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AN APPLICATION OF OSL DATING TO TEST THE
PERCHED-DUNE MODEL ON COASTAL DUNES AT
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AN APPLICATION OF OSL DATING TO TEST THE PERCHED-DUNE MODEL ON COASTAL DUNES AT ARCADIA, MICHIGAN

By

Bradley Ethan Blumer

A THESIS

Submitted to
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ABSTRACT

AN APPLICATION OF OSL DATING TO TEST THE PERCHED-DUNE MODEL ON COASTAL DUNES AT ARCADIA, MICHIGAN

By

Bradley Ethan Blumer

Coastal sand dunes commonly occur along the northeastern shore of Lake Michigan. Many of these landforms are perched dunes that mantle tall (~ 100 m) coastal bluffs. It is believed that these dunes enlarge during high lake levels when waves destabilize bluffs and eolian sand is transported inland. Previous studies relied heavily on radiocarbon dating of buried soils to reconstruct dune chronologies. Given the uncertainties associated with this method, however, erroneous temporal linkages may have been made. As a result, this study tested the perched-dune model by reconstructing the history of eolian sand deposition at three exposures at Arcadia dunes in Benzie County, MI. I hypothesized that periods of dune growth are associated with high lake levels. This hypothesis was tested by directly dating eolian sand deposits using optical stimulated luminescence (OSL).

Results indicate that the Arcadia dunes mantle ~ 90 m of glacial sediment. Formed near the top of this deposit is a well developed paleosol that was buried by eolian sand ~ 4.5 ka. Episodes of dune growth subsequently occurred at ~ 3.5, 1.7, 1.0, 0.9, 0.7, 0.4 and 0.3 ka. Most of these episodes generally coincided with estimated high lake levels. One major period of dune construction clearly occurred during a lake regression, however, suggesting that perched dunes may also be supplied with sand during low lake phases. Although some synchrony in dune growth was observed across the field, some periods were asynchronous, suggesting local factors may play a role in sand deposition.

ACKNOWLEDGMENTS

I would like to extend my appreciation those agencies and individuals who have made this thesis possible. First I would like to thank Xiomara Cordoba-Lepczyk, Brad Gerlach, and Glen Chown from the Grand Traverse Regional Land Conservancy for providing funding, resources, and access to their beautiful property at Arcadia Dunes. I would also like to thank the Department of Geography at Michigan State University for their funding and resources; and Steven L. Forman at the University of Illinois – Chicago Luminescence Dating Research Laboratory for the radiometric dating.

This thesis is not only a product of my efforts, but the efforts of a group of people who generously contributed their time and knowledge to the project. Special thanks are extended to my committee members: Alan Arbogast, Randy Schaetzl, and Catherine Yansa for their insights and support throughout all phases of this study. I would also like to thank my colleagues in the Department of Geography for their help, including Kristy Stanley, Lesley Fusina, and Briana Walker. I especially thank Trevor Hobbs, Crosby Savage, and Mark Devisser for their work in the field. Their assistance in this research is appreciated as much as making the experience of fieldwork enjoyable and unforgettable.

Finally, I would like to thank and dedicate this thesis to my parents, family, and friends who have always supported me throughout my endeavors. No matter where I am your support always makes me feel at home. You have given me inspiration, integrity, guidance, and most of all the motivation which keeps me moving forward.

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CHAPTER I

INTRODUCTION AND PROBLEM DEFINITION

Coastal sand dunes are common along the eastern shores of the Great Lakes, due to prevailing westerly winds and an abundant sand supply (Figure 1:1; Arbogast, 2009). These dunes have been the focus of geomorphic and geologic research at many prominent locations (Figure 1:1) for over 100 years. The earliest studies focusing on Lake Michigan's dunes were largely qualitative in nature, with a focus on describing basic geomorphic characteristics and ecology (Cowles, 1899; Dow, 1937). Olson (1958 a, b) was the first to associate dune mobilization with an initiating event, specifically the association of foredune formation with low lake levels. This model was generically applied to the larger coastal dunes, along with a general assumption that the dunes essentially formed during the waning phases of the Nipissing high stand, which was a high phase of Lake Michigan that occurred approximately 5 – 4 thousand years ago (Dorr and Eschman, 1970; Buckler, 1979). This hypothesis remained largely untested until the late 20th century.

Beginning in the 1990s, a new generation of researchers began to systematically reconstruct the evolution of the large coastal dunes along Lake Michigan. This new interest focused on detailed stratigraphic investigations and age determination of buried organic deposits, using radiocarbon (^{14}C) dating of buried soils to test the prevailing hypothesis (e.g., Dorr and Eschman, 1970; Buckler, 1979) by correlating periods of eolian activity with reconstructed lake-level curves (e.g., Thompson, 1992; Thompson and Baedke, 1995; Thompson and Baedke, 1997; Baedke and Thompson, 2000).

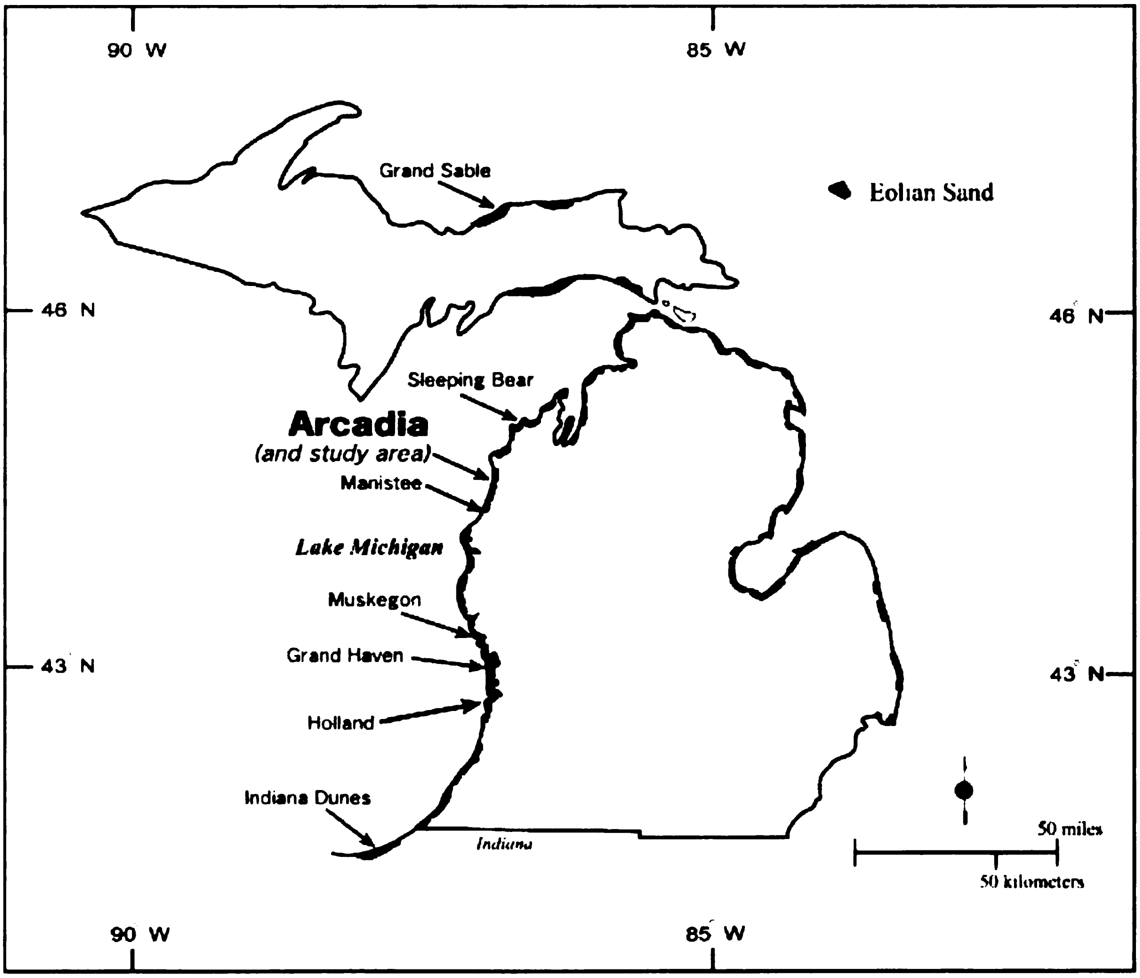

Figure 1:1 – Prominent areas of coastal sand dunes along the eastern shore of Lake Michigan and the southern shore of Lake Superior. Notable dunes and previously studied areas are named, along with the general location of the study area north of Arcadia, Michigan (modified from Lepczyk and Arbogast, 2005).

The studies of the 1990s indicate that large coastal dunes can be subdivided into high-perched systems that mantle glacial headlands and low-perched dunes that cover topographically lower lake plains (Arbogast, 2009). Radiocarbon dating of buried soils in high-perched dune fields indicate that the onset of dune construction correlates to high lake levels during the Nipissing, and has continued periodically until recent times (Snyder, 1985; Marsh and Marsh, 1987; Anderton and Loope, 1995). Also, observations of active perched dunes indicate that they are usually supplied by sand from glacial sediment in the upper portions of coastal bluffs (Dow, 1937; Marsh and Marsh, 1987). These studies led to the development of the perched-dune model, which states that coastal dunes are mobilized during high lake levels when waves can undercut and destabilize the bluff (Marsh and Marsh, 1987; Anderton and Loope, 1995).

Other studies have focused on the low-perched dunes that dominate Lake Michigan's southern and eastern shores. Radiocarbon dating from a number of studies confirms that the dunes first formed during the Nipissing high stand (Arbogast and Loope, 1999; Arbogast et al., 2002a; Hansen et al., 2002). In contrast to prior assumptions (e.g., Dorr and Eschman, 1970; Buckler, 1979), however, these investigations indicate that dunes grew episodically throughout much of the late Holocene, with many periods of dune growth actually occurring during high lake levels (e.g., Arbogast and Loope, 1999; Van Oort et al., 2001; Hansen et al., 2002; Arbogast et al., 2004; Fisher and Loope, 2005; Lepczyk and Arbogast, 2005). These conclusions suggest that the relationship between lake levels and low *or* high-perched dune activity is essentially the same, and supports the widespread application of the perched-dune model on Lake Michigan's coastal dunes.

Most of the studies conducted thus far on coastal dune fields along Lake Michigan have based their conclusions largely on radiocarbon dates from paleosols which are assumed to generally represent the time of burial by eolian sand. However, these conclusions may be erroneous because radiocarbon dating only provides a maximum limiting age of burial, i.e. the eolian sand can be no older than the paleosol on which it is deposited. Also, organic matter in the A horizon may be much older than the time of burial, become mixed, or contaminated due to the translocation of humates downward into the paleosol, thereby providing erroneously younger ages for this buried soil.

Given the uncertainties associated with radiocarbon dating in eolian environments, a more accurate test of dune construction may choose to focus on the direct timing of eolian sand deposition. Such a record can be constructed using Optically Stimulated Luminescence (OSL) dating, which dates the time of deposition of dune sand rather than the time of burial of organic matter (e.g., Bailey et al., 2001; Roberts and Plater, 2007). By using OSL, samples from different locations within the same eolian unit can be dated, which helps bracket its time of deposition. This sampling method allows for the collection of samples outside of buried soils, which may be poorly formed, discontinuous, or absent.

In addition to possible erroneous interpretations (e.g., Bailey et al., 2001; Roberts and Plater, 2007) of radiocarbon dates in dune environments, there is also a spatial gap in the chronological record of dune construction. Many studies have been conducted on low-perched dunes along the southeastern shore of Lake Michigan (e.g., Arbogast and Loope, 1999; Van Oort et al., 2001; Hansen et al., 2002; Arbogast et al., 2004; Fisher and

Loope, 2005; Lepczyk and Arbogast, 2005). In contrast, few studies have focused on high-perched dunes of northwest Lower Michigan despite the fact that they are some of the most prominent dunes along the Lake Michigan shoreline (Snyder, 1985). Investigating their formation would further clarify the history of eolian sand deposition along Lake Michigan.

The purpose of this thesis is to test the validity of the perched-dune model proposed by researchers (e.g., Anderton and Loope, 1995; Arbogast and Loope, 1999, Loope and Arbogast, 2000, Arbogast et al., 2002a). In that context I will test the hypothesis that perched dunes grow due to an increase in sand supply during high lake stages using OSL dates from eolian sand units in a high-perched dune field north of Arcadia in northwest Lower Michigan (Figure 1:1). This is the first study in which OSL dating has been exclusively used to establish a depositional history of high-perched dunes along Lake Michigan. As such, it will contribute to the rapidly emerging record of coastal dune evolution along the Lake Michigan shore.

CHAPTER II

RELEVANT RESEARCH

This chapter discusses relevant literature pertaining to coastal dune formation along the eastern shore of Lake Michigan. The discussion will begin with early work in eolian processes and landforms, with a focus on the world's deserts and interactions between wind and sand (e.g. Bagnold, 1941). A brief discussion of post-glacial Lake Michigan lake levels follows, which provides the context for dune studies along the coast. Following the glacial-lake discussion, the chapter moves to coastal dune research along Lake Michigan, with an initial focus on early studies associated with foredunes and plant succession (e.g., Cowles, 1899; Olson, 1958a, b, c). Subsequently, the chapter shifts to the recent research associated with coastal dune chronology and evolution (e.g., Anderton and Loope, 1995; Arbogast and Loope, 1999, Loope and Arbogast, 2000; Arbogast et al., 2002a). Finally, international coastal dune studies are discussed, with special attention paid to the methods used to date eolian sand.

Eolian Processes

In the context of this investigation, the discussion of eolian processes focuses primarily on the sand fraction of sediment. Given that sand dunes are products of wind erosion and deposition, it is first necessary to discuss eolian processes as they relate to the movement of sand particles. In this context, this part of the chapter begins with the fluid nature of air. Subsequently, it moves to dune formation and classification.

Like water, air is a fluid capable of transporting sediment (Bagnold, 1941). In order for sediment to mobilize, wind velocity must overcome a fluid threshold greater than that of resisting forces such as gravity, friction, and cohesion. Small particles, such as clay and silt (< 2 μm to .05 mm), may become *suspended* at low velocities and travel long distances due to their relatively low fluid threshold (< 30 cm/sec; Bagnold, 1941). In contrast, sand grains (0.05 to 2.0 mm) have a relatively high fluid threshold between ~ 35 to 64 cm/sec. As a result, sand is usually transported short distances, making it much more important than silt or clay to the formation of dunes

The mobilization of eolian sand follows a predictable pattern. Wind tunnel experiments demonstrate that as velocity increases, sand grains begin to roll downwind in a process called *creep* (Figure 2:1[*]; Bagnold 1941). If wind velocity increases further, the particles begin to bounce downwind through the process of *saltation*. Saltating grains travel through the air in trajectories based on their density, shape and mineralogy (Bagnold, 1941; Gerety and Slingerland, 1983; Greeley et al., 1983). Additional sand will become entrained when saltating particles impact immobile grains on the surface and force them upward into the windstream.

Formation of Sand Dunes

Eolian transport of sand occurs over relatively short distances, as long as wind speed remains above the fluid threshold. Deposition of wind-blown sand begins locally when obstacles such as vegetation or man-made structures cause wind speeds to decrease (Olson, 1959b). As additional sand is deposited it begins to acquire a characteristic dune

[*] Images in this thesis are presented in color.

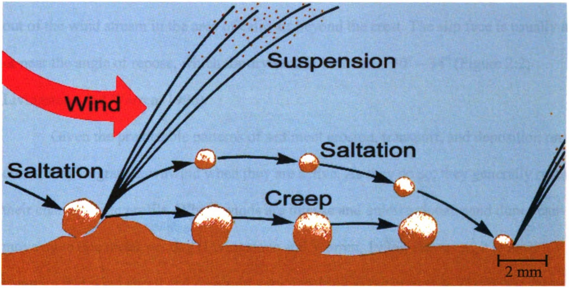

Figure 2:1 – Methods of eolian transport by suspension, saltation, and surface creep. Wind can cause creeping of sand grains which leads to saltation. Other particles ejected during saltation may become suspended in the air and deflate out of the area. (Modified from WERU, 2007).

shape with three common profile elements: the *backslope*, *crest* and *slip face* (Figure 2:2). The backslope is a shallow (generally 8° to 13°) windward slope where erosion and sand transport dominate (Livingstone and Warren, 1996). At the peak of the dune is the crest, where erosion is in equilibrium with deposition. Leeward of the crest is the slip face, which is a steeper slope where sand deposition dominates because sand grains fall out of the wind stream in the area of calm air beyond the crest. The slip face is usually at or near the angle of repose, which for dry sand is between ~ 30° – 34° (Figure 2:2; Livingstone and Warren, 1996).

Given the predictable patterns of sediment erosion, transport, and deposition on dunes, they migrate downwind when they are active. As they do so, they generally retain their characteristic profile. Where winds are intense and unidirectional, sand dunes can move up to one meter per day (Livingstone and Warren, 1996). However, it is more common that wind speed and direction vary, resulting in the numerous diagnostic forms that are used to classify sand dunes (Figure 2.3). These various forms are basically dependent on three variables, including 1) wind speed/direction, 2) the amount and type of vegetation cover, and 3) sand supply (Livingstone and Warren, 1996).

In the context of dune classification, Livingstone and Warren (1996) identify two general classes of dunes: *free* and *anchored*. As the name implies, free dunes form in mostly dry, unvegetated regions (i.e. deserts) where mobilization easily occurs. These dunes are self-sustaining, and usually form in areas with little topographic or vegetative obstruction (Livingstone and Warren, 1996). A variety of dune forms occur within the category of free dunes. *Transverse dunes* are ridges of sand oriented perpendicular to unidirectional winds. These dunes typically develop where sand supplies are high and net

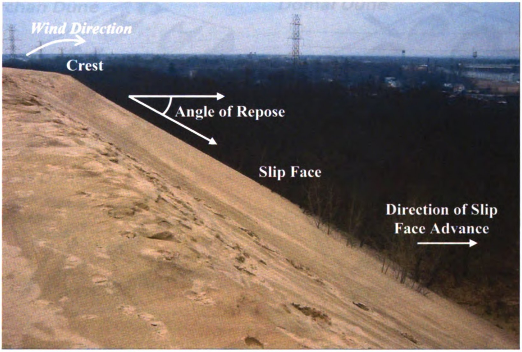

Figure 2:2 – Angle of repose and slip face on the lee slope of Mt. Baldy, Indiana Dunes National Lakeshore. Sand is blown over the crest where it is deposited on the slip face, which is at or near the angle of repose. Trees in this photograph are ~10 meters tall. Photographed by author.

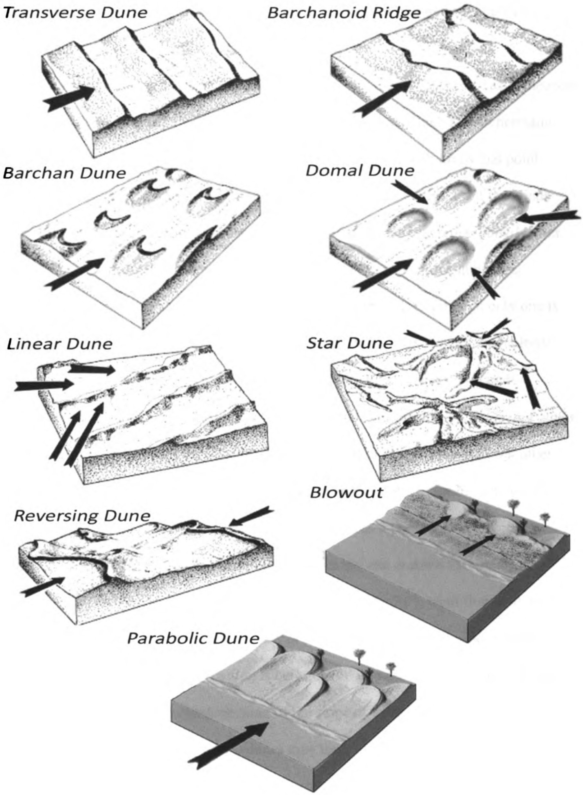

Figure 2:3 – Generalized dune forms, arrows indicate prevailing wind direction. (Modified from McKee, 1979; Tarbuck et al., 2005).

sand transport is normal to their crest, which forms slip faces oriented in roughly the same direction (Livingstone and Warren, 1996).

As sand supplies decrease *barachnoid ridges* may form, which have asymmetrical crests. This dune form is a transitional stage to *barchan dunes*, which form when sand supply is relatively scarce. Barchan dunes are crescent shaped, with limbs that point downwind, and generally form on firm surfaces such as desert pavement. If the winds become multi-directional the limbs may erode, creating oblong or circular *domal dunes.* *Linear dunes* form when winds are oriented parallel to the crest. These dunes are distinctive because they have slip faces on either side of the ridge, although only one is typically active at any time (Livingstone and Warren, 1996). *Star dunes* have multiple ridges and slip faces radiating from a central peak. They form in areas with multi-directional winds and high sand supply (Livingstone and Warren, 1996). *Reversing dunes* form where opposing winds persist, creating two slip faces oriented toward each other.

In contrast to free dunes, Livingstone and Warren (1996) described anchored dunes as those which form around fixed obstacles, such as topographic obstructions. Large topographic obstructions, such as mesas or scarps, can create a variety of topographically anchored dunes. If the scarp is very steep (> 50°), wind flow will diverge at the base of the obstruction, creating a low pressure area in which an *echo dune* may form. When the slope of the obstruction is between 30° and 50°, sand is deposited against the scarp in a *sand ramp* or *climbing dune*. If the slope of the scarp is below 30° (or if a sand ramp is already present) *cliff-top* dunes may form on top of the obstruction. If the obstruction also has a lee slope (e.g. a mesa), sand may accumulate into *falling dunes* in the area of calm air behind the obstruction. If the obstruction is relatively small, air may simply flow around the obstacle, depositing *lee dunes*, similar to the limbs of a barchan dune (Livingstone and Warren, 1996).

The dunes described to this point share the general character of being unvegetated, which allows sand to move freely. In contrast to these dunes, a few dune forms are distinctive because they are partially vegetated, which causes some predictable morphologies. The most basic vegetated dune form is a *blowout* (Figure 2:3), which is a saucer-shaped depression that develops when eolian erosion occurs on a local scale. Once a blowout is initiated, wind flowing over the dune is channeled towards the apex of the blowout, causing downwind elongation (Carter, 1988). Young or poorly developed blowouts begin as saucer shaped depressions, and eventually extend into long, narrow troughs (Hesp, 2002). Vegetative cover of the surrounding landscape tends to limit the expansion of blowouts.

If winds strengthen and the blowout enlarges a *parabolic dune* may form. This type of dune is distinctive because it has an open-faced, curved shape with two long arms anchored near the original blowout position (Figure 2:3; Livingstone and Warren, 1996). Vegetation anchors the arms against wind erosion, which forces wind down the concave backslope and over the crest (Livingstone and Warren, 1996). Sand is deposited on the slip face of the convex lee side, and rests at or near the angle of repose (Figure 2:2, 2:3). Parabolic dunes can be classified by their length to width ratio; ratios less than 0.04 are 'lunate'; between 0.4 and 1.0 are 'hemicycle'; between 1.0 and 3.0 are 'lobate'; and greater than 3.0 are 'elongate' (Pye, 1990).

In summary, entrainment of sand particles in wind occurs at relatively high velocities due to the high fluid threshold of sand. As a result of this relationship, sand is mostly transported via *creep* and *saltation*, which can be easily interrupted due to a decrease in wind velocities around obstructions. As sand is deposited it acquires a characteristic shape with three profile elements, the *backslope*, *crest*, and *slip face*. Dunes that are self propagating, such as those in sandy deserts, are called *free*. In contrast, *anchored* dunes form around either topographic obstructions or are influenced by vegetation. Both free and anchored dunes are classified by their shape in plan view, and comprise most of the basic dune morphologies found globally.

Lake Michigan Lake Level Studies

The formation of coastal dunes along Lake Michigan is directly related to post-glacial lake-level changes in the basin (e.g., Snyder, 1985; Anderton and Loope, 1995; Arbogast and Loope, 1999; Arbogast et al., 2002). This section focuses on the deglacial history of the Great Lakes region and the reconstruction of Holocene lake levels, which provide the context for my coastal dune research.

Early Holocene Geologic History

During the late Pleistocene, the Great Lakes region was dominated by the Laurentide Ice Sheet, which reached its most southerly extent in Ohio and Indiana approximately 20,000 years ago (Larson and Schaetzl, 2001). By about 16,000 years ago the ice sheet, which had fully occupied the Lake Michigan basin, began to recede to the north (Figure 2:4A; Larson and Schaetzl, 2001). Subsequent readvances of the Lake Michigan lobe covered much of the northern part of the basin, including northwest Lower

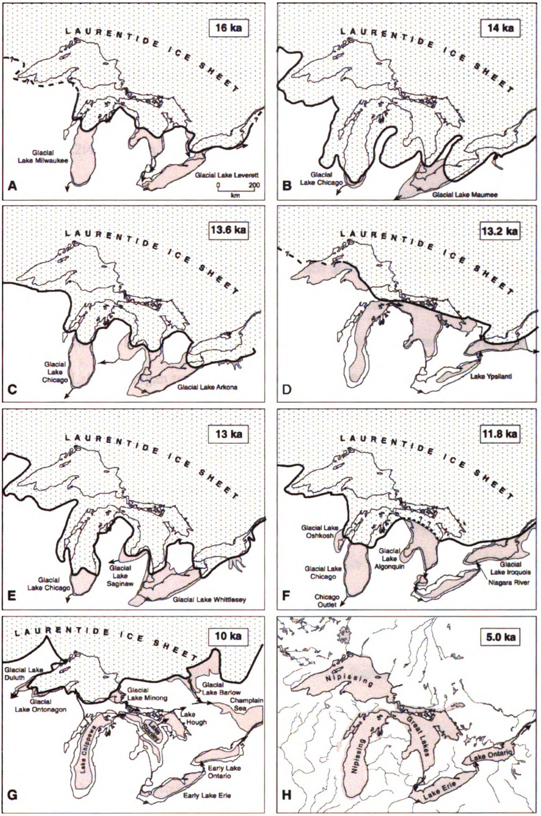

Figure 2:4 – Locations and general chronology of the major proglacial lakes during the retreat of the Laurentide Ice Sheet (modified from Larson and Schaetzl, 2001).

Michigan. This complex history resulted in a series of proglacial lakes that occupied the Lake Michigan basin (Figure 2:4B-E). The last advance that reached northern Lower Michigan was the Greatlakean about 14,000 cal. yr. B.P.[*] (Evenson et al., 1976). At this time, pro-glacial Lake Chicago was at its highest phase and drained through the Chicago outlet (Figure 2:4F; Larsen, 1987).

Once ice receded north of the Mackinac Straits (~ 13,000 cal. yr. B.P.), Lakes Michigan and Huron became confluent, forming glacial Lake Algonquin (Larsen, 1987). This high lake phase was short-lived and ended when the topographically low North Bay outlet was uncovered by retreating glacial ice about 12,080 cal. yr. B.P. (Larsen, 1987). The rapid draining of Lake Algonquin through the isostatically depressed North Bay outlet caused Lake Michigan to enter the Chippewa low phase (Fig 2:4 G; Mickelson et al., 1982; Larsen, 1987; Larson and Schaetzl, 2001). At the peak of this phase, Lake Michigan was at least ~ 60 m lower than present (Hansel et al., 1985).

As Lake Michigan reached its lowest Chippewa elevation, isostatic uplift of the North Bay outlet slowly caused the Great Lakes to rise during the Nipissing transgression (Larsen, 1987). By approximately 9,000 cal. yr. B.P. Lakes Michigan and Huron became confluent once again (Larsen, 1987). The Nipissing transgression ultimately reached its highest level at about 5,300 cal. yr B.P. (Larsen, 1994) and is estimated to have been about five meters above historical average (Figure 2:4H, Baedke and Thompson, 2000).

[*] Calibrated years before present (calibrated radiocarbon age before 1950 A.D.). All radiocarbon dates have been calibrated by the author from radiocarbon years using Fairbanks0107 calibration curve (Fairbanks et al., 2005), unless calibrated ages were given by the original source. A standard deviation of 100 years was assumed for approximate ages or if the standard deviation was missing. Uncalibrated approximate ages are not given in text.

Late Holocene Lake-Level Studies

The end of the Nipissing transgression marks the point in the history of Lake Michigan where lake level changes associated directly with post-glacial isostatic adjustments ended. By about 5,150 cal. yr. B.P. isostatic rebound caused the abandonment of the North Bay outlet in favor of the current drainage route through Port Huron (Larsen, 1987). By 3,530 cal. yr. B.P. (Larsen, 1987) incision of the Port Huron outlet caused lake levels to drop to near modern levels (Figure 2:5; Baedke and Thompson, 2000). Lake levels rose once again approximately 3,000 cal. yr. B.P. to the Algoma high stand, which may have been caused by stabilization of the Port Huron outlet, or due to climatic factors (Figure 2:5; Larsen, 1985). Lake levels fell approximately 2,250 cal. yr. B.P. and rose again to above normal levels from 1,900 to 1,100 cal. yr. B.P. (Baedke and Thompson, 2000). Since then, lake levels have stabilized to within ± 1 m of historical yearly averages (Figure 2:5; Baedke and Thompson, 2000).

A variety of studies have contributed to the current record of Late Holocene lake level changes, with most of them associated with the reconstruction of beach-ridge ages and elevation. The first of these studies was by Thompson (1992), who studied a strandplain along the southern shore of Lake Michigan at Tolston Beach in northwestern Indiana. He found the age of ridges in the ridge-swale complex decreased lakeward. This observation, coupled with elevation data, was used to reconstruct a lake-level curve that shows an overall downward trend in the level of Lake Michigan during the late Holocene. This curve reflects the lowering of lake levels following the Nipissing phase of Lake Michigan at approximately 4,200 cal. yr. B.P. By about 3,100 cal. yr. B.P. lake levels

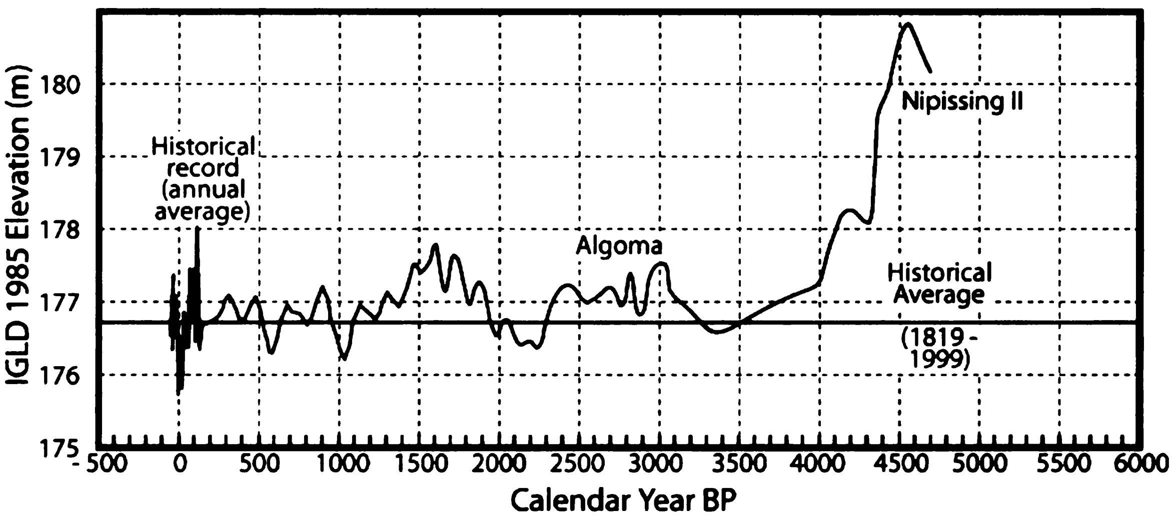

Figure 2:5 – Late Holocene lake-level estimations for Lake Michigan (Modified from Baedke and Thompson, 2000).

increased, possibly representing the Algoma high phase. Thompson (1992) noted other high stands occurring at about 2,300, 1,700, and 600 cal. yr. B.P.

Thompson's (1992) study indicated that ridges are constructed in a cyclic nature in response to quasi-periodic lake-level fluctuations. The largest fluctuation is approximately 1.8 – 3.7 meters every 500 to 600 years (Thompson, 1992). Nested within this cycle are meso-scale fluctuations of 0.8 – 0.9 meters every 140 – 160 years. The meso-scale cycle is responsible for the ~150 year construction of beach ridges at Tolston Beach. Beach ridges are also constructed during a short term cycle of every 24 to 35 years with a magnitude of 0.5 – 0.6 meters (Thompson, 1992).

Following Thompson's (1992) study, investigations of beach ridges were expanded by Thompson and Baedke (1995) to the Thompson Embayment in the northern part of Lake Michigan. The lake-level curve constructed for this location also shows a general fall in lake levels (Thompson, 1992), but at a faster rate due to isostatic rebound (Thompson and Baedke, 1995). A notable dissimilarity in ridge construction was found from 2,800 – 1,700 cal. yr. B.P., when dune ridge construction at Thompson Embayment was reduced relative to Tolston Beach. Thompson and Baedke (1995) attributed this to a reduction in sediment supply from local headlands near the Thompson Embayment. Conversely, Tolston Beach had larger amounts of sand supplied via longshore and beach drift from the east and west coastal areas of Lake Michigan (Thompson, 1992).

Lichter (1995) continued this type of research on the beach ridges along the northern shores of Lake Michigan. He used radiocarbon dating of macrofossil remains of early-successional dune plants from swales in dune-ridge complexes at Wilderness State Park to date periods of dune construction. Using this method, Lichter (1997) reported nine radiocarbon dates from the toe slopes of beach ridges at Wilderness State Park. With these dates in hand he calculated the mean return period of high lake levels at 35.7 years, which is very similar to that reported by Thompson (1992).

Thompson and Baedke (1997) continued their previous work on lake-level fluctuations at three additional beach-ridge complexes along Lake Michigan. They used stratigraphic and chronologic evidence from the ridges to determine the elevation and time of ridge development. In a singular context, each complex yields only a partial chronology of lake-level fluctuations. Overlapping the lake-level curves this study provides a better record of late Holocene lake levels than previous studies (e.g., Thompson, 1992; Lichter, 1995; Lichter, 1997).

The lake-level curves from Thompson and Baedke (1997) were combined by Baedke and Thompson (2000) to create a unified lake-level curve for all of Lake Michigan that is also normalized by removing isostatic adjustments. The result was the most detailed record of late-Holocene changes in Lake Michigan to date (Figure 2:5). This curve does not include a record of ~ 150 and ~ 33 year fluctuations reported earlier (e.g.; Thompson, 1992), but these changes are assumed to have occurred.

These studies comprise the lake-level record to which coastal dune activity is compared. However, it should be noted that these reconstructions are based on calibrated radiocarbon ages from ridge-swale complexes. The calibration of the radiocarbon ages and their subsequent association with ridge construction may introduce additional error into the lake-level curves. These issues are addressed later in this chapter.

Coastal Dune Research

Early Research

Coastal dunes along Lake Michigan have been the focus of research for over 100 years. The first geologic/geomorphic investigation of the dunes was conducted by Cowles (1899), who studied the role of vegetation in the formation of Lake Michigan dunes. He proposed a model of foredune formation, which begins with *embryonic* or *stationary* beach dunes, which then become modified into *secondary embryonic* dunes. In this model, the beach functions as the sand source for foredunes as winds sweep across that part of the landscape. Once they begin to move, the coarsest sand grains are subsequently deposited as wind diminishes around pockets of vegetation, whereas finer particles are deposited on the lee side of the mound. This depositional pattern results in a dune cross-section with a steep backslope and gentle lee face.

Critical to Cowles' (1899) model is the ability of perennial plants (i.e. beach grasses) to grow *and* propagate throughout embryonic dunes. The most important plant species, according to Cowles, is *Ammophila arundinacea* (sand reed, syn. marram grass), which is a perennial grass that grows in response to sand burial. It also propagates radially via rhizomes, which in turn sprout upward-growing stems. The nature of marram grass is responsible for the lateral and vertical growth of foredunes.

Cowles (1899) also proposed that dune geomorphology is influenced by the type of vegetation present. Grasses and reeds are the most common vegetation on foredunes because they rapidly expand across dunes. Dunes anchored by grass and reeds are less than a meter in height and have very low slopes. Conversely, foredunes anchored by sand cherry and cottonwood are very tall and steep due to the upward growth ability of these species. In Cowles' (1899) model, vegetation that anchors foredunes will eventually propagate lakeward, causing the formation of a new embryonic dune that will reduce sand supply to inland foredunes. At this point the inland foredunes are modified by the introduction of slow-growing species, such as shrubs, thus creating secondary embryonic dunes.

Eventually the secondary embryonic dunes may become so tall that the anchoring plants are unable to reach the water table, or are exposed to the desiccating effects of wind exposure (Cowles, 1899). This development reduces vegetative cover and increases erosion, at which point the dunes may become *wandering* dunes. These dunes migrate with the prevailing wind, with sand being transported up a gentle backslope, over the crest, and deposited on the slip face. The wandering dune, according to Cowles (1899), is the building block for more complex dune landscapes, including blowouts and parabolic dunes.

Cowles (1899) concluded his model with *established* dunes, which are classified based on their dominant plant communities. These dunes are essentially wandering dunes which have been stabilized by dominant plant species, such as basswood, maple, oak, and evergreen. Generally the distribution of these species is related to the moisture, aspect, and latitude of sand dunes. The established dunes within a dune complex are subject to erosion, at which time they would again be wandering. Although Cowles (1899) largely focused on vegetation, his descriptions of foredune and dune complex geomorphology created the foundation for further research on Lake Michigan coastal dunes.

Following the Cowles, (1899) study, the next research on coastal dunes did not occur until Dow (1937) investigated dunes along the northeast coast of Lake Michigan.

He defined these dunes as "perched dunes", because they mantle tall coastal bluffs composed of glacial sediment (Figure 2:6). Dow's (1937) study focused on determining the source of eolian sand in these dunes. He hypothesized four possible scenarios: 1) sand of beach origin, transported directly upslope, 2) sand of beach origin that is transported upslope as lake erosion approaches the base of the bluff, 3) eolian sand derived from glacial sediments, or 4) a combination of the three processes. Based on observations of sand transportation at the perched dunes, he concluded that most of the sand is derived from the glacial sediments on which the dunes are perched, i.e. local reworking of deposits by wind. This is especially true where coastal bluffs are quite tall (~100 m), although certainly some proportion of beach sand is transported to the top of the bluff.

In the same vein as Cowles (1889), Tague (1946) conducted a qualitative description of dunes in the Grand Marais Embayment at Warren Dunes along Lake Michigan's southeastern shore. He assumed that foredunes are the primary dune form and are supplied with sand from beaches. Foredunes may then erode, and eventually morph into large secondary coastal dunes. The terminology and process of Tague's (1946) conclusions were largely modeled after Cowles (1899).

Unlike Cowles (1899), Tague (1946) was interested in the history of coastal dune evolution. He sought to use sand dune geomorphology and shape to infer a chronology of dune evolution, which he admitted was a relative dating method. Dunes closest to the lakeshore, like foredunes, were assumed to be younger than those farther inland. Based

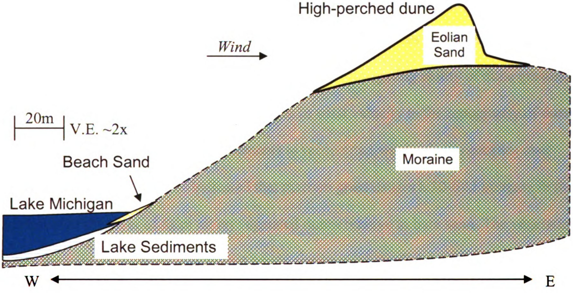

Figure 2:6 – Cross section of a typical high-perched dune. This dune is perched upon a moraine, with erosion occurring on the windward face of the dune and moraine bluff. Mobilized sand moves up the slope and is deposited on the slip face. (Adapted from Anderton and Loope, 1995)

on the position of dunes in relation to former lake terraces, he proposed that the largest dunes at Grand Marais were constructed during the Nipissing high stand.

One of the most influential dune studies on Lake Michigan was done by Olson (1958a, b, and c) who, like Cowles (1899), also studied the role vegetation plays in forming foredunes. Olson (1958b) used plant structure and distribution to reconstruct changes in dune morphology such as sand deposition, erosion, and stabilization. By excavating foredunes and studying annual growing patterns of beach grasses, he was able to estimate annual rates of deposition on foredunes. Like Cowles (1899), Olson (1958b) associated foredune formation to the stabilizing and propagating ability of dune grasses.

Olson (1958a) also attributed changes in wind velocity to the topography of dune ridges. He used anemometer measurements to reconstruct wind velocity profiles over vegetated foredunes. Results indicated that marram grass causes a 30-fold increase in the thickness of the layer of calm air on vegetated dunes relative to bare sand. The vegetation alters the wind velocity profile so that winds approaching the foredune become compressed and accelerates. This effect increases erosion on the backslope, while a sharp decrease in wind velocity behind the vegetation causes sand deposition onto the slip face.

Whereas Cowles (1899) focused on vegetation as the impetus for foredune formation, Olson (1958c) proposed an alternative model for the evolution of Lake Michigan foredunes. He suggested that foredune formation begins with submerged offshore bars. When lake levels decrease these bars become subaerially exposed and may separate from the lake to form foredunes (Figure 2:7). Pioneer vegetation then begins to stabilize the foredunes which, along with topography (e.g., Olson, 1958b) enhances

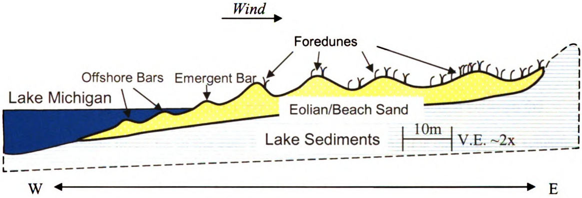

Figure 2:7 – Cross section of a typical foredune complex. The foredunes in this example have formed due to dropping lake levels, with the most eolian activity occurring at the dunes nearest the lake. As lake levels drop, foredunes farther back in the complex become stabilized by vegetation. (Adapted from Olson, 1958c).

foredune enlargement. Continued lowering of the lake would expose more offshore bars, and create a ridge-swale complex.

Modern Dune Research

Research on Lake Michigan's coastal dunes waned after the 1950's. In 1970, Dorr and Eschman published *Geology of Michigan* and loosely applied Olson's (1958c) model to large parabolic coastal dunes along Lake Michigan. They suggested that the large coastal dunes are relicts of lowering lake levels following the Algonquin and Nipissing high stands, and that they have fundamentally been stable since that time. Their conclusions were essentially mirrored by Buckler (1979), who introduced the term *barrier dunes* for the large dunes that provide a topographic obstruction between the beach and inland locations to the east. In addition to providing this classification scheme, he stated that the most extensive period of dune formation occurred during the waning phases of the Nipissing high stand when sediment supply was increased due to wide beaches (e.g., Tague, 1946).

Following Buckler's (1979) study, the next research focusing on Lake Michigan coastal dunes was conducted by Snyder (1985), who investigated the perched dunes at Sleeping Bear Dunes National Lakeshore. These dunes mantle a tall (~ 100 m) coastal bluff composed of glacial sediment in the Manistee Moraine. Snyder's study was possible because erosion of the bluff had revealed several buried soils which he dated to reconstructed the radiocarbon (^{14}C) chronology of eolian activity.

The oldest radiocarbon date Snyder (1985) acquired was from a paleosol formed in the uppermost part of the glacial sediment. This soil provided a date of 4,559 ± 226 cal. yr. B.P. (4,065 ± 160 B.P.). Another radiocarbon date of 2,781 ± 161 cal. yr. B.P. (2,675 ± 145 B.P.) was acquired from a paleosol formed in a unit of eolian sand above the glacial sediment. A third date from a soil stacked higher in the eolian sequence dated to 688 ± 178 cal. yr. B.P. (730 ± 250 B.P.; Crane, 1956). Snyder (1985) interpreted the radiocarbon dates to mean that the first period of dune construction began after the Nipissing high stand, approximately 3,170 cal. yr. B.P.

Snyder's (1985) interpretation of the chronology at Sleeping Bear may be erroneous, as radiocarbon dates only indicate a maximum limiting date of soil burial. That is, the radiocarbon dates coincide to oldest possible time of burial, and could be older, because organic material accumulates over time before it is buried. The paleosol formed at the contact between glacial sediment and eolian sand must have formed prior to burial, probably during a period of stability which coincided to the Chippewa low phase. This interpretation is supported by a radiocarbon age of ~ 4,559 cal. yr. B.P., and suggests that the paleosol was buried due to the onset of eolian activity during the Nipissing transgression. Formed in the unit of eolian sand above the glacial sediment was a paleosol which was buried about 2,800 years ago, during the Algoma high stand (Baedke and Thompson, 2000). This likely indicates a second period of dune construction, rather than the minimum limiting age of eolian activity as Snyder (1985) had suggested.

Following the Snyder (1985) study, research on coastal dunes in Michigan surged, beginning in the 1990s. These investigations systematically focused on reconstructing the geomorphic evolution of large dunes by analyzing dune stratigraphy in conjunction with radiometric dating. This new wave of dune studies began when Anderton and Loope

(1995) studied the Grand Sable dunes on the south shore of Lake Superior in Michigan's Upper Peninsula (Figure 1:1). Grand Sable is a classic perched dune field in the Dow (1937) model, in that it mantles high bluffs composed of glacial sediment. The dunes contain a sequence of buried soils and eolian sand units, which suggest periodic activation. Anderton and Loope (1995) used radiocarbon dating of the buried soils to infer the time of burial by eolian sand, events which they then correlated to lake levels (e.g., Larsen, 1985)

Anderton and Loope (1995) found that the basal and best developed soil in the Grand Sable dune field formed in the uppermost part of glaciofluvial sediment and dated to 5,925 – 5,339 cal. yr. B.P. (5,170 – 4,640 B.P.). This basal soil has O/A/E/Bs1/Bs2/C horizonation, and was identified as the Sable Creek Soil, first recognized by Farrell and Hughes (1985). The stratigraphic location and development of the Sable Creek Soil suggests that it formed over a relatively long period of time, probably during a period of stability which coincided with the Houghton low phase of Lake Superior. Above the Sable Creek Soil are several paleosols formed in eolian sand, which were buried during periods of dune activity, most notably at about 3,776 and 526 years ago. The basal soil and overlying paleosols formed in eolian sediment are chronologically similar to those found by Snyder (1985) at Sleeping Bear Dunes.

Anderton and Loope (1995) correlated these periods of activity to late Holocene fluctuations in Lake Superior (e.g., Colman et al., 1994). They hypothesized that when lake levels increase, waves undermine the coastal bluff. The bluff face then erodes to form a barren colluvial slope from which sand is deflated and transported to the perched dunes. This model incorporates Dow's (1937) hypothesis regarding the source of perched dune sand with the destabilizing effect of high lake levels, and synthesizes them as the *perched-dune model* (Figure 2:8).

Following Anderton and Loope's (1995) study, Arbogast and Loope (1999) investigated a variety of high-relief dunes mantling low lacustrine terraces along the eastern shore of Lake Michigan. They tested the prevailing hypothesis that these dunes formed during the Nipissing high stand (Dorr and Eschman, 1970; Buckler, 1979). This was accomplished through stratigraphic analysis at four sites between Manistee and Grand Haven, in association with radiocarbon dating of buried soils (e.g., Anderton and Loope, 1995; Snyder, 1985).

Arbogast and Loope (1999) focused on basal paleosols that are formed in the uppermost part of the underlying lake sediments. These soils are Entisols trending toward Spodosols. Radiocarbon dating of these soils indicated that the time of burial differed between sites. The oldest age was collected from a thin eolian sand unit overlying lacustrine sediment near Muskegon in western Lower Michigan. The date indicates that this soil was buried at ~ 4,600 cal. yr. B.P. Dunes at the Jackson and Nugent quarries became active at roughly 4,100 cal. yr. B.P., when dune sand buried soils formed in lacustrine sediment. In contrast to these ages, the dune at the Rosy Mound quarry was the last to activate at ~ 3,100 cal. yr. B.P. Overlying this soil are five units of eolian sand, each separated by weakly developed soils. In contrast, overlying the basal soil at the other three sites is one unit of eolian sand with Spodosols formed at the modern surface.

Although soil burial was not always contemporaneous between all of the sites, each burial event correlated to high lake levels. Arbogast and Loope (1999) associated the onset of dune construction with the Nipissing high stand at three sites along the Lake

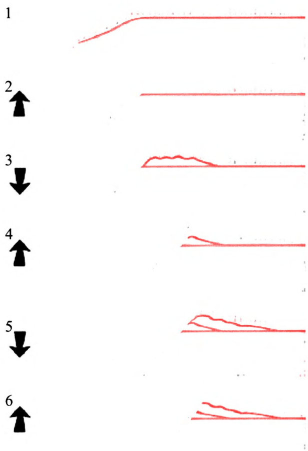

Figure 2:8 – The perched-dune model.

1 – A moraine is left after the last retreat of the Laurentide ice sheet. Vegetation grows and soils form. A low proglacial or post-glacial lake is present.

2 – Lake levels rise, waves cut into the base of the bluff. Some sand is mobilized and buries the soil on top of the bluff (pink). Other materials are deposited as lake sediment, or winnows into the lake.

3 – Lake levels drop, the bluff face stabilizes. Soils form. Exposed lake sediments may get worked into foredunes (pink).

4 – Lake levels rise, flooding foredunes and again cutting into the bluff. Sand mobilizes and buries soils on top of the bluff.

5 – Lake levels drop, the bluff face stabilizes. Soils form. Exposed lake sediments may get worked into foredunes (pink).

6 – Again, lake levels rise, flooding foredunes and again cutting into the bluff. Sand mobilizes and buries soils on top of the bluff. (Modified from Loope and Arbogast, 2000)

Michigan shoreline in western Lower Michigan. The dune at Rosy Mound, however, apparently activated during the Algoma phase. Their results indicated that these coastal sand dunes were actually formed during or after the Nipissing high stand and may have been active during more recent high lake levels (e.g. post-Algoma). This conclusion is in disagreement with the previous assumption that the dunes actually formed during waning phases of the Nipissing (Dorr and Eschman, 1970; Buckler, 1979).

At about the same time that Arbogast and Loope (1999) determined that large dunes on lake terraces along Lake Michigan began growing during the Nipissing high stand, a study by Arbogast (2000) estimated the time of stabilization of a perched dune field on the southern shore of Lake Superior at Nodaway Point. Arbogast (2000) hypothesized that the dunes at Nodaway Point were last active during the Nipissing high stand, and stabilized thereafter. To test this hypothesis, Arbogast (2000) used a combination of soils analysis and radiometric dating (^{14}C and OSL) to determine the ages of dune formation. A radiocarbon date from a basal paleosol formed in glacial sediment indicated that burial occurred approximately 3,480 – 3,160 cal. yr. B.P., while an OSL sample from the upper portion of the overlying eolian unit dated to 3.38 ± 0.16 ka*. These dates are in excellent agreement, and indicate that the dunes first formed about 3,500 years ago, and began to stabilize about 3,300 years ago. However, these ages seem to indicate that dune activity lagged behind the Nipissing high stand. This delay may be due to post-Nipissing coastal erosion or slow stabilization of the dune field.

As the coastal dune research in Michigan accelerated, Loope and Arbogast (2000) conducted a study that more rigorously tested the hypothesis that sand is mostly supplied to large dunes on topographically low lake terraces during periods of high lake levels.

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* OSL ages in this chapter are provided in ka, or kiloannum (1,000 years) before present.

Their study examined a variety of dunes along the shoreline of northwest Lower Michigan that mantle moraines, outwash plains, and lake plains. Loope and Arbogast (2000) collected radiocarbon samples from dune paleosols to infer the time of burial by eolian sand. Their results indicated that the onset of dune construction, and subsequent periods of activity, were usually synchronous to high lake phases. They concluded that most dune construction occurred after the Nipissing high stand, during ~150-year high stands of Lake Michigan (e.g., Thompson, 1992; Thompson and Baedke, 2000).

Arbogast et al. (2002a) continued research on massive coastal dunes along Lake Michigan by dating a sequence of buried soils discovered in several lake-facing exposures near Holland, Michigan. The basal paleosol here is formed in the uppermost lake sediments and was buried at about 5,500 cal. yr. B.P. Above the basal paleosols was a sequence of buried Entisols indicating episodic dune construction punctuated by stable periods during which soils formed. According to Arbogast et al (2002a), about 75% of dune building occurred between ~ 4,000 and 2,500 cal. yr. B.P. The dunes subsequently stabilized for a relatively long period of time, resulting in the development of an Inceptisol with A/E/Bs/C horizonation. This soil was subsequently buried ~ 500 years ago. In the past ~ 300 years the dunes have grown episodically with at least two brief periods of stability occurring.

The stratigraphy and chronology of the dunes at Holland (Arbogast et al., 2002a) is similar to other locations along the Lake Michigan shoreline. Van Oort et al. (2001) expanded on the research of Arbogast et al. (2002a) by dating paleosols cropping out of dunes at Van Buren State Park, Michigan. The basal paleosol here is a horizontal peat layer that caps lacustrine sands. Several radiocarbon samples were acquired from the peat layer across the study area and indicated that the peat formed between ~ 6,170 to 4,880 cal. yr. B.P. Dates from the upper portion of the peat suggest that it was buried by eolian sand during the Nipissing high stand.

As at Holland (Arbogast et al., 2002a), the lower part of the eolian sands contains a sequence of Entisols that represent brief periods of landscape stability and soil formation during the evolution of the dune (Van Oort et al., 2001). These soils provided dates ranging from 3,690 – 1,970 cal. yr. B.P. In the uppermost eolian deposits a relatively well developed buried soil was discovered. Like at Holland, this soil is an Inceptisol with Spodic characteristics that represents a relatively long period of landscape stability and soil formation. Charcoal from the A horizon of this Inceptisol dates to ~ 500 – 100 cal. yr. B.P., which suggests that the soil developed for ~1,500 years before it was buried.

In association with the research conducted at Holland (Arbogast et al., 2002a) and Van Buren State Park in Michigan (Van Oort et al., 2001), it was noted that a series of small dunes occur in the lee of the massive parabolic dunes. These *backdunes* were believed by Tague (1946) to be older than the larger dunes and that they probably formed during earlier high lake phases (e.g., Calumet, ~13,600 to 13,100 cal. yr. B.P., Hansel et. al., 1985). A study by Hansen et al. (2002) tested Tague's (1946) chronology by obtaining OSL ages on the sand deposits contained within the dunes. Results from Hansen et al. (2002) indicate that the backdunes in the Marias Embayment near Holland are much younger than Tague (1946) thought. A basal sample provided an age of 4.3 ka, whereas those taken from the crests of the dunes range in age from 4.0 – 3.7 ka. These dates are contemporaneous with the lowermost deposits in the lakeward parabolic dunes which were studied by Arbogast et al. (2002a) and suggest that the entire dune system was active during that period of time. The onset of backdune growth may have begun during the Nipissing high stand, with final stabilization between 4 and 3.5 ka. This stabilization may have been caused by a cooling after the mid-Holocene Altithermal (hypsithermal), and/or increased shelter from onshore winds caused by the growth of massive parabolic dunes (Hansen et al., 2002).

Hansen et al. (2004) described the history of massive parabolic dunes at Green Mountain Beach, Michigan, based on radiocarbon dates from buried soils. The basal paleosol here dates to 5,890 – 5,310 cal. yr. B.P., which again correlates with the Nipissing high stand. Like at Van Buren State Park (Van Oort et al., 2001), a sequence of sand deposits and associated Entisols overlie the basal soil and provided dates from 4,900 to 3,800 cal. yr. B.P. This sequence is capped with an Inceptisol which dates to about 1,350 cal. yr. B.P. Above that are several Entisols which date to 500 to 300 cal. yr. B.P., and suggest periods of remobilization after 1,100 cal. yr. B.P.

By 2004, several studies (e.g., Van Oort et al., 2001; Arbogast et al., 2002a; Hansen et al., 2004) had reported the presence of numerous buried soils in coastal dunes on Lake Michigan's southeastern shore. The most prominent of these is a soil with spodic characteristics and A/E/Bs/BC/C horizonation that is usually found in the upper third of the dune deposits. This soil has been identified along a ~ 200 km stretch of Lake Michigan that extends from Indiana Dunes National Lakeshore in Indiana to Montague, Michigan and suggests a period of regional stability. Given the development and widespread nature of this soil it was informally named the Holland Paleosol by Arbogast et al. (2004).

The age of the Holland Paleosol was inferred from a variety of data, including the character of the soil and information from Arbogast et al. (2002a) and Hansen et al. (2004). This integration indicated to Arbogast et al. (2004) that the Holland Paleosol formed in eolian sand that was deposited between 3,000 and 2,500 cal. yr. B.P., and was subsequently buried about 1,000 – 900 cal. yr. B.P. at the Holland, Michigan type section and at about 400 cal. yr. B.P. at Van Buren State Park. As in other studies (Hansen et al., 2004; Van Oort et al., 2001), the deposits overlying the paleosol contain several Entisols which date to less than ~ 300 cal. yr. B.P., depending on location. Arbogast et al. (2004) were uncertain as to the environmental cause for the stable interval in which the Holland Paleosol formed and tentatively linked it to the possibility that the active coastline may have been more westward at that time.

Most recent dune studies have largely focused on determining the chronology of the formation of massive coastal dunes, however some other research has focused on the specific geomorphic processes that influence foredunes. Van Dijk (2004) focused on contemporary Lake Michigan dune processes, specifically small-scale volumetric changes associated with foredunes at P.J. Hoffmaster State Park in Michigan. She used erosion pins, sand traps, and microclimate instruments to quantify changes in topography on a seasonal basis, and found that the foredunes had increased in height and volume over the three year study, and that sand mobility was highly variable from year to year. Moreover, most of the sand deflated from the beach was trapped in the foredune ridge and was not transported farther inland. The patterns of sand transportation were correlated to seasonal changes. During the summer there was little geomorphic change, as beach grasses sufficiently reduced wind velocity in front of the foredune. In the autumn, intense weather systems and reduced vegetation led to an increase in sand mobilization. Winter presented the harshest conditions, when strong winds could potentially move sand over mostly unvegetation foredunes. However, the dunes were sometimes protected by snow cover and wet or frozen surfaces. Coastal ice also reduced shoreline sediment supply and onshore winds, which in turn eliminated wave erosion. During the spring, a shift in wind speed and direction reduced the amount of sand transported onto the foredune. Foredunes then became more stabilized as vegetation once again anchored the foredunes.

Although most dune research has focused on the southeastern part of the Lake Michigan coast, recent detailed investigations have moved to the northeast portion of the coast. In this context, Lepczyk and Arbogast (2005) studied the geomorphology of coastal dunes at Petoskey State Park in northwest Lower Michigan. This study identified five geomorphic units, including 1) lake terrace, 2) parabolic dunes, 3) onlap dunes, 4) linear/shadow dunes and 5) active dunes. Interpretation of these units, in association with radiocarbon and OSL dating, was used to reconstruct the geomorphic history of the study area.

The oldest landform recognized at Petoskey State Park is a lake terrace that was identified in the area east of the dunes. According to Lepczyk and Arbogast (2005), the elevation of the terrace indicates that it formed during the Nipissing I high stand of Lake Michigan. This surface is buried by the parabolic dunes immediately to the west, which began to form 4,815 – 4,425 cal. yr. B.P. when eolian sand buried a well developed Spodosol formed in the uppermost part of the underlying lake sediments. OSL ages from the crest of the parabolic dunes indicate of the dunes stabilized at ~ 2.2 ka in the southern part of the dune field and ~ 1 ka in the northern area.

As the name implies, the onlap dunes mantle the western slope of the parabolic dunes. These dunes are smaller than the parabolic dunes, suggesting that they took less time to form (Lepczyk and Arbogast, 2005). The onlap dunes probably began forming around 2,800 cal. yr. B.P., when they buried a paleosol formed in a thin unit of eolian sand capping lake sediment. The onlap dunes apparently continued to bury this paleosol as they migrated inland, as indicated by a radiocarbon date of ~ 1,700 cal. yr. B.P. obtained on a paleosol buried beneath the crest of the these dunes.

The linear and shadow dunes lie immediately to the west of the onlap dunes. The age of these dunes is uncertain because they do not contain any buried soils that could be dated. They must be younger than ~ 3.2 ka, however, because they mantle the Algoma lake terrace and are lakeward of the onlap dunes. Dates from the active dunes indicate that they began to form about 1,000 years ago, and underwent a period of stability during which a paleosol formed. This soil was buried between 500 and 300 cal. yr. B.P. A younger soil above that was buried at about 300 cal. yr. B.P. Results from this study indicate that the dunes at Petoskey State Park began forming during the Nipissing high stand and were active at later times, perhaps in response to fluctuating lake levels.

As can be seen from this review, a considerable body of research has focused on reconstructing the chronology of dune construction along Lake Michigan's eastern shore (e.g., Arbogast and Loope, 1999; Loope and Arbogast, 2000; Arbogast et al., 2002; Van Oort et al., 2001; Hansen et al., 2002; Arbogast et al., 2004; Lepczyk and Arbogast, 2005) and Lake Superior's southern shore (e.g. Anderton and Loope, 1995; Arbogast, 2000). Results from these studies suggest that topographically low dunes that mantle lacustrine sediments respond to fluctuating lake levels in much the same manner as those perched upon tall glacial bluffs (e.g. Dow, 1937; Snyder, 1985; Anderton and Loope, 1995). As a result, Arbogast (2009) proposed the classification of coastal dunes based on topographic position. In this scheme, dunes perched high above lake level are called *high-perched dunes* (Figure 2:6), whereas those mantling topographically-low lake terraces are called *low-perched dunes* (Figure 2:9). This classification system links high and low-perched dunes together on the basis that the fundamental reasons for their formation are the same (Figure 2:10).

International Dune Studies

Although a large literature base for Lake Michigan coastal dunes exists, research on dunes elsewhere in the world provides a broader context for dune studies in the Great Lakes region. Despite the difference in processes which control sea levels compared to Lake Michigan, the effect of fluctuating sea levels on coastal dunes is generally similar. In that context, this section summarizes several coastal dune studies from abroad which focus on coastal dune processes. Also, dating techniques pioneered by international researchers and critical to modern dune studies will be discussed.

A large body of research on coastal dune processes has been completed in Australia, specifically along the southeastern coast of that country. Hesp (1988) defined a quantitative method to categorize foredunes based on geomorphology and vegetative cover. His classification deals with derivations of established foredunes, which are laterally continuous along the shoreline. These 'classic' foredunes are stage one of his

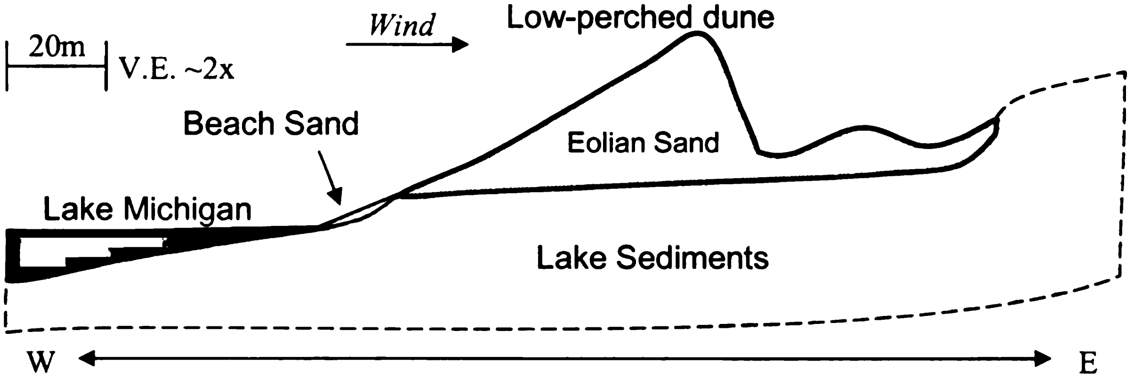

Figure 2:9 – Cross section of a typical low-perched dune. The dune is perched upon lake sediments, with erosion occurring on the windward face. Deposition occurs on the slip face. The landscape behind the dune is relatively stable, as it is protected from the wind. (Adapted from Loope and Arbogast, 2000).

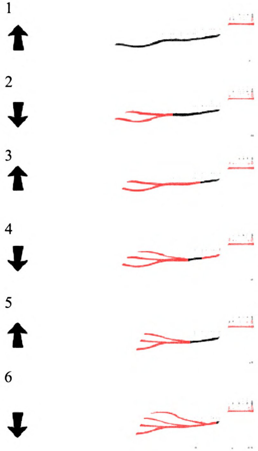

Figure 2:10 – The perched-dune model, as applied to low-perched dunes.

1 – High lake levels cut into truncated lake sediments (green), mobilizing sand (pink). Soils become buried.

2 – Lake levels drop, the dune stabilizes. Soils form. Exposed lake sediments may get worked into foredunes (pink).

3 – Lake levels rise, the dune destabilizes, which mobilizes sand. Soils become buried.

4 – Lake levels drop, the dune stabilizes. Soils form. Exposed lake sediments may get worked into foredunes.

5 – Lake levels rise, the dune destabilizes, which mobilizes sand. Soils become buried.

6 – Lake levels drop, the dune stabilizes. Soils form. Exposed lake sediments may get worked into foredunes. (Modified from Loope and Arbogast, 2000)

five categories, and are 90% to 100% covered with vegetation. Stage two foredunes have developed unvegetated troughs, with vegetation cover ranging from 75% to 90%. Stage three foredunes contain blowouts and have hummocky topography, with 45% to 75% vegetative cover. Stage four foredunes have been considerable eroded, and are only 20% – 45% covered with vegetation. Finally, stage five foredunes are only remnant knobs separated by blowouts, with 5% – 20% vegetation.

More recently, Hesp (2002) revisited the subject of foredunes by focusing on their formation and geomorphology. He indicated that foredunes fall into two main types: *incipient* and *established* (e.g., Hesp, 1988). Incipient (sometimes referred to as embryonic) foredunes are those developing nearest the water due to a reduction in wind velocity caused by vegetation. Incipient foredunes grow laterally and vertically due to expansion of vegetation via rhizomes. They are accretionary features which eventually stabilize due to dense vegetation and a decrease in sand supply.

Established dunes develop from incipient dunes, and are differentiated by their greater morphological complexity, height, width, age, and geographic position. The development of established foredunes depends on numerous factors, including sand supply, vegetative cover, storm and wind stress, wave stress, and anthropogenic forces. One of the most important of these factors is wave stress, which can increase or decrease depending on sea levels. As sea levels fall, incipient foredunes form near the water as the beach progrades. This increases the supply of beach sand, which may allow the formation of new incipient foredune towards the sea, and begin to create a foredune plain. If vegetation is unable to rapidly move into the new beach area, an increase in sediment supply may effectively bury the nearest foredune vegetation, thus causing instability.

When sea levels rise, waves will presumably erode the backslope, causing a blowout and the landward retreat of the foredune.

Hesp (2002) related the varied outcome of sea level changes to his 1988 model of established foredunes, and incorporated scenarios for accretion and erosion cycles for the southeastern coast of Australia. In this model, established foredunes in a stable or prograding environment will trend towards stage one foredunes, which are heavily vegetated. This may be counteracted by an erosive environment, or by rapid wave erosion events. In this scenario, the foredunes would erode to stage five, move landward, or be eroded by blowouts.

Hesp (2002) also discussed the morphology and development of blowouts. Blowouts have two general types: saucer and trough. They are generally initiated by a decrease in vegetation, either by erosion or climate reasons. Wind is then focused into the area of decreased vegetation, resulting in the mobilization of sand. *Trough* blowouts tend to form where topography is steep, and wind is funneled down the narrow corridor of the blowout. Sand is deflated from this basin and transported inland, causing a linear extension of the blowout. Over time, or where topography is lower, blowouts tend to be wider. These larger depressions are called *saucer* blowouts, where erosion is reduced and sand transport occurs in multiple directions. Often, the complex flow of wind over saucer blowouts can cause expansion in all directions. Both saucer and trough blowouts, given time, may evolve into parabolic dunes.

Although Australian dune studies have focused on the processes that form coastal dunes, studies in other parts of the world focus on the physics of dune building. In Europe, research by Arens et al. (2002) tested the prevailing theories on changes in grain-size distribution during transport. This was achieved using vertical saltation traps which collect sand being transported in the wind. These traps collected sand in five cm increments and provided data which was used to create sediment profiles at each of the locations.

Their results showed that mean grain size either decreases or is not related to height. Sorting on the foredune is poorer than on interior dunes, probably because of a progressive loss of coarse grains in the direction of sand transport. As sand is transported inland, wind velocity decreases, thus making it less likely that large grains will be transported here.

Anderson and Walker (2006) expanded on previous studies (e.g., Arens et al., 2002) of airflow and sand transport in British Columbia, Canada. Their study examined the influence of vegetation and topography on sand transportation in backshore foredune/parabolic dune complexes on a temporal scale of hours and years. This analysis was done by measuring wind velocities using anemometers and measuring sediment transport using grain impact sensors and traps. These data was incorporated with measurements of vegetation density and topography to create a model of wind and sediment transport from near the shoreline to the backdune complex.

Anderson and Walker (2006) found that the topography of foredunes at their study area greatly influences air flow. Air is compressed and accelerated over incipient foredunes, and in this study, the wind was then channeled into a trough blowout, where it accelerated as much as 1.8 times. Regardless of the incident wind direction, the blowout acts as a conduit for sediment and wind flow towards the backdune complex. When air exits the blowout, wind flow expands and decelerates to ~ 60% of the original wind velocity. Because of this, much of the sediment eroded from the blowout is deposited on the lee side in a depositional lobe.

Anderson and Walker (2006) concluded that the parabolic dune in the backshore was supplied with sand from the foredune complex and from long distance sand suspension due to intense storms. Sand supply via grainfall may be responsible for maintaining active dunes far inland from the shoreline in a mostly vegetated foredune plain.

While many international dune studies focus on the physics and geomorphology of sand dunes, a considerable amount of effort has been devoted to understanding the chronology of dune construction. This is particularly true on the North Atlantic coast of Europe, where the possibility of rising sea levels has lead researchers to reconstruct the historical effect of such changes. Work done using a combination of radiocarbon and luminescence dating methods (e.g., Murray and Clemmensen, 2001; Bailey et al., 2001) has revealed regional trends in eolian activity across Europe, and provide excellent context for dating methods used in the Great Lakes region.

Many studies in Europe have used radiocarbon dating of paleosols to construct a history of dune activity. For example, Wilson and Braley (1997) dated organic material (soil organic matter, peat, terrestrial or marine shells) from paleosols in seaward blowouts and gullies on coastal dunes in Northern Ireland. They found that the dunes were first active approximately 7,000 cal. yr. B.P., and were subsequently active during transgressive sea level phases. However, they also found periods of eolian activity from 3,100 to 2,400 and 650 to 50 cal. yr. B.P., which they attribute to climate changes or site-specific factors. Other studies across the region have found similar periods of dune activity related to climate and local factors rather than sea level changes (Sommerville et al., 2007; Bailey et al., 2001; Murray and Clemmensen, 2001).

A study using similar methods was completed in 2000 by Haslett et al. in Brittany, France, where Holocene dunes are perched on Pleistocene deposits. They collected radiocarbon samples from buried soils, and associated the onset of dune construction with the date of burial of the basal paleosol. Their date indicates that initial periods of dune building began as sea levels rose, approximately 4,300 cal. yr. B.P., and subsequently stabilized along with sea levels, approximately 3,250 cal. yr. B.P.

The relationship between sea levels and dune activity led Haslett et al. (2000) to propose the *perched dune development model*, which states that increasing sea level increases beach sand supply. This transgression is followed by stabilizing sea levels, which in turn allows dunes to become stabilized by vegetation. Finally, the high sea levels erode into the underlying sediment of the perched dunes, resulting in a landward retreat of these dunes which are often backed against bedrock cliffs. Although similar to the Great Lakes *perched-dune model* (Dow, 1937; Anderton and Loope, 1995), the presence of bedrock cliffs results in a different morphology in this particular dune complex. Despite the differences in the two models, the relationship between water levels and dune activity has been identified internationally.

Radiometric Dating in Eolian Environments

Most of the aforementioned studies rely on radiocarbon (^{14}C) dating of organic materials in buried soils to interpret eolian depositional chronologies. However, uncertainties associated with using radiocarbon dating in eolian environments have lead to an increasing trend in the use of OSL dating (e.g., Bailey et al., 2001; Roberts and Plater, 2007). This section discusses the use of radiocarbon and OSL dating in eolian environments and the rationale for the sole use of OSL in this investigation.

Radiocarbon dating uses the ratio of ^{14}C (radiocarbon) to ^{12}C and ^{13}C (stable carbon) to estimate the age of organic materials. The theory is that all living organisms remain in equilibrium with atmospheric ratios of stable and radiocarbon. Upon death of the organism, the radiocarbon isotope continues to undergo ionizing radiation and decay to ^{14}N, thus reducing the ratio of radiocarbon to stable carbon. By comparing this ratio to the half-life of radiocarbon (5568 years), a radiocarbon age can be derived (Stuiver and Polach, 1977).

Though the rate of radioactive decay of ^{14}C remains constant, they amount of radiocarbon in the atmosphere has fluctuated over time, and thus causes error in radiocarbon ages. Fortunately, past ratios of atmospheric carbon has been estimated through the analysis of tree-rings, corals, and other carbonaceous materials (Fairbanks et al., 2005). This allows radiocarbon ages to be calibrated from radiocarbon ages into calendar ages (cal. yr. B.P.) with great accuracy (Fairbanks et al., 2005).

Another possible source of error in radiocarbon dating is the reservoir effect, which is caused by the contamination (or incorporation) of "old" carbon with a very low ratio of radiocarbon to stable carbon into organic materials. This most commonly occurs in areas with high water tables and a large amount of "old" carbon, such as carbonate rocks (Schaetzl and Anderson, 2005). The reservoir effect can be corrected for to provide a 'reservoir corrected age' (Stuiver and Polach, 1977). Also, field sampling methods can help reduce the chance of contamination, in particular the collection of isolated wood and charcoal fragments from paleosols undisturbed by root activity (Schaetzl and Anderson, 2005).

Previous dune studies (e.g., Anderton and Loope, 1995; Loope and Arbogast, 2000) relied on radiocarbon dates of carbonaceous materials in paleosols to infer the time of burial by eolian sand. Much of the critique of radiocarbon analysis focuses around the relationship between soil formation and eolian activity, rather than problems with the method itself (e.g., Bailey et al., 2001; Roberts and Plater, 2007). Researchers believe that age errors exist in their samples due to the time-transgressive nature of dune stabilization and soil development (Wilson and Braley, 1997), which creates uncertainty as to the association of radiocarbon dates and soil stability. As such, radiocarbon dating only provides a maximum limiting age of burial, i.e. the eolian sand can be no older than the paleosol on which it is deposited. Also, organic matter in the A horizon may be much older than the time of burial, become mixed, or contaminated due to the translocation of humates downward into the paleosol, thereby providing erroneously younger ages for this buried soil.

Optical stimulated luminescence (OSL) dating reduces the uncertainties in correlating dune activity with soil development, as it directly dates the time of burial of eolian sand. As a result, OSL should better estimate periods of eolian activity, thus making correlations with lake-level curves potentially more accurate. The basic concept behind OSL dating is fairly straightforward. The method is based on the fact that common geologic minerals, such as quartz and feldspar, have imperfections in their crystalline structure. These imperfections trap electrons released by surrounding sediment due to ionizing radiation (e.g. alpha, beta and gamma radiation); thus they are referred to as ‘electron traps’ (Duller, 2004). The number of electrons trapped in the grains increases relative to the rate of irradiation and the elapsed time. This phenomenon allows quartz and feldspars to be effective dosimeters (Duller, 2004).

When exposed to light the host minerals luminesce, or release their electrons and reset the traps. Under laboratory conditions the number of electrons released can be measured (equivalent dose), and by comparing that to the rate at which the minerals are being irradiated (dose rate) the last date of light exposure can be calculated with the following equation:

$$age(yr) = \frac{D_e\,(Gy)}{D\ (Gy/yr)}$$

Where D_e = Equivalent dose in grays (1 Gy = 1 J/kg)

D = Dose rate in grays per year

OSL also has drawbacks that must be addressed. Although feldspar does accumulate a luminescence signal, a phenomenon known as ‘anomalous fading’ has occasionally caused dates to be underestimated, as the equivalent dose seems to stabilize for hundreds or thousands of years. Several methods were proposed to correct for the problem, including the sole use of quartz as the dosimeter, since it does not seem to have an anomalous fading problem (Duller, 2004). Thus, there has been a general shift towards quartz as the dosimeter of choice.

Another issue with OSL dating is that of the inconsistent reset of electron traps of sand grains during transport and deposition, or conversely contamination with recently reset grains after burial or during collection. This could result in erroneously old or young dates, respectively. Fortunately, sand deposited into dunes should always occur subaerially (in the presence of light), thus resetting the traps. Also, great care should be taken to collect samples from stable areas in an opaque enclosure to prevent accidental reset of the electron traps.

OSL dating has advanced greatly in the last few decades, in part to the development of the regenerative dose procedure, which was summarized by Duller (2004). In this general methodology, one or more small samples (aliquots) are stimulated using a source light, where their D_e is measured. Then, the samples are subjected to a laboratory source of radiation and measured again for their luminescence signal. Their efficacy in retaining the known lab radiation is then used to estimate the sample's effectiveness at retaining radiation. This methodology has arisen for two main reasons, one: the relationship between D and D_e is not absolute, and requires calibration of laboratory equipment. Secondly, the amount of radiation that can be stored in a quartz grain is limited. As D_e increases, the electron traps 'fill up', this making it less likely that emitted electrons will become trapped, this asymptotic relationship must be accounted for. The older method of multiple aliquot regenerative dose (MAR) uses dozens of aliquots from the main sample are given different radiation doses, the results of which are combined to estimate a 'growth curve', or luminescence response to radiation. Problems arose due to the different number of mineral grains per aliquot, and the time consuming nature of preparing dozens of samples.

The newer approach is single aliquot regenerative dose (SAR) developed largely by European researchers (e.g., Murray and Wintle, 2003). In this method, one aliquot is repeatedly subjected to laboratory radiation, and the ‘growth curve’ is adjusted accordingly after each run. This reduces the amount of analytical scatter, as uncertainty is now measurable thanks to the reproducibility of the method. Unlike older methods and technologies used for OSL, which were often expensive and unreliable (Duller, 2004), SAR allows for fast and precise age determination (Banerjee et al., 2001; Wintle and Murray, 2006).

Studies in Europe has also been on the forefront of research which compare OSL and radiocarbon dates, since they have the advantage of hundreds of years of written records with which to compare dates. OSL dates on eolian sand in North Wales, UK have correlated to within 11 years of an historical record of a dune’s construction (± 0.14 ka; Bailey et al., 2001). OSL also shows comparable (or better) matches with historical records and radiocarbon dates on peaty paleosols between eolian sand deposits in Jutland, Denmark (Murray and Clemmensen, 2001).

A thorough comparison of radiocarbon and OSL dates was completed by Wilson et al. (2004), who found excellent agreement between the two. The success of OSL has lead to its heavy use in dating eolian activity (Wilson et al., 2004), as radiocarbon dates on paleosols must postdate the material in which they are formed, and predate any overlying sediment (e.g., Bailey et al., 2001; Roberts and Plater, 2007).

Nielsen et al. (2006) used OSL to date an extensive beach ridge complex near Jerup, Northern Jutland, Denmark. They found that beach ridges began forming about 2.7 ka and halted at 1 ka, even though sea levels continued to regress. Despite the discrepancy with sea level changes, Nielsen et al. (2006) found their OSL ages to be quite reliable. In most cases OSL samples within the same unit are chronologically consistent, and correlate well to radiocarbon ages.

Nielsen et al. (in press) reexamined the beach ridge complex near Jerup, and incorporated more OSL samples into their chronology. The results indicate that the rate of dune ridge formation is highly variable, ranging from nearly 3 to 90 years per ridge, with an average of 24 years per ridge. The rates of ridge formation seem to respond to disruptions in sediment supply caused by a large sea spit to the north. Periods of rapid beach ridge growth from 4 to 2 ka correlate to a period of low sediment deposition at the spit. Conversely, the rate at which beach ridges formed decreased from 2 to 1 ka while the spit grew considerably. The growth of the spit, and therefore the sediment supply at Jerup, was linked to changes in sea level.

In summary, a significant body of research exists relating the evolution of coastal sand dunes to the elevations of water bodies, both in the Great Lakes and abroad. The destabilizing nature along the Lake Michigan shore due to lake-level fluctuations, in conjunction with a well established chronology of lake levels, allows researchers to test the perched-dune model in several different dune environments. All of these factors come together nicely at high-perched dune fields in northwest Lower Michigan, where I tested the perched-dune model using OSL.

CHAPTER 3

STUDY AREA

This study focuses on the history of eolian sand deposition on a high-perched dune field in northwest Lower Michigan, just north of the Manistee-Benzie County line (Figure 3:1). The property is owned by the Grand Traverse Regional Land Conservancy (GTRLC), and is called the *Arcadia Dunes Conservation Area* by that organization. Although this land parcel includes approximately 2428 ha of forested and grassland nature preserve, the study area specifically centers on the coastal sand dunes perched on the Manistee moraine within an approximately 3 km^2 area between highway M-22 and the lakeshore (Figure 3:1). As a result of this focus, the study area is referred to as *Arcadia Dunes* for the remainder of this thesis.

Land Ownership History

The Manistee-Benzie County area was originally occupied by the Native Americans of the Ottawa tribe. They farmed, hunted, and fished in the lakeshore area during the summer and fall, and migrated inland during the winter season (Arcadia Area Historical Society, 2008). The area was largely unsettled by Europeans until the 1860's, when Arcadia became one of the many logging boomtowns in northern Lower Michigan (Arcadia Area Historical Society, 2008). Arcadia was particularly active throughout the 1880's and 1890's, thanks to the construction of a narrow gauge railway used for timber extraction and access to Lake Michigan shipping. Much of the land was owned by sawmills and lumber

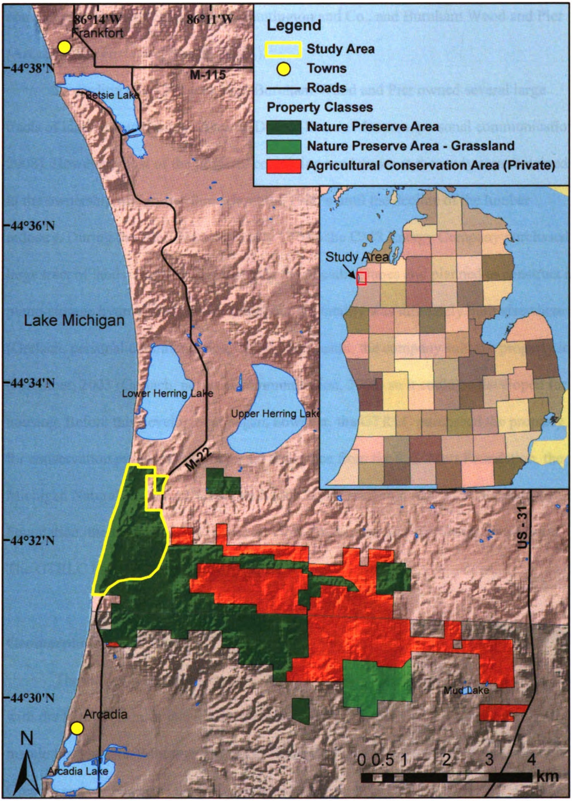

Figure 3:1 – Location of the study area within the Arcadia Dunes Conservation Area. The focus of this study is the area between highway M-22 and the lake, within the boundaries of Arcadia Dunes.

companies such as the Starke Bros., Huntington and Co., and Burnham Wood and Pier (Arcadia Area Historical Society, 2008).

According to a 1901 plat map, Burnham Wood and Pier owned several large tracts of land surrounding the Arcadia Dunes (Benzie Museum, personal communication, 2008). However, most of the property containing sand dunes at the study area remained in the ownership of families throughout the 1900's until the decline of the lumber industry. During the late 1960's and early 1970's the CMS Energy Company purchased a large tract of land in the area that contains the Arcadia Dunes and planned to construct a pump-storage hydroelectric power plant, but the facility was ultimately built elsewhere (Gerlach, personal communication, 2008). As a result, the company sold the property in December, 2003 (Gerlach, personal communication, 2008) so it could be developed for housing. Before this development began, however, the GTRLC purchased the property for conservation purposes with financial assistance from the C.S. Mott Foundation, the Michigan Natural Resources Trust Fund, the Carls Foundation, the W. K. Kellogg Foundation, the Herbert H. and Grace A. Dow Foundation and the Kresge Foundation. The GTRLC has owned the property since.

Geomorphology

The early geomorphic history of the study area is closely tied to events associated with the Late Wisconsin glaciation of northern Lower Michigan. The Arcadia Dunes mantle the Manistee Moraine (Dow, 1937), which is generally associated with the Greatlakean readvance of the Lake Michigan lobe at about 14,000 cal. yr. B.P. (Larson and Schaetzl, 2001; Evenson et al., 1976). During this most recent glacial period, ice advanced as small sublobes into pre-existing lowlands formed during previous glacial stages (Leverett and Taylor, 1915). This ice deposited the sandy till and outwash that composes the Manistee Moraine (Leverett and Taylor, 1915; Farrand, 1988; Blewett, 1990). The moraine is characterized by irregular ridges that extend perpendicular to the shoreline in the Manistee-Benzie county area (Figure 3:2; Farrand and Bell, 1982). As the sublobes retreated, they continued to deposit outwash which covered the moraine and filled the lowlands. These lowlands were later filled with lacustrine sediments from subsequent high lake phases (Dorr and Eschman, 1970).

In association with this history, the study area contains four geomorphic features: active dunes, stable dunes, glacial uplands, and a lake plain/ridge-swale complex. To the north and south of the study area are lake plains and ridge-swale complexes which separate high ridges of the Manistee Moraine (Figure 3:3). However, the entire study area is located on top of glacial uplands (Figure 3:3).

Dunes in the study area consist mostly of long ridges that generally run parallel to the shoreline (Figure 3:4). Dow (1937: p. 430) named the major lake-ward ridge "razorback". In turn, this ridge contains a "wind-rift" (Figures 3:4, 3:5; Dow, 1937: Plate II, Fig. 2) that is a large blowout which extends inland approximately 400 m. Extending from this blowout are the limbs of a parabolic dune named "Old Baldy" (Figure 3:6). Several stable, densely vegetated dune ridges occur northwest of Old Baldy and extend to the Herring Lake plain a few kilometers to the north (Figure 3:4). Erosion of the active lake-ward dune ridges has revealed an excellent sequence of buried soils and associated

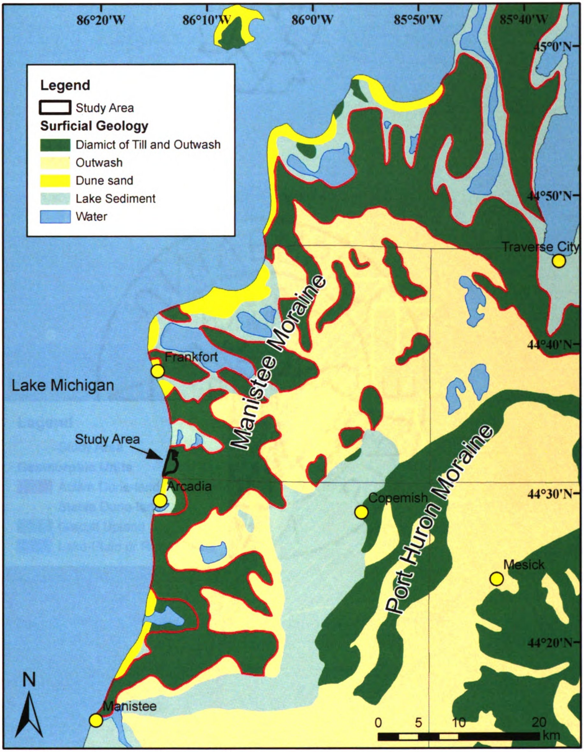

Figure 3:2 – Surficial geology of northwest Lower Michigan. (Source: Michigan CGI, 2008). The Manistee Moraine (after Dow, 1937; and Farrand, 1988) is outlined. Note its looping nature north and south of the study area.

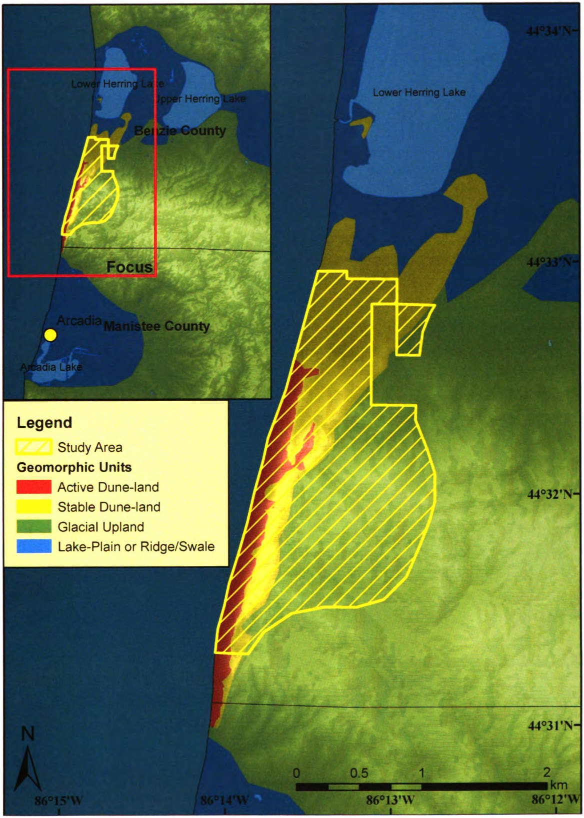

Figure 3:3 –Gemorphic units of Arcadia Dunes and the surrounding area.

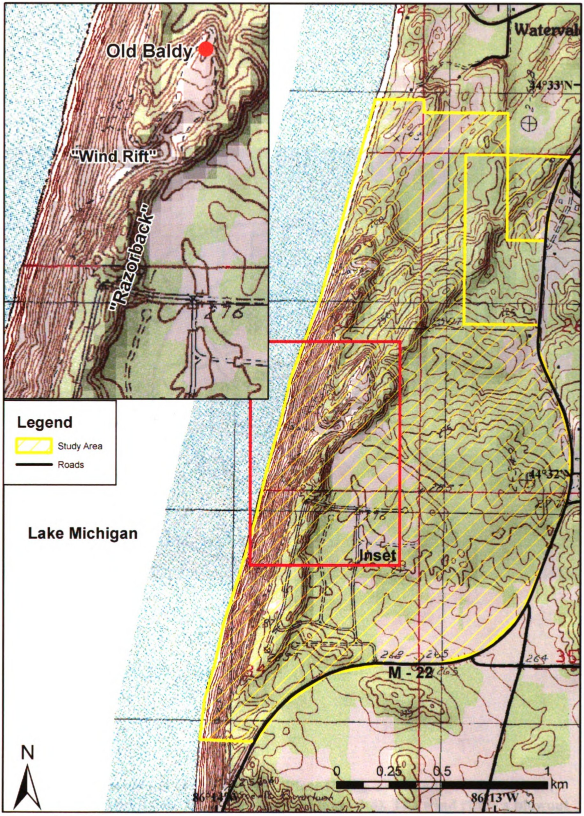

Figure 3:4 – Topographic map of the study area (yellow). Focus box is on the landform feature names given by Dow (1937). Contour interval: 5 m.

Figure 3:5 – Photograph of Dow's (1937: Plate II, Fig. 2) "Wind rift" taken from near the top of Old Baldy, facing southwest (Figure 3:4). The wind rift is a blowout formed in Dow's (1937) "Razorback", which is the large vegetated ridge on the left of the image (Figure 3:4). Photographed by author.

Figure 3:6 – Photograph of "Old Baldy", the crest of a large, active parabolic dune at Arcadia Dunes (Figure 3:4). Background: view of Lake Michigan to the north-northwest, and Lower Herring Lake on the right side of the image. Photographed by author.

units of eolian sand (Figure 3:7). These features are the primary focus of this investigation because they were used to reconstruct periods of eolian sand deposition and stability in the dune field.

Climate

The climate of the study area is strongly influenced by the close proximity of Lake Michigan and the prevailing westerly winds (MDACP, 1989). The nearest coastal climate station is at Manistee, Michigan (44°, 13' N, 86° 17' W), approximately 33 km to the south-southwest of the study site. Data from this station are representative of the study area.

The climate is classified as humid continental, with warm summers and cold winters. The monthly high temperature ranges between -5.2° C in January to 20.7° C in July, with a mean of 8.1° C annually. From a statistical perspective, the annual growing season ranges from the last and first freezes of the year, which occur on May 10 and October 16, respectively (MDACP, 1989).

Average annual precipitation is 78.5 cm and is spread fairly evenly throughout the year. Much of this precipitation comes in the form of snowfall, which averages 234.2 cm annually. The study area has a snow-pack of at least 2.5 cm an average of 91 days per season. The proximity of Lake Michigan increases cloudiness and precipitation during the fall and early winter. These effects are reduced when ice builds up on the lake during late winter (MDACP, 1989).

Figure 3:7 – View of one of the lake-ward exposures at Arcadia dunes. Background shows Lake Michigan and other high-perched dunes on the Manistee Moraine to the north. L. Crosby Savage III sits for scale on one of the lake-ward exposures in which a buried soil outcrops. Photographed by author.

Soils

The soils at the study area are consistent with soils elsewhere in northern Lower Michigan that have formed in sandy sediments, and can be generalized on the basis of their development (Kroell III, 2008). The densely vegetated portions of the moraine contain soils of the Kaleva-Grattan and Kaleva-Benona associations, which are level to very steep, excessively drained, sandy soils on lake plains, outwash plains, and sand dunes. Common soil series of these associations are the Benzonia, Benona and Grattan. These series are Spodosols which usually develop inland on the moraine, especially in low or flat areas and contain moderately well developed spodic horizons. The Benzonia series is a sandy, isotic, mesic Lamellic Haplorthod, and is the best developed of these soils because it has an A-E-Bhs1-Bhs2-Bs1-Bs2-E&Bt horizonation, followed by the Benona series, a sandy, mixed, mesic Lamellic Haplorthod with A-E-Bs1-Bs2-Bw-E'-E&Bt horizonation. The least developed soil is the Grattan series, a sandy, mixed, mesic Entic Haplorthod with A-E-Bs1-Bs2-BC-C horizonation (Kroell III, 2008).

Most soils at Arcadia Dunes are Entisols, such as the Nordhouse, Coloma, and Plainfield sands of the Nordhouse association. These soils form on level to very steep terrain and are excessively drained. The best developed of these soils is the Nordhouse series, a mesic, uncoated Spodic Quartzipsamment with A-E-Bs-C1-C2 horizonation. The Coloma series is a mixed, mesic Lamellic Udipsamment, with A-E-Bw1-Bw2-Bw3-Bw4-E&Bt horizonation. The Plainfield sands are very poorly developed, mixed, mesic Typic Udipsamments with A-Bw1-Bw2-BC-C horizonation. The majority of Entisols in the active dune areas are an Udorthent-Udipsamment complex, which are Entisols formed in sandy and coarse-loamy materials. They usually have an A-C horizonation, and may contain a weakly developed B horizon (Kroell III, 2008).

Vegetation

Pre-settlement vegetation of the study area was primarily a mixed northern hardwood forest comprised of the dominant beech (*Fagus grandifolia*), sugar maple (*Acer saccharum*), and hemlock (*Tsuga Canadensis*; Michigan CGI, 2008). No evidence exists indicating that the entire study area was ever clear-cut, although a small part of the moraine was once cleared for agriculture. Portions of this area were later replanted as stands of Norway spruce (*Picea abies*) and pine (*Pinus*; Gerlach, personal communication, 2008).

The current vegetation generally depends on the water-holding capacity of the varying surficial sediments within the study area. The driest areas, such as beaches or active dunes, are typically not vegetated, or are limited to xerophytic species. The most prominent species are grasses, such as marram grass (*Ammophila breviligulata*) and wheat grass (*Agropyron*), and several species of dune willows (*Salix*; e.g., Olson, 1958b). They dominate the foredunes and beach due to their ability to withstand burial events.

In contrast to the active dunes, more stabilized areas have been colonized by shrubs like red ozier dogwood (*Cornus stolonifera baileyi*), choke cherry (*Prunus virginiana*) and sand cherry (*P. pumila*) (Olson, 1958b). The very steep, but relatively stable slopes also have large areas of moss (*Polytrichum pilferum*), which grow in large colonies (e.g., Marsh and Koerner, 1972).

Most of the study site is densely forested. These areas have been stable for long periods of time, with common tree species including American beech (*Fagus grandifolia*), sugar maple (*Acer saccharum*), red maple (*Acer rubrum*), bigtooth aspen (*Populus grandidentata*), quaking aspen (*Populus tremuloides*), northern red oak (*Quercus rubra*), eastern hemlock (*Tsuga Canadensis*), eastern white pine (*Pinus strobus*), red pine (*Pinus resinosa*) and jack pine (*Pinus banksiana*; Kroell III, 2008). Understory vegetation includes trillium (*Trillium grandiflorum*), brackenfern (*Pteridium aquilinum*), jack-in-the-pulpit (*Arisaema triphyllum*), lily of the valley (*Maianthemum canadense*), wintergreen (*Gaultheria procumbens*), twisted stalk (*Streptopus amplexifolius*), bloodroot (*Sanguinaria canadensis*), Dutchman's breeches (*Dicentra cucullaria*), Canada mayflower (*Maianthemum canadense*), true Solomon's seal (*Polygonatum biflorum*), American starflower (*Trientalis borealis*), and trout lily (*Erythronium americanum*; Kroell III, 2008).

CHAPTER 4

METHODS

The goal of this study was to reconstruct the evolution of the Arcadia Dune field and to test the validity of the perched-dune model. This goal was accomplished using a variety of field and laboratory methods, including landscape interpretation, surveying, soil description, sediment analysis, and radiometric dating techniques. The following chapter describes these methods.

Geomorphic Identification

Before I entered the field, I attempted to learn as much as I could about the Arcadia Dunes through topographic analysis and map interpretation. I first analyzed the landscape on the Elberta, MI 7.5 minute topographic map and digitized the landforms in ArcMap 9.2 (ESRI, 2008). In this process I included the identification of prominent dune ridges named by Dow (1937) in his investigation of the study area. The resulting geomorphic features were then integrated with a state-wide surficial geomorphology map (Farrand and Bell, 1982) to properly identify their extent in the Manistee-Benzie area.

After my initial map investigations, I began field reconnaissance of the study area. During this time the geomorphic maps were field-checked for accuracy. Also, areas were selected to gather information about the chronology of eolian activity at the study area. In particular, I looked for outcrops in eroded sand dunes which contained buried soils. As a result of this work, three exposures were identified (Figure 4:1). Two of the exposures

Figure 4:1 – Aerial photograph of the exposures of the study area, subset shows the exposures on topographic map. "Old Baldy" is identified for reference. Contour interval: 5 m.

were eroded out of the lake-ward dune ridge (Figure 4:1; Exposures 1 and 2), while the third was in a blowout of an inland dune (Figure 4:1; Exposure 3).

Stratigraphic Survey and Soil Sampling

Once the exposures were discovered, I subsequently devoted my time to learning as much about them as possible. The first step in this assessment involved an elevation survey to determine the stratigraphy of buried soils and sediment units displayed at the exposures.

The two lake-ward exposures were surveyed using a device I constructed to measure tall vertical exposures. This survey device consisted of a 3-m pole that was used to form a right angle to a piece of measured string extended up the slope. Application of this survey instrument in the field required three people. One person held the pole plumb (with the use of a level) while another climbed immediately upslope to measure the horizontal distance. While this measuring occurred, the third person insured that the string formed a right angle by visually checking the string with a T-square affixed to the top of the pole. This technique was repeated for both lake-ward exposures to establish stratigraphic boundaries and to calculate their elevations above Lake Michigan. The elevation of the third exposure was estimated based on the USGS topographic map.

In the process of surveying the site stratigraphy, special care was taken to identify and measure the elevation of buried soils that outcropped in the exposures. When soils were identified they were cleaned and described for horizonation, thickness, depth from the surface, boundary, field-moist color, structure, and texture according to NRCS standards (Schoenberger et al., 2002). One sample of sediment was also taken from each horizon for laser particle size analysis. At the contact between eolian sand and underlying glacial sediments, samples were taken every ten centimeters to identify the exact location of the lithologic boundary.

Age Determination of Eolian Sediments

After the soils were identified, described, and sampled, I turned my attention to acquiring samples for age determination. In this study, samples were collected from the C horizon of soils using PVC pipe (72 cm long) to prevent exposure to light. Each pipe was wrapped in electrical tape, driven into the cleaned pit face, and then sealed with duct tape upon extraction. A large amount of the surrounding sand was also collected to calculate the dose rate. In the case of very thick eolian units, samples were collected from the upper and lower portions of the deposit to provide bracketing ages for their deposition. Twelve samples were collected in total, five from exposure one, six from exposure two, and one from exposure three. The samples were sent to the Luminescence Dating Laboratory at the University of Illinois at Chicago where they were analyzed using the Single Aliquot Regenerative Dose (SAR) method (Wintle and Murray, 2006) by Steve Forman and his personnel.

OSL ages acquired in this study are reported in ka (1,000 years) before 2000 A.D., with one standard deviation of error (e.g., Wintle and Murray, 2006). The luminescence curves are assumed to have a normal distribution, which implies that about 95% of the data were within two standard deviations of the mean OSL age. The ages in this study are reported at 1 σ, which means that there is a 68% probability that the deposition of eolian sands occurred during that time frame.

Sediment Texture Analysis

In order to accurately describe the stratigraphy of the exposures, sediment samples were checked using particle size analysis. This information is used to describe and identify the type of material in which buried soils at the study site formed.

Particle size analysis of soils was run in the Geography Department at Michigan State University, East Lansing, MI. Samples from each soil horizon were dried in a 65° C oven overnight to remove moisture. Each sample was then put through a 2000 µm sieve to remove any coarse or organic fragments. Following this step, each sample was homogenized by running them three times through a sample splitter. A small aliquot (~ 2 grams) of each sample was then placed in a vial, along with distilled water and five ml of dispersant ($(NaPO_3)_6$ and Na_2CO_3). The samples were shaken for three hours prior to particle size analysis to disperse the sediment.

Particle size distributions were measured using a Malvern Mastersizer 2000 laser particle size analyzer with a Hydro MU manual wet dispersion unit. The machine uses the diffraction of a laser beam through the suspended sediment to calculate particle size distributions between 0.02 and 2000 µm (Malvern Instruments Ltd., 2008). The output of this system is a table describing the percent of each particle size range, as set by the user.

In summary, geomorphic examination and mapping was used to identify suitable locations for stratigraphic analysis. Three exposures containing a stacked sequence of eolian units were surveyed for stratigraphy and described for soil development using profiles. Exposure 1 was analyzed using four profiles (1A-1D), Exposure 2 using three profiles (2A-2C), and Exposure 3 using one profile (3A). OSL and sediment samples were collected from the exposures, and are used to interpret the eolian history of Arcadia Dunes in Chapter 5.

CHAPTER 5

RESULTS AND DISCUSSION

This chapter reconstructs the record of eolian sand deposition at Arcadia Dunes and relates it to other studies throughout the region. First, the stratigraphy and OSL dates acquired from the three exposures within the study area will be discussed. The chronology of eolian activity is then discussed in the context of the perched-dune model and compared to results from other relevant studies throughout the region.

Dune Stratigraphy

This section discusses the stratigraphy of the three exposures outlined in Chapter 4 (Figure 4:1). These exposures contain a stacked sequence of eolian sand units that are bounded by paleosols. The sequence of these deposits is first discussed, with a focus on the thickness of the eolian units and their associated OSL ages. These ages are presented in Table 5:1. The nature and general development of paleosols contained within the units is also discussed. The information from the stratigraphy and OSL dates is used to interpret periods of eolian activity in the remainder of this chapter.

Exposure One

Exposure 1 (Figure 5:1) is located in the southern part of the study area (Figure 4:1). This exposure is about 130 m tall, extending from the beach to the top of the dune where it reaches an elevation of 310 m above sea level (masl). This exposure is quite steep, with a slope of approximately 61%.

Table 5:1 – Single aliquot regeneration ages for the three exposure at the high-perched dunes at Arcadia Dunes in northwest Lower Michigan.

Laboratory #	Aliquots	Equivalent Dose (Gy)[a]	U (ppm)[b]	Th (ppm)[b]	K^2O (%)[b]	Cosmic dose (Gy/ka)[c]	Dose rate (Gy/ka)[d]	SAR age (ka)[e]
Exposure 1								
UIC2128	32/40	0.96 ± 0.06	0.9 ± 0.1	3.1 ± 0.1	1.99 ± 0.02	0.18 ± 0.02	2.36 ± 0.11	0.400 ± 0.040
UIC2124	32/40	1.87 ± 0.12	0.8 ± 0.1	2.0 ± 0.1	1.97 ± 0.02	0.06 ± 0.01	1.91 ± 0.10	0.970 ± 0.010
UIC2121	39/40	3.34 ± 0.25	0.5 ± 0.1	1.7 ± 0.1	1.82 ± 0.02	0.04 ± 0.01	1.89 ± 0.10	1.765 ± 0.190
UIC2119	40/40	6.56 ± 0.41	0.7 ± 0.1	2.1 ± 0.1	1.72 ± 0.02	0.04 ± 0.01	1.88 ± 0.10	3.495 ± 0.335
UIC2118	40/40	10.15 ± 0.44	0.9 ± 0.1	2.9 ± 0.1	2.75 ± 0.02	0.02 ± 0.01	2.88 ± 0.14	3.530 ± 0.300
Exposure 2								
UIC2122	41/50	0.79 ± 0.11	0.6 ± 0.1	2.0 ± 0.1	2.30 ± 0.02	0.18 ± 0.02	2.49 ± 0.12	0.310 ± 0.050
UIC2123	29/40	1.30 ± 0.15	0.6 ± 0.1	2.0 ± 0.1	1.84 ± 0.02	0.15 ± 0.02	2.05 ± 0.10	0.630 ± 0.085
UIC2129	38/40	1.70 ± 0.15	0.7 ± 0.1	2.2 ± 0.1	2.28 ± 0.02	0.03 ± 0.01	2.38 ± 0.11	0.710 ± 0.080
UIC2125	39/40	2.39 ± 0.17	0.7 ± 0.1	2.6 ± 0.1	2.53 ± 0.03	0.03 ± 0.01	2.62 ± 0.13	0.910 ± 0.095
UIC2127	37/40	7.04 ± 0.32	0.5 ± 0.1	1.7 ± 0.1	1.54 ± 0.02	0.02 ± 0.01	1.62 ± 0.08	4.340 ± 0.380
UIC2120	40/40	12.64 ± 0.82	0.8 ± 0.1	2.9 ± 0.1	2.70 ± 0.02	0.02 ± 0.01	2.80 ± 0.14	4.500 ± 0.445
Exposure 3								
UIC2126	37/40	9.94 ± 0.59	0.6 ± 0.1	1.9 ± 0.1	2.26 ± 0.02	0.17 ± 0.02	2.44 ± 0.12	4.070 ± 0.380

[a]Equivalent dose determined by the single aliquot regenerative dose method under blue light excitation (470 nm) (Murray and Wintle, 2003)on the 150-250 μm quartz fraction.

[b]U, Th and K20 values determined by ICP-MS, Activation Laboratory Ltd., Ontario.

[c]Contains a cosmic rate dose rate component from Prescott and Hutton (1994).

[d]A long-term moisture content of 5 ± 2% was assumed.

[e]All errors are at one sigma and ages are from reference year AD 2000. Analyses preformed by Luminescence Dating Research laboratory, Dept. of Earth & Environmental Sciences, Univ. of Illinois-Chicago.

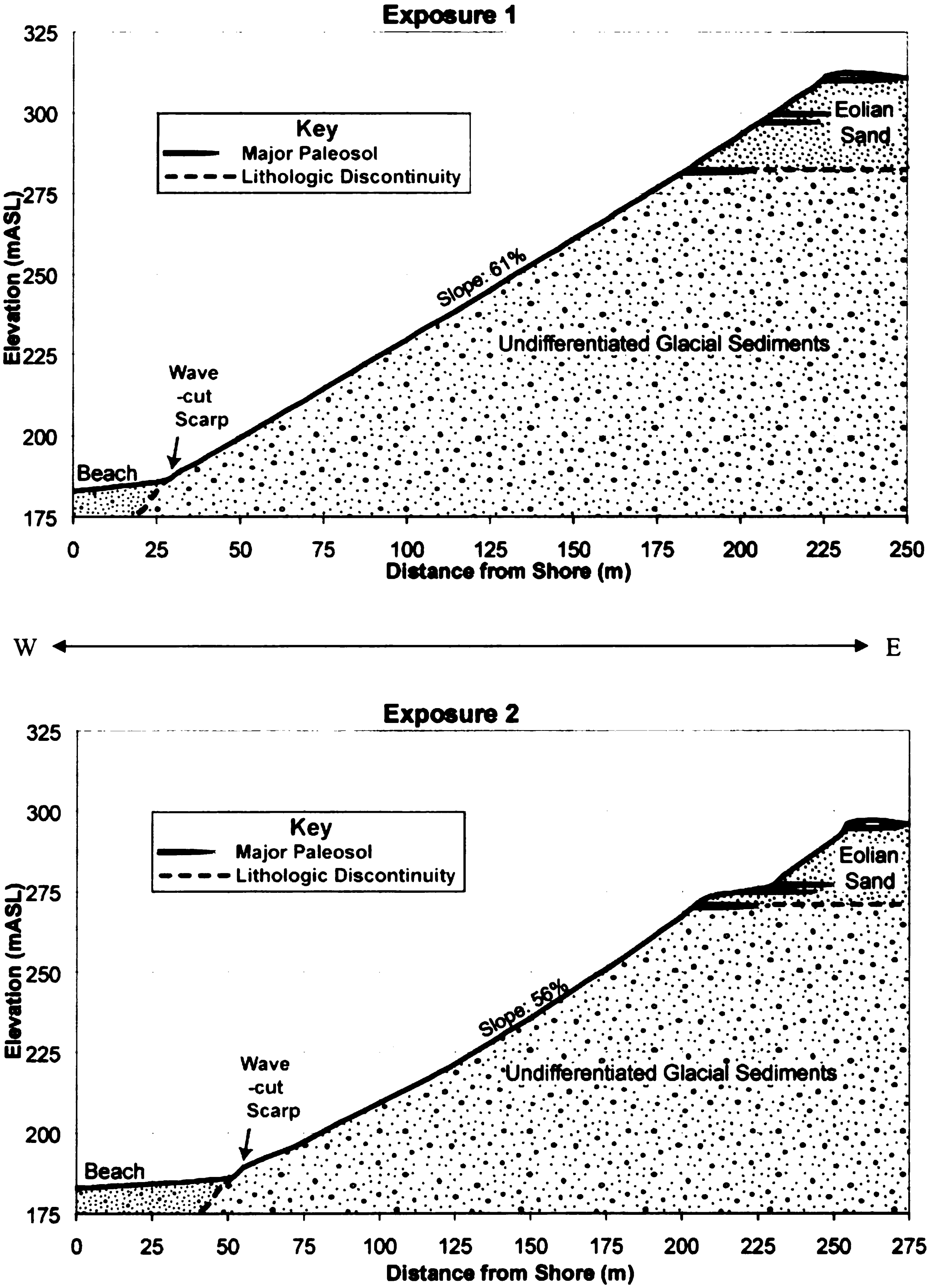

Figure 5:1 – Profile view and general stratigraphy of the two lake-facing bluff exposures.

Exposure 1 contains seven stratigraphic units that were visible in this study (Figure 5:2). The basal unit (Unit I) consists of undifferentiated glacial sediment that is approximately 100 m thick above lake level. In general, the sediment in Unit I consists of mixed sand and gravel. Particle size analysis of six samples from the uppermost meter of Unit I indicate that the sediment in this part of the unit is comprised of 90% to 100% sand, with the following fractions: very coarse sand (<1%), coarse sand (14.2%), medium sand (54.9%), fine sand (30.0%) and very fine sand (0.8%). Coarse fractions ranged from about 0% to 15% by mass, with occasional cobbles and boulders in the uppermost meter which were not collected from the unit.

Formed in the uppermost part of the glacial sediments is an Entisol, trending towards a Spodosol, that occurs at an elevation of 284 masl and at a depth of 28.43 m from the surface of the exposure. This soil was described in soil profile 1D (Figure 5:3, Table 7:1) and exhibits A-E-Bs-C horizonation with the following thickness and colors, respectively: 7Ab (4 cm thick; 10YR 3/3), 7Eb (20 cm thick; 10YR 5/4), 7Bsb (43 cm thick; 7.5YR 4/6), and 7C (>20 cm thick; 10YR 6/4).

Overlying Unit I is about 28 m of eolian sand that extends to the top of the exposure. This contact is marked by an abrupt shift in the particle size distribution from coarse sand and gravel of the glacial sediment to fine, well-sorted eolian sand (Figure 5:4). The eolian sand at this contact ranges from 86% to 95% pure sand, with 3% to 12% silt and 0.8% to 2% clay. The breakdown of the sand fraction is approximately: very coarse sand (<1%), coarse sand (1.7%), medium sand (29.5%), fine sand (45.9%), and very fine sand (18.7%).

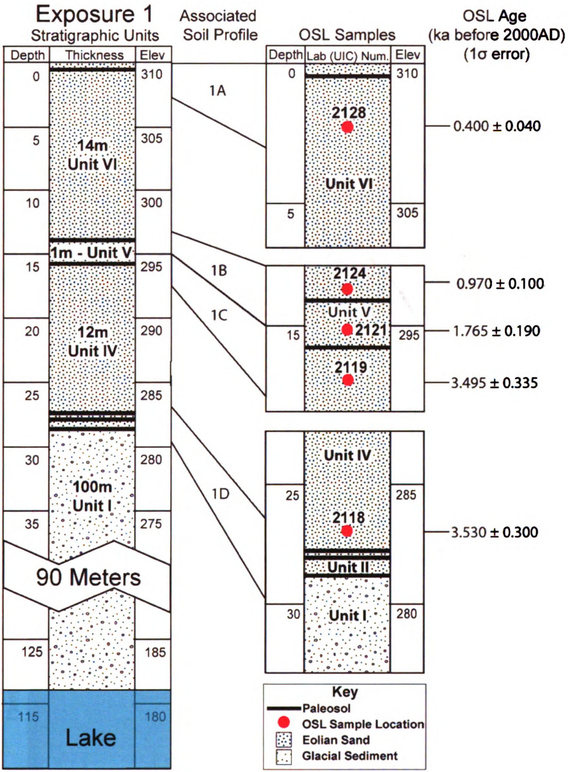

Figure 5:2 – Generalized stratigraphy of Exposure 1. Major eolian units and paleosols are shown in relation to their associated soil profile. Generalized OSL sample locations are labeled with UIC sample number and associated age. Depth and elevation in meters.

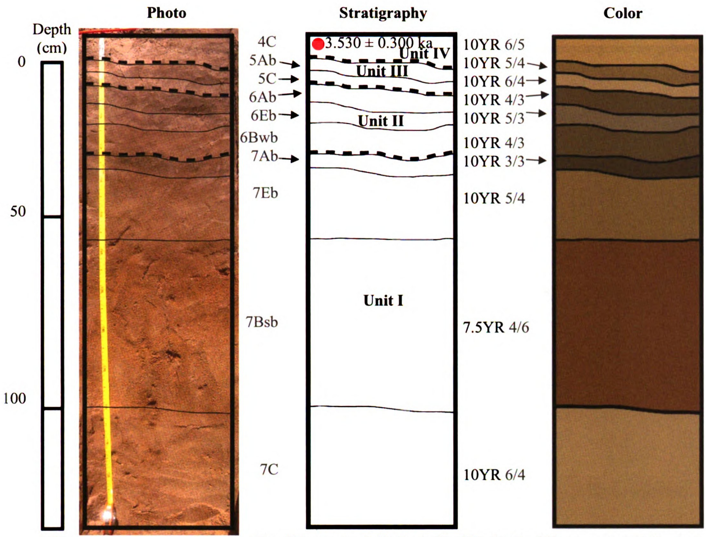

Figure 5:3 – The lowermost soils of Exposure 1 (Soil profile 1D). Dashed lines represent the boundary between stratigraphic units. Soils 5 and 6 formed in fine eolian sand. Soil 7 formed in glacial sediment. Red dots are representative locations of OSL samples.

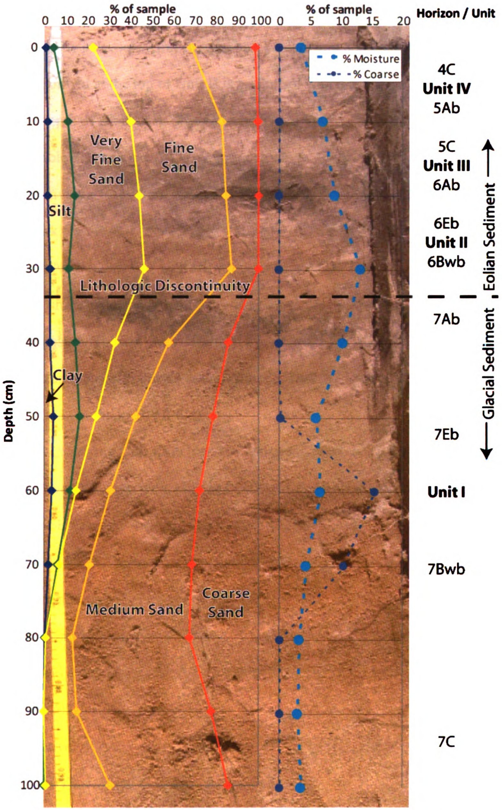

Figure 5:4 – Depth plot of texture, moisture, and coarse fraction content across the eolian sand/glacial sediment contact of Exposure 1.

Contained within the eolian sand deposits are six stratigraphic units which are bounded by paleosols. Unit II is the basal eolian unit and consists of about 23 cm of fine sand. Formed at the top of Unit II is an Entisol which occurs at an elevation of 284 masl and a depth of 28.20 m in the exposure. This soil was described in soil profile 1D (Figure 5:3, Table 7:1) and has A-E-Bw horizonation, with thickness and colors values of (respectively): 6Ab (6 cm thick; 10YR 4/3), 6Eb (7 cm thick; 10YR 5/3), and 6Bwb (10 cm thick; 10YR 4/3).

Above Unit II is Unit III, which consists of 10 cm of fine sand. Formed at the top of this unit is an Entisol that occurs at an elevation of 284 masl and a depth of 28.10 m in the exposure. This soil was described in soil profile 1D (Figure 5:3, Table 7:1) and has A-C horizonation, with thickness and colors of (respectively): 5Ab (5 cm thick; 10YR 5/4), and 5C (5 cm thick; 10YR 6/4).

Overlying Unit III is Unit IV, which is a relatively thick (11.8 m) deposit of fine sand. Two OSL ages were obtained in this Unit. The lowermost sample was collected near the base of the deposit from a depth of 27.95 m and returned an age of 3.53 ± 0.3 ka[*] (UIC 2118). The uppermost sample was obtained near the top of the unit from a depth of 16.31 m and provided an age of 3.495 ± 0.335 ka (UIC 2119). Formed in the uppermost part of Unit IV was a weakly developed Spodosol, which occurs at an elevation of 297 masl, and a depth of 15.53 m in the exposure. This soil was described in soil profile 1C (Figure 5:5, Table 7:2) and has A-E-Bs-C horizonation, with thickness and colors of (respectively): 4Ab (10 cm thick; 10YR 3/4), 4Eb (29 cm thick; 10YR 6/4), 4Bsb (30 cm thick; 10YR 5/6), 4C (>25 cm thick; 10YR 6/5).

[*] OSL ages acquired in this study are provided in ka, or kiloannum (1,000 years) before 2000 A.D.

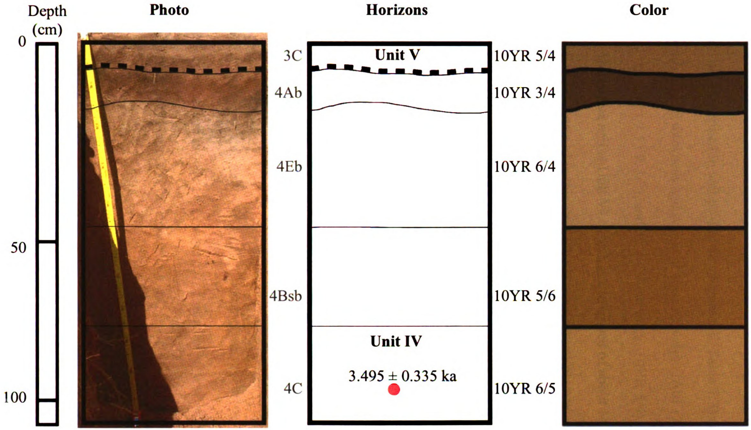

Figure 5:5 – The lowermost of the two middle soils of Exposure 1 (Soil profile 1C). Dashed line represents the boundary between stratigraphic units. These soils formed in fine eolian sand. Red dots are representative locations of OSL samples.

Continuing upward in the eolian sequence, the next deposit is Unit V, which is a thin (~ 1 m) unit of fine sand. An OSL sample was collected from this unit at a depth of 14.86 m and returned an age of 1.765 ± 0.19 ka (UIC 2121). Formed at the top of unit V is an Entisol which occurs at an elevation of 298 masl, and a depth of 14.6 m in the exposure. This soil was described in soil profile 1B (Figure 5:6, Table 7:3) and exhibits A1-A2-C horizonation, with thickness and colors of (respectively): 3Ab1 (7 cm thick; 5YR 2.5/1), 3Ab2 (4 cm thick; 7.5YR 3/2), and 3C (>70 cm thick; 10YR 5/4).

Above that soil is Unit VI, which is comprised of about 14 m of fine sand. Two OSL ages were obtained in this unit. The lowermost sample was collected near the base at a depth of 14.23 m and returned an age of 0.970 ± 0.1 ka (UIC 2124). The uppermost sample was obtained from the top of the unit at a depth of 1.32 m and provided an OSL age of 0.4 ± 0.04 ka (UIC 2128). Formed at the uppermost part of Unit VI is a weakly developed Spodosol which occurs at a depth of 22 cm. This soil was described in soil profile 1A (Figure 5:7, Table 7:4) and exhibits A-E-Bs-C horizonation, with the following thickness and colors, respectively: 2Ab (19 cm thick; 10YR 3/1), 2Eb (6 cm thick; 10YR 6/2), 2Bsb (7 cm thick; 10YR 6/3), and 2C (>48 cm thick; 10YR 7/2).

Unit VI is buried directly by Unit VII, which is comprised of 22 cm of fine sand. Formed at the top of this unit is an Entisol which occurs at the modern surface. This soil was described in soil profile 1A (Figure 5:7, Table 7:4) and exhibits no discernable soil development; it contains a C horizon with a color of 10YR 4/3 and a thickness of 22 cm.

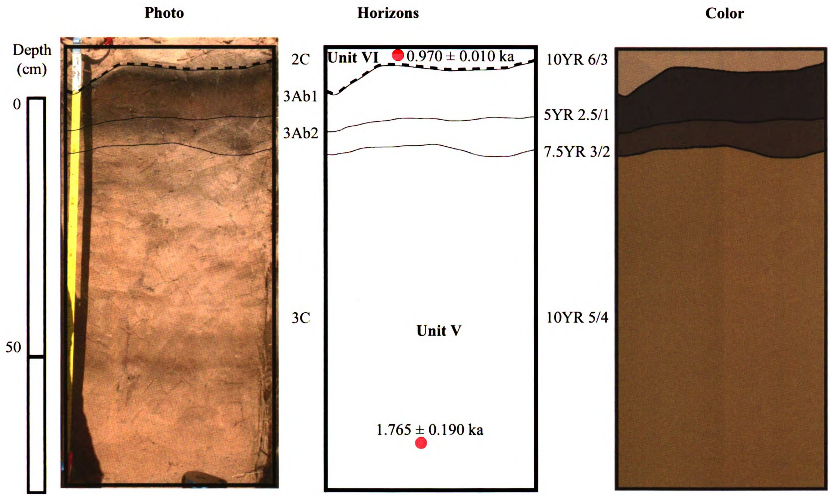

Figure 5:6 – The uppermost of the middle soils of Exposure 1(Soil profile 1B). Dashed line represents the boundary between stratigraphic units. These soils formed in fine eolian sand. Red dots are representative locations of OSL samples. The dark band at ~ 50 cm was not visible upon inspection of the profile, and is probably residual moisture left after cleaning the profile face for photography.

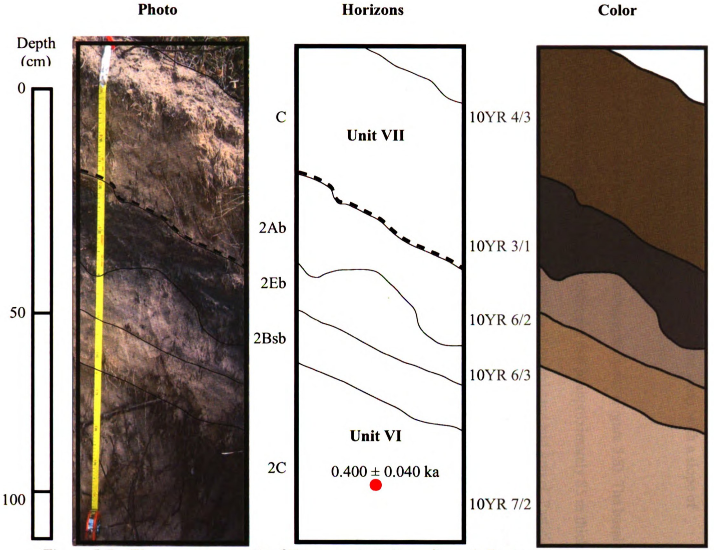

Figure 5:7 – The uppermost soils of Exposure 1 (Soil profile 1A). Dashed line represents the boundary between stratigraphic units. These soils formed in fine eolian sand. Red dots are representative locations of OSL samples.

Exposure Two

Exposure 2 is located in the central part of the study area (Figure 4:1) approximately 600 m north-northeast of Exposure 1. This exposure is about 115 m high, extending from the beach to the top of the dune where it reaches an elevation of 295 masl. Like Exposure 1, the surface of Exposure 2 is quite steep, with a slope of approximately 56% (Figure 5:1).

Exposure 2 contains eleven visible stratigraphic units (Figure 5:8). The basal unit (Unit I) consists of undifferentiated glacial sediment that is approximately 85 m thick above lake level. In general, this sediment consists of mixed sand and gravel. Particle size analysis of six samples from the uppermost meter of Unit I indicate that the sediment in this part of the unit is comprised almost entirely of sand, with the following fractions: very coarse sand (<1%), coarse sand (19.8%), medium sand (42.4%), fine sand (22.7%) and very fine sand (4.6%). Silt and clay ranged from 0% to about 5%. Coarse fractions ranged from about zero to 25% by mass, with occasional cobbles and boulders in the uppermost meter which were not collected from the unit.

Formed in the uppermost part of the glacial sediments is an Entisol with spodic characteristics that occurs at an elevation of 269 masl and at a depth of 26.06 m in the exposure. This paleosol was described in soil profile 2C (Figure 5:9, Table 7:5) and exhibits A-E-Bs-C horizonation with the following thickness and colors, respectively: 11Ab (4 cm thick; 10YR 3/2), 11Eb (15 cm thick; 10YR 5/4), 11Bsb (70 cm thick; 7.5YR 4/6), 11C (>10 cm thick; 10YR 5/6).

Overlