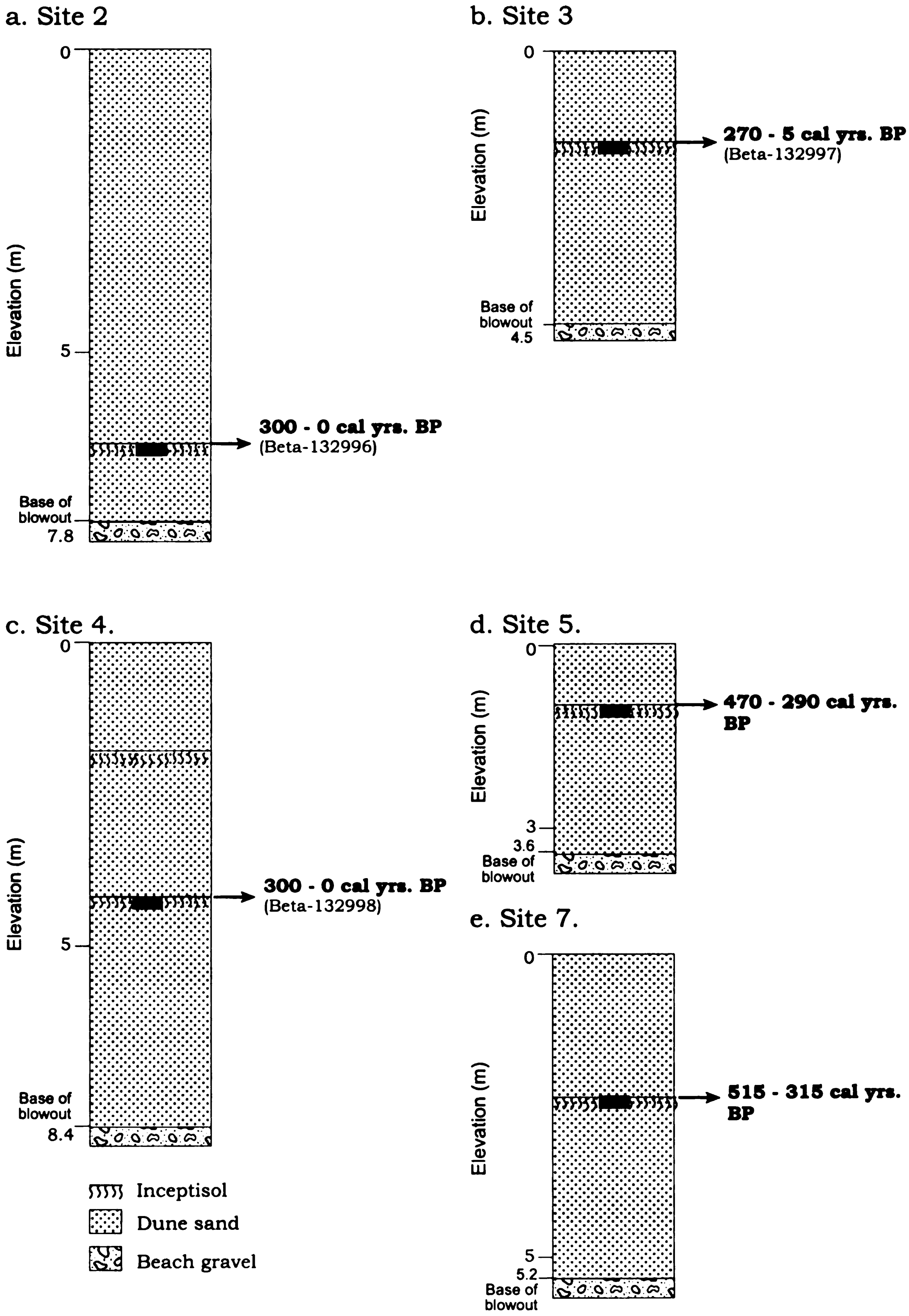

Figure 5:8. Investigations in the blowout dunes. See Figure 5:4 for geographic location of each site.

communication), which indicates that the period of stability was probably at least 100 years. Soil development in the buried soils is very weak (A/C profile), which supports the hypothesis that these periods of stability were relatively short lived (probably around 100 years).

The stratigraphic units observed at the blowout dunes (sites 2-5 and 7) suggest that several cycles of eolian activity/stability contributed to their formation. Sand accumulation initiated sometime after a lake level drop, around 1000 years ago. At the present time between 1-2 m of this unit are preserved in the northern part of the blowout dunes (sites 5 and 7; Figure 5:8). This initial accumulation was followed by a period of stability, as evidenced by a buried soil at sites 5 and 7. This period was probably short lived based on the degree of development of the paleosol (A/C profile) and the type of trees observed on this surface (white pine). Renewed eolian activity around 500 - 300 cal. yrs. B.P. buried this soil. Following deposition of this second unit, a new period of stability occurred. Renewed eolian activity buried the soil that developed during this period of stability. Radiocarbon dates from sites 2-4 indicate that burial occurred sometime after 300 yrs. B.P. A second paleosol at site 4 buried by eolian sand indicates that at least one more cycle of eolian activity occurred at this site.

In summary, at least 3 periods of eolian activity are evident in the blowout dunes from field and radiocarbon evidence. The data clearly shows a cyclicity of eolian activity and landscape stability. The fact that

3-4 periods of eolian activity occurred in the blowout dunes in the past 400 years separated by paleosols suggests that the periods of eolian activity were probably short lived (only a few decades). This would require large volumes of sand moving through the system in short periods of time.

Recent Changes in Land Cover

One of the goals of this research was to provide DNR with information about recent processes that could be used for park management. In order to meet this goal, changes in land cover between 1965 and 1998 were interpreted from aerial photographs (Figures 5:9 and 5:10). This information is particularly significant because vegetation cover plays a major role in the stability of sand dunes (Olson 1958a). Five land cover categories were defined and mapped from aerial photography. Dense forest are areas that appear heavily forested on the photos, while the areas mapped as forest savanna correspond to areas were the surface was visible through the canopy. The transition zone refers to areas with scattered vegetation (grass and shrubs). Areas mapped as unvegetated sand were areas were no vegetation was present, and, finally, urban areas are areas outside the park, where the predominant land use is urban. Some major trends were observed from the interpretations of aerial photographs:

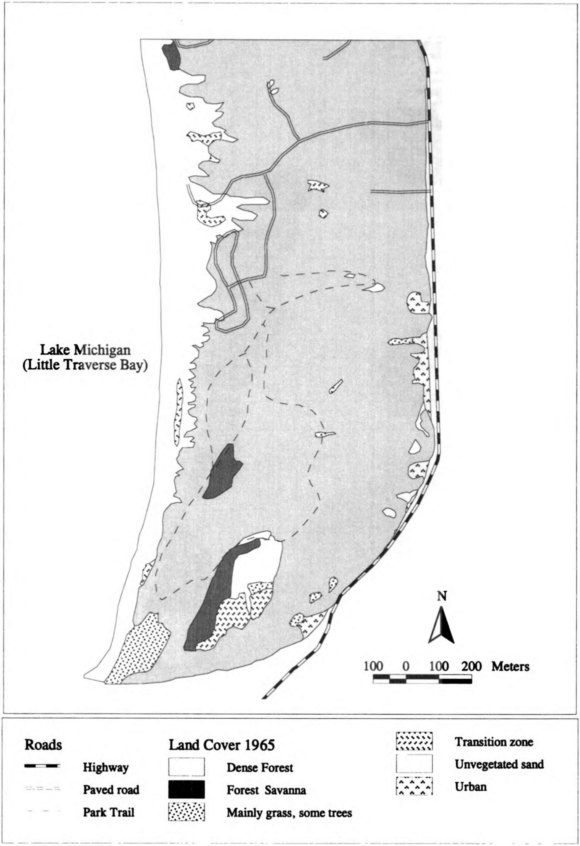

Figure 5:9. Land Cover Map 1965.

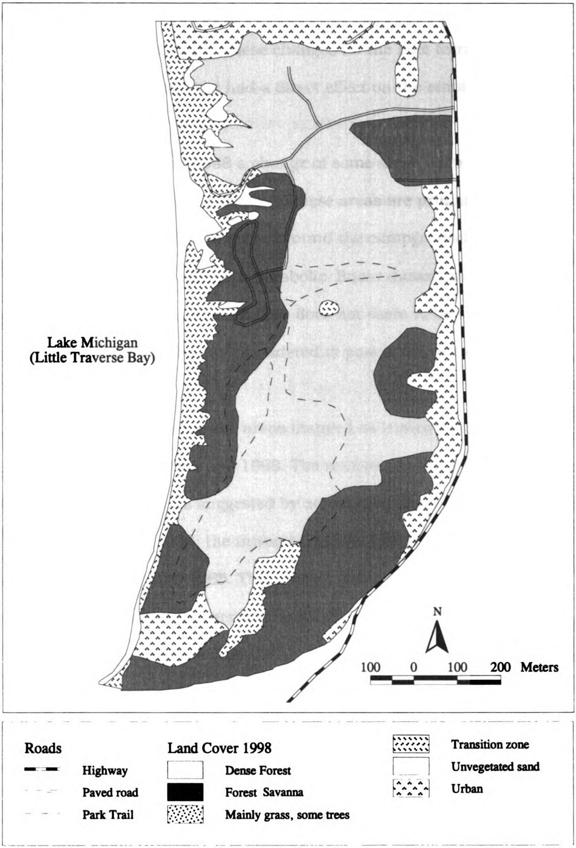

Figure 5:10. Land Cover Map 1998.

1. Intense urbanization of the areas surrounding the park occurred between 1965 and 1998. These changes on the lake terrace landform assemblage and have not had a direct effect on the stability of the sand dunes in the park.
2. Between 1965 and 1998 a change of some areas from dense forest cover to forest savanna is evident. These areas are primarily related to heavy use in the park and occur around the campgrounds, roads, and trails, and affect the incipient parabolic, linear dunes, and the lee side of the blowout dunes. This process does not seem to be significant at the present time and is not considered to pose a threat to the stability of the dunes.
3. A marked increase in the areas mapped as transition zone occurred between 1965 and 1998. The maximum extent of unvegetated sands, as suggested by aerial photographs, occurred around 1965. By 1998, the unvegetated areas were significantly reduced compared to 1965. This change is likely related to the strong efforts by the park management to stabilize the foredune and blowout dunes by planting dune grass.

Interpretation of aerial photographs indicated that changes in forest density are concentrated along the areas of heavy use (i.e. around the campgrounds and trails). This is partially a result of the construction of the park infrastructure, but undeniably, the constant traffic of pedestrians along the trails has contributed to erosional problems in the linear and incipient parabolic dunes landform assemblage. Additionally, users of the Dunes Campground use the blowout dunes as a pathway to and from the beach despite the efforts by park management to direct the traffic of pedestrians elsewhere. This impairs the efforts of park management to reduce the volumes of sand mobilized across the active dunes. Furthermore, heavy use of the foredune for recreational purposes may reduce the vegetation cover over the summer and leave the foredune unprotected against the fall storms, enhancing erosion. Sediments eroded from the foredune would eventually reach the blowout dunes and contribute to the active sedimentation over the Dunes Campground.

Although human impact does not appear to be the primary driving force of eolian processes affecting the active dunes, it appears to enhance the natural erosional processes. As mentioned previously, human activities may potentially contribute to loss of vegetative cover on the dunes, hence enhancing and accelerating the naturally active eolian processes. The processes of deposition and erosion on these dunes are highly sensitive to changes on the fragile vegetation cover.

The geomorphic history and recent vegetation changes investigated in this study represent a valuable management tool for coastal management. Figure 5:11 shows an E-W topographic cross-section of the park, including the dunes campground. Visual interpretation of this profile suggests that large volumes of sand can be mobilized across this landscape over relatively short periods of time (100s of years).

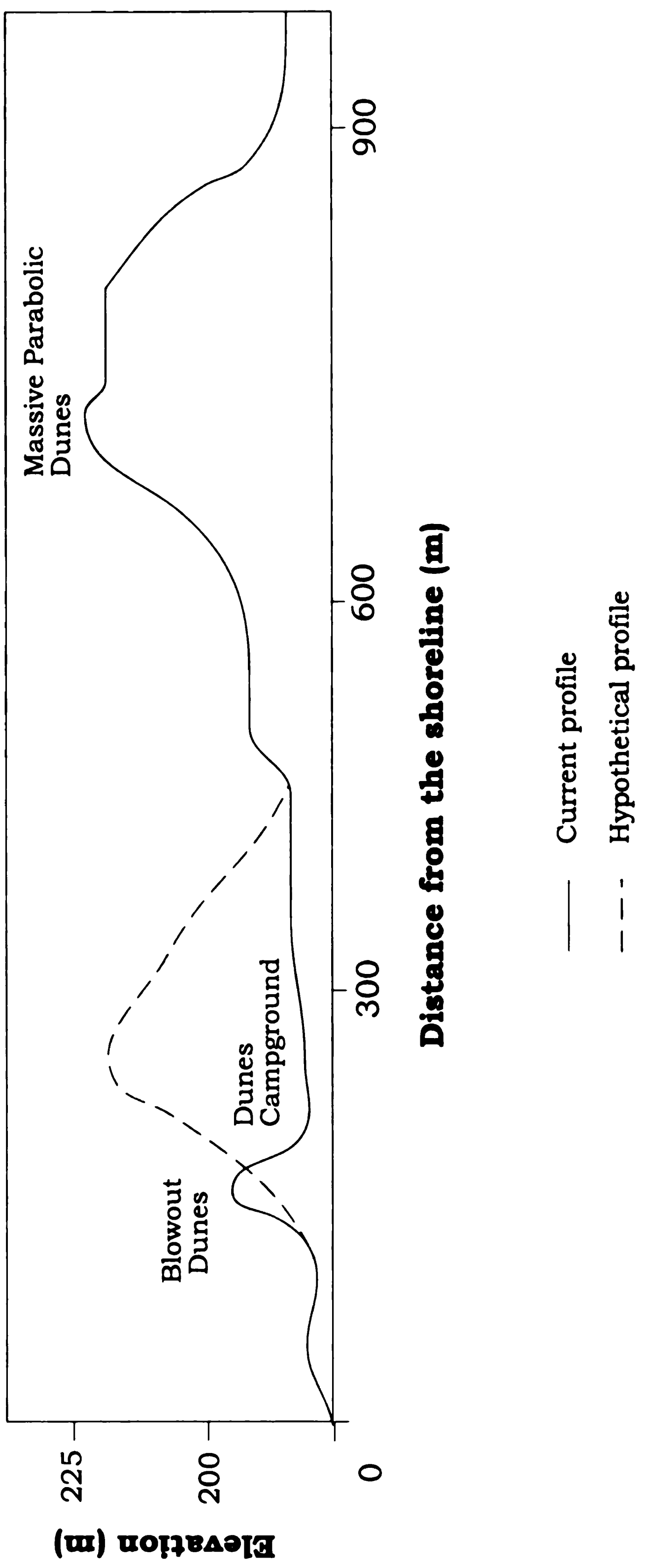

Figure 5:11. Cross section showing the position and size of the active dunes compared to the massive parabolic dunes. Dashed line shows hypothetical size of blowout dunes given that time and sediment supply remain similar to the conditions of formation of the massive parabolic dunes. Vertical exageration: 3x.

Comparison of the size of the massive parabolic dunes with the blowout dunes suggests that, given that adequate conditions are present along the shoreline (i.e. sand supply and minimum vegetative cover) the blowout dunes have the potential to reach the magnitude of the larger dunes in the park. Results from this study suggest that eolian processes have been active on the park area over the recent geological past and will continue to be active in the future, regardless of human intervention.

SUMMARY OF GEOMORPHIC HISTORY AND RELEVANCE TO CURRENT RESEARCH IN LAKE MICHIGAN COASTAL DUNES

In summary, evidence from this study indicates that multiple, yet predictable; cycles of eolian activity are responsible for the formation of the dunes at Petoskey State Park (Figure 5:12). Figure 5:12 shows that, at times, eolian processes were active across at least two landform assemblages. Radiometric evidence suggests that eolian processes were not uniform throughout landform assemblages and that several assemblages were probably active synchronously throughout the evolution of the park's landscape.

In general, the investigations conducted at Petoskey State Park reveal that dunes are older in the eastern part of the park and become progressively younger towards the west (Figure 5:5). A simplified cross section of the sand dunes showing subsurface distribution of major depositional assemblages is presented on Figure 5:13.

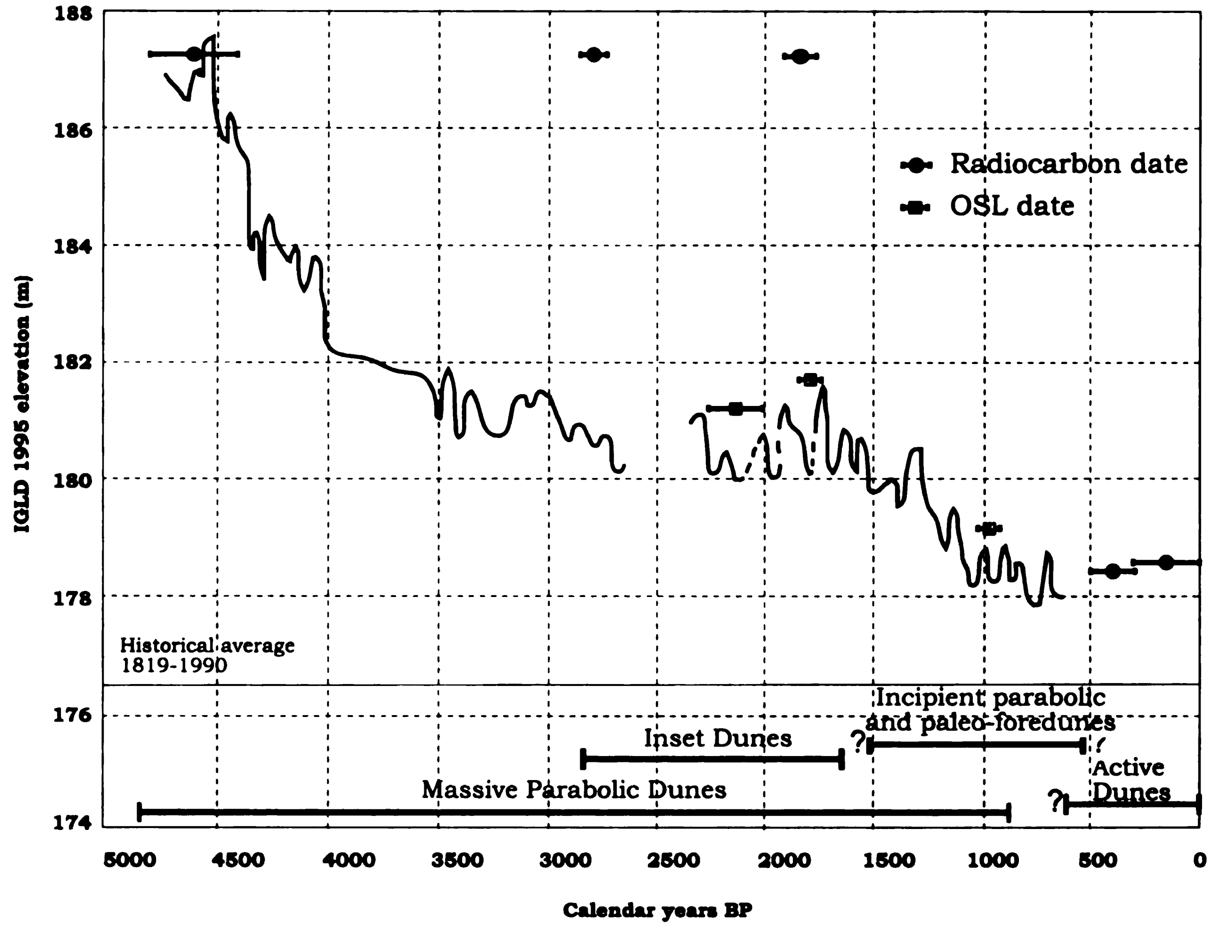

Figure 5:12. Relation between radiometric data and the relative late Holocene lake level for Sturgeon Bay, approximately 30 km north from Petoskey State Park. (Modified from Thompson and Baedke, 1997). Radiocarbon dates are plotted at the elevation at which the sample was taken.

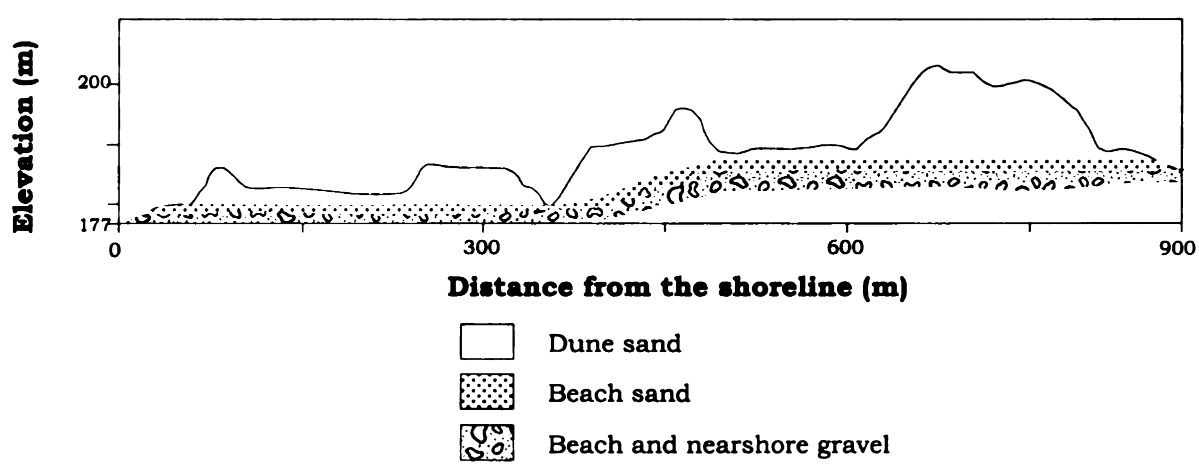

Figure 5:13. Schematic cross-section of the depositional units at Petoskey State Park. Depth of the gravel is unknown. Stratigraphic contacts estimated from sampled locations. Vertical exageration: 5x.

The impressive dune sequences present in the park are clear evidence of shoreline progradation and eolian vertical aggradation. The dune complex apparently developed in response to overall lake level fall during the late Holocene (Figures 5:12, 5:13). The protected location of Petoskey State Park, on Emmet Bay, allows for a positive sediment budget in this reach of the shoreline (Thompson and Baedke, 1997). Hence, vertical aggradation of eolian sediments is favored.

The overall geomorphic history of Petoskey State Park is very complex and is dominated by cycles of eolian activity and stability that occurred apparently in conjunction with lake level fluctuations. Such complexities result from the numerous factors that influence eolian and shoreline processes. As discussed in Chapter 2, a number of factors such as base level fluctuations, presence and type of vegetation cover, sediment availability, among others, influence periods of eolian activity (Carter, 1990). The results from this study suggest that eolian processes have been active across the park only in the past 4000 years. The spatial focus of eolian activity has shifted as a response to external factors. Evidence from this study also suggests that several landform assemblages were active synchronously (Figure 5:12).

The controls on cycles of stability and eolian activity are difficult to assess on the sole basis of the data provided by this study. Furthermore, a straightforward comparison of the radiometric dates obtained in this study with results from previous studies is not possible due to the limitations of radiocarbon and OSL dating. Additionally, inconsistent dating methods and materials throughout studies make correlations difficult. However, the results from this study can be reasonably compared with the conceptual models of Thompson and Baedke (1995, 1997), Arbogast and Loope (1999), and Loope and Arbogast (2000).

According to Thompson and Baedke (1995), foredunes probably initiate as beach ridges that are subsequently colonized by vegetation when lake levels drop. The pioneer vegetation acts as a trap for sediments, contributing to vertical aggradation and, therefore, to the formation of a dune cap on the ridge. A sequence of linear dunes may develop as lake level drops (if sediment supply is abundant). Based on numerous radiocarbon dates, Thompson and Baedke (1997) concluded that one beach ridge (and probably the overlying linear dune) form on some embayments along the shoreline about every 30 years. The presence of parabolic dunes in strandplains in Illinois is explained by erosion of the lakeward linear dunes and vertical aggradation of at least part of such dunes into the next landward (older) dune (Thompson and Baedke, 1995).

Arbogast and Loope (1999) and Loope and Arbogast (2000) specifically addressed the formation of parabolic dunes on lake terraces along the Lake Michigan shoreline. They associated the development of blowouts and vertical aggradation of parabolic dunes with quasi-periodic lake level fluctuations of the order of ~150 years. Their studies have made significant contributions to our understanding of the basic processes controlling shoreline behavior and coastal dune evolution. However, it is important to acknowledge that both of these interpretations about linear dunes (Thompson and Baedke, 1995, 1997) and parabolic dunes (Arbogast and Loope, 1999, Loope and Arbogast, 2000) are heavily dependent on the reliability and temporal resolution of radiocarbon evidence. Evidence from this study lacks the temporal resolution to test the hypothesis about the timing of parabolic dune aggradation. However, the magnitude of the parabolic dunes in the park, especially those in the eastern part, supports Loope and Arbogast's (2000) model, who hypothesized that parabolic dune aggradation occurs at the expense of foredunes during highstands (Figure 5:14). Comparison of the massive parabolic dunes and the linear and incipient parabolic dunes landform assemblages suggests that the slope of the nearshore probably plays an important role on the preservation of a sequence linear dunes versus a sequence of parabolic dunes (Figures 5:13 and 5:14). Slope controls the magnitude of the horizontal retreat of the shoreline as lake level falls. In areas such as the strand plains studied by Thompson and Baedke (1995, 1997) or Litcher (1998), where the slope of the nearshore is low, a small decrease in lake levels would cause a significant horizontal retreat of the shoreline, therefore enhancing the processes of beach ridge and posterior linear dune aggradation. At sites like Petoskey State Park, where the nearshore slope

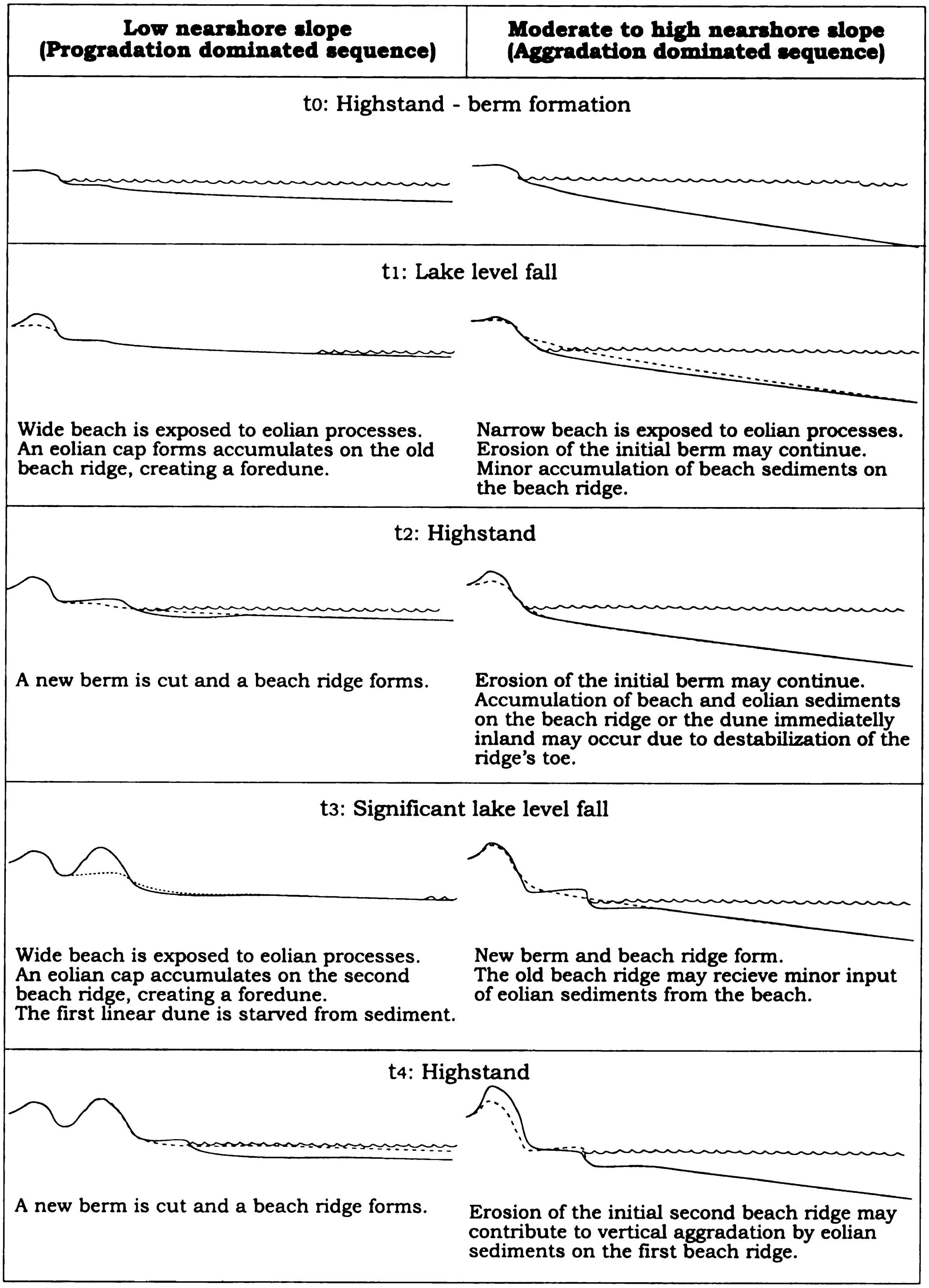

Figure 5:14. Schematic diagram showing the effect of the nearshore slope on dune aggradational and progradational sequences. Dashed lines show the profile of the previous stage. Repetition of these cycles over time yield to linear dunes sequences (left) or parabolic dune fields (right).

is relatively high, a more significant change in lake level is required to cause a horizontal retreat of the shoreline of sufficient magnitude for a new beach ridge (and subsequent foredune) to form. A small increase on lake level, on the other hand, would be enough for erosion of the foredune to occur. Destabilization at the base of the foredune would cause a rearrangement of the dune slopes due to gravitational processes. Such processes would most likely affect the vegetation cover on the foredune, hence allowing wind erosion of the foredune and transport and deposition of sand into the next dune.

Studies of modern shorelines have demonstrated that significant erosion of foredunes may occur as a response to a single storm event (LaMoe, 1987; Carter, 1988). However, the writer failed to find reports about significant blowout or parabolic dune aggradation related to such storm events. How would the current processes along the Petoskey shoreline be reconciled with the middle and late Holocene record? A possible explanation for this discrepancy could be that, because of their protected location (in bays and embayments), linear and parabolic dune fields along the shoreline display attenuated impacts of single storm events. Furthermore, a single storm event erodes foredunes in a matter of hours, removing the sediments and redistributing them along the shoreline. Because this happens in only a few hours, the ability of eolian processes to rework a significant volume of sediments into the older dunes is impaired, especially in the presence of wet sand.

Evidence from this study, as well as previous studies of coastal dunes (Livingstone and Warren, 1996), suggests that blowouts are indicative of a negative sediment budget on the dunes. As discussed earlier, blowouts may be initiated on a number of ways (e.g. due to storms, human impacts, an opening on the vegetation cover, among others). The early stages of blowout formation may occur even before the formation of the eolian cap itself, or anytime in the process of dune formation. Field observations along the foredune at Petoskey State Park revealed some localized areas (unvegetated, 1-2 m wide) where erosion appears to be the dominant process (Figure 5:15). Further development of such gaps into blowouts like those observed on the blowout dunes depends on the sediment budget on the dune. Vertical aggradation of the dunes probably occurs as at the expense of foredunes when lake levels are high (Thompson and Baedke, 1997; Loope and Arbogast, 2000). When lake levels are low, a new foredune may form. This foredune would act as a trap for sediments and the dune immediately inland from it would be starved of sediments. Evidently, this would translate into a negative sediment budget for the starved dune; hence, winds would have erosional power as they move over the windward face of this dune. Vegetation or topographic gaps on this dune would yield to enhanced wind erosion and blowout development. A hypothetical cycle of lake level rise and fall and its effect of eolian processes as complied from

Figure 5:15. Vegetation and erosional gaps in the foredune at Petoskey State Park.

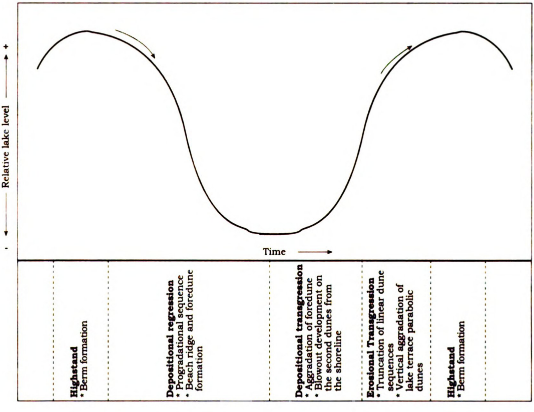

Figure 5:16. Hypothetical schematic cycle of lake level rise and fall and response of the shoreline-dune system. Compiled from Thompson and Baedke (1995), and Loope and Arbogast (2000) and this study.

Thompson and Baedke (1997), Loope and Arbogast (2000), and this study is presented on Figure 5:16.

In summary, dune formation at Petoskey State Park is closely related to lake level fluctuations and is, hence, cyclical. The cycle of dune formation initiates during highstands, when a berm is eroded. This berm could be potentially colonized by vegetation and act as a trap for sediments, initiating a foredune as lake level falls. Foredune aggradation continues as relative lake levels begin to rise. At this time abundant sediments are available. The second dune from the shoreline (a paleo-foredune) becomes starved of sediments due to vertical growth of the foredune and blowout development may occur on this paleo-foredune. Increasing rates of lake level rise may cause erosion. The magnitude of the erosion episode depends on the rate and magnitude of lake level rise as well as on the slope of the nearshore. Severe erosion may result on vertical aggradation of paleo-foredunes at the expense of the foredune. Repetition of this cycle at different scales may result on parabolic dune fields such as that at Petoskey State Park.

Chapter 6

CONCLUSIONS AND RECOMMENDATIONS

The goal of this study was to reconstruct the late Holocene geomorphic history of Petoskey State Park in the context of providing useful geomorphic information for park management. The investigations conducted at the park indicated that Petoskey State Park lies within a dune field that has been active for the past 4000 years. During this time multiple cycles of eolian activity occurred, some of which are preserved in the park. The focus of erosion and/or deposition has migrated throughout the park with time, probably in relationship to lake level fluctuations. Results from this study suggest that the dunes at Petoskey State Park constitute a progradational and aggradational sequence developed as a response to Lake Michigan fluctuations during the past 4000 years.

Five landform assemblages were mapped in the park, which have distinct morphologic characteristics and geomorphic histories, these are: (1) lake terrace, (2) massive parabolic dunes, (3) inset dunes, (4) linear and incipient parabolic dunes, and (5) active dunes. In a general sense, the older and larger dunes are the most inland ones; their age and size decreases towards the shoreline. Only the active dunes landform assemblage displayed active eolian processes at the time of this study.

Along these dunes, active blowouts are the dominant feature, which poses a problem for park management because they are encroaching on the Dunes Campground. No active eolian processes were observed on other landform assemblages. However, the linear and incipient dunes landform assemblage undergoes heavy use by park visitors and evidence of initial loss of vegetative cover was observed during this study, especially along park trails and close to campgrounds.

The previous conclusions have great implications for park management and planning. Evidence from this study suggested that large volumes of sand could move through this system over relatively short periods of time. Human activities in the park have the potential to accelerate these processes from a geologic time scale into a human time scale, causing serious damage to park infrastructure. Efforts by park authorities to minimize these effects include extensive planting of protective beach grass over the blowouts and foredune, as well as placing numerous warnings asking park visitors to avoid climbing the dunes. However, the problems have persisted and additional measures should be taken should current stabilization efforts fail:

1. Improve fencing structures to prevent access to the blowout dunes.
2. Provide users of the Dunes Campground with a short access to the beach. Establishing a new route would prevent users from seeking shortcuts to the beach through the dunes. This would

be particularly effective because that type of pedestrian traffic contributes to loss of vegetation cover from the dunes.

3. Expansion of the park toward less sensitive areas. If possible, expansion efforts should be directed to the Tannery Creek Campground, especially because this campground is located in a less sensitive area than the Dunes Campground.

Great potential for future studies exists on Lake Michigan's dune fields. Several aspects of coastal dune formation are still unclear after decades of research:

1. What are the factors controlling site - specific dune development? Studies conducted on Lake Michigan dunes have concentrated on the regional perspective. Previous studies have demonstrated that, on a regional scale, lake level fluctuations are the primary controlling factor on dune formation (Olson, 1958c; Thompson, 1992; Lichter, 1995; Thompson and Baedke, 1995; Petty et al, 1996a; Petty et al, 1996b; Lichter, 1997, Thompson and Baedke, 1997, Arbogast and Loope, 1999; Loope and Arbogast, 2000). However, detailed studies of specific parabolic dune fields need to be conducted in order to understand a better predict local aspects of dune formation and its impact on coastal management.
2. Are there modern analogs to the processes described by previous studies? If "the present is key to the past",

understanding current processes on Lake Michigan sand dunes, may help understand the processes that lead to the current configuration of dune fields along the shoreline. Future research should address modern as well as ancient processes. Issues such as the blowout initiation and growth, vertical aggradation on parabolic dunes, and timing of development of transgressive dune fields, among others need to be addressed by future research. The relations between these processes and lake level changes need to be established based on current observations.

REFERENCES

Alfred, S.D., Hyde, A., and R.L. Larson, 1973. Soil Survey of Emmet County, Michigan. Soil Conservation Service, United States Department of Agriculture in Cooperation with Michigan Agricultural Experiment Station, p. 1-96.

Arbogast, A., and W. Loope, 1999. Maximum limiting Ages of Lake Michigan Coastal Dunes: Their correlation with Holocene Lake Level History. Journal of Great Lakes Research, v. 25 (2), p. 372-382.

Barrett, L.R. and R.J. Schaetzl, 1998. Regressive Pedogenesis Following a Century of Deforestation: Evidence for Depodzolization. Soil Science, v. 163, p. 482-497.

Brown, D., 1998. Digital Terrain Analysis Geo 428 Course Packet. Department of Geography, Michigan State University, p. 3-1 to 3-5.

Buckler, W., 1979. Dune type inventory and barrier dune classification study of Michigan's Lake Michigan shore, Michigan Geological Survey Report of Investigation 23.

Calver, J.L., 1946. The Glacial and Postglacial history of the Platte and Crystal Lake depressions, Benzie County, Michigan. Occasional papers on the Geology of Michigan, Publication 45, Michigan Department of Conservation, Geological Survey Division.

Carter, R., 1988. Coastal Environments. Academic Press, London, 617 p.

Carter, R., K. Nordstrom, and P. Psuty, 1990. The study of coastal dunes. In: K. Nordstrum, N. Psuty, and R. Carter, editors. Coastal Dunes: form and processes. John Wiley and Sons, New York, p. 1-11.

Cleland, N.N., 1979. A reconstruction of post-glacial vegetation in the Northwest portion of Michigan's Lower Peninsula. Master's Thesis, Department of Botany and Plant Pathology, Michigan State University, p. 1-15.

Cowles, H., 1899. The ecological relations of the vegetation on the sand dunes of Lake Michigan. Botanical Gazette, v. 27, p. 95-117, 167-202, 281-308, 361-391.

Curray, J., 1964. Transgressions and Regressions. In: R. Miller ed., Papers in Marine Geology, Macmillan, New York, p. 167-189.

David, P., 1978. Why dunes are parabolic: the wet-sand hypothesis. Geological Association of Canada and Geological Society of America, Annual Meetings, Toronto, Ont., Abstracts with Programs, v. 3, p. 385.

Delcourt, P., W. Petty, and H. Delcourt, 1996. Late Holocene formation of Lake Michigan beach ridges correlated with a 70-yr. oscillation on global climate. Quaternary Research, v. 45, p. 321-326.

Dreimanis, A., and P.F. Karrow, 1972. Glacial History of the Great Lakes – St. Lawrence region, the classification of the Wisconsin(an) Stage, and its correlatives. In: 24th International Geological Congress, Montreal, Quaternary Geology, sec. 12, p. 5-15.

Dorr, J., and D. Eschman, 1970. Geology of Michigan, The University of Michigan Press: Ann Arbor, Michigan.

Easterbrook, D.J., 1999. Surface Processes and Landforms. Prentice Hall, Upper Saddle River, New Jersey, p. 471-488.

Eschman, D.F., 1985. Summary of the Quaternary History of Michigan, Ohio, and Indiana.. Journal of Geological Education, v. 33, p. 161-167.

Eschman, D.F., and Karrow, P.F., 1985. Huron Basin Glacial Lakes: A Review. In: P.F. Karrow and P.E. Calkin Eds. Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30, p. 79-93.

Evenson, E.B., Farrand, W., Eschman, D.F., Mickelson, D.M., and L.J. Maher, 1976. Greatlakean Substage: A replacement for the Valderan Substage in the Lake Michigan Basin. Quaternary Research, v. 6, p. 411-424.

Farrand, W.R., and D.L. Bell, 1982. Map of Quaternary Geology of Southern Michigan. MDNR, Geologic Survey Division.

Farrand, W.R., and Eschman, D.F., 1974. Glaciation of the Southern Peninsula of Michigan: A Review. Michigan Academician, v. 7, p. 31-56.

Goldsmith, V., 1978. Coastal Dunes. In: R. Davis Ed., Coastal Sedimentary Environments, Springer-Verlag, New York, p. 171-203.

Hansel, A.K., Mickelson, D.M., Schneider, A.F., and Larsen, C.E., 1985. Late Wisconsinan and Holocene History of the Lake Michigan Basin. In: P.F. Karrow and P.E. Calkin Eds. Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30, p. 40-53.

Hansel, A.K., and D.M. Mickelson, 1988. A Reevaluation of the Timing and Causes of High Lake Phases in the Lake Michigan Basin. Quaternary Research, v. 29, p. 113-128.

Hough, J.L., 1958. Geology of the Great Lakes. University of Illinois Press, Urbana, 313 p.

Hyde, R. and R. Watson, 1983. Radiative and meteorological control on the movement of sand at Lake Mungo, New South Wales, Australia. In: M. Brookfield and T. Ahlbrandt eds., Developments in Sedimentology 38: Eolian sediments and processes, Elsevier, New York, p. 311-324.

Karrow, P.F., Anderson, T.W., Clarke, A.H., Delorme, L.D., and M.R. Sreenivasa, 1975. Stratigraphy, Paleontology, and Age of Lake Algonquin Sediments in Southwestern Ontario, Canada. Quaternary Research, v. 5, p. 49-87.

Larsen, C.E., 1985. Lake Level, Uplift, and Outlet incision, the Nipissing and Algoma Great Lakes. In: P.F. Karrow and P.E. Calkin Eds. Quaternary Evolution of the Great Lakes. Geological Association of Canada Special Paper 30, p. 63-77.

Larson, G., Lowell, T., and N. Ostrom, 1994. Evidence for the Two Creeks interstade in the Lake Huron basin. Canadian Journal of Earth Sciences, v. 31, p. 793-797.

Leverett, F., and F.B. Taylor, 1915. The Pleistocene of Indiana and Michigan and the History of the Great Lakes. United States Geological Survey Monograph 53, 529 p.

Litcher, J., 1995. Lake Michigan beach ridge and dune development, lake level and variability in regional water balance. Quaternary Research, v. 44, p. 181-189.

Litcher, J., 1997. AMS radiocarbon dating of Lake Michigan beach ridge and dune development. Quaternary Research, v. 48, p. 137-140.

Litcher, J., 1998. Primary succession and forest development on coastal Lake Michigan sand dunes. Ecological Monographs, v. 68, p. 487-510.

Livingstone, I., and A. Warren, 1996. Aeolian Geomorphology: An Introduction. Addison Wesley Logman Limited, Singapore, p. 64-101.

Loope W., and A. Arbogast, 2000. Dominance of a ~150 year cycle of sand supply change in late Holocene dune-building along the eastern shore of Lake Michigan. Quaternary Research, v. 54, p. 414-422.

McKee, E., 1979. A study of global sand seas: U.S. Geological Survey Professional paper 750-D, p. 108-114.

Martin, H.M., 1957. Outline of the Geologic History of Cheboygan County. Michigan Department of Conservation - Geological Survey Division, 6 p.

MDNR, 2000. http://www.midnr.com/ParkRegions.asp?linkID=185insert=1&id=2

Mickelson, D.M., Clayton, L., Fullerton, D.S., and Borns Jr., H.W., 1983. The Late Wisconsin Glacial Record of the Laurentide Ice Sheet in the United States. In: S.C. Porter Ed. Late Quaternary Environments of the United States. V. 1: The Late Pleistocene. University of Minnesota press, Minneapolis, p. 3-37.

Olson, J., 1958a. Rates of succession and soil changes on southern Lake Michigan sand dunes. Botanical Gazette, v. 119 p. 125-170.

Olson, J., 1958b. Lake Michigan dune development, 2: Plants as agents and tools in geomorphology. Journal of Geology, v. 66, p. 473-483.

Olson, J., 1958c. Lake Michigan dune development, 3: lake level, beach, and dune oscillations. Journal of Geology, v. 66, p. 473-483.

Petty, W., P. Delcourt, and H. Delcourt, 1996. Holocene lake level fluctuations and beach ridge development along the northern shore of Lake Michigan, USA. Journal; of Paleolimnology, v. 15, p. 147-169.

Pye, K., 1990. Physical and Human influences on coastal dune development between the Ribble and Mersey estuaries, northwest Engalnd. In: K. Nordstrum, N. Psuty, and R. Carter, editors. Coastal Dunes: form and processes. John Wiley and Sons, New York, p. 339-359.

Rieck, R.L., and H.A. Winters, 1993. Drift volume in the southern peninsula of Michigan – a prodigious Pleistocene endowment. Physical Geography, v. 14, p. 478-493.

Schaetzl, R., 2001. Late Pleistocene Ice Flow Directions and the Age of Glacial Landscapes in Northern Lower Michigan. Physical Geography, v. 22: in press.

Schaetzl, R., Drzyzga, S., Weisenborn, B., Kinkare, K., Cordoba, X., Shein, K., Dowd, C., and Linker, J., (in revision). Correlation and mapping of Lake Algonquin shorelines in Michigan. Annals of the Association of American Geographers.

Schenk, C.J., 1990. Processes of Eolian Sand Transport and Deposition. In: S. Fryberger, L. Krystinik, and C. Schenk Eds., Modern and Ancient Eolian Deposits: Petroleum exploration and production, Rocky Mountain Section S.E.M.P., Denver, p. 2-1 to 2-4.

Schenk, C.J., 1990. Eolian Dune Morphology and Wind Regime. In: S. Fryberger, L. Krystinik, and C. Schenk Eds., Modern and Ancient Eolian Deposits: Petroleum exploration and production, Rocky Mountain Section S.E.M.P., Denver, p. 3-1 to 3-5.

Scott, I., 1942. The dunes of Lake Michigan and correlated problems. 44th Annual Report of the Michigan Academy of Science, Arts and Letters, v. 44, p. 53-61.

Sharkey, R.F., 1972. History of Petoskey State Park. MDNR internal document.

Sherman D.J. and S. Hotta, 1990. Aeolian sediment transport: theory and measurement. In: K. Nordstrum, N. Psuty, and R. Carter, editors. Coastal Dunes: form and processes. John Wiley and Sons, New York, p.17 - 38.

Short, A. and P. Hesp, 1982. Wave, beach and dune interactions in southeast Australia. Marine Geology, v. 48, p. 259-284.

Soil Survey Staff, 1993. Soil Survey Manual. USDA Handbook No. 18. U.S. Govt. Print Office.

Stuiver, M. and P. Reimer, 1993. Radiocarbon calibration program revision 3.0. Radiocarbon, v. 35, p. 215-230.

Thompson, T.A., 1992. Beach ridge development and lake level variation in Lake Michigan. Sedimentary Geology, v. 80, p. 305-318.

Thompson, T. and S. Baedke, 1995. Beach ridge development in Lake Michigan: shoreline behavior in response to quasi-periodic lake level events. Marine Geology, v. 129, p. 163-174.

Thompson, T. and S. Baedke, 1997. Strand plain evidence for late Holocene lake level change variations in Lake Michigan. Geologic Society of America Bulletin, v. 109, p. 666-682.

Thwites, F.T., 1943. Pleistocene of part of northeastern Wisconsin. Geological Society of America Bulletin, v. 54, p. 87-144.

Zumberge, J.H., and W.S. Benninghoff, 1969. A mid-Wisconsin peat in Michigan, U.S.A. Pollen et Spores, v. 11, p. 585-601.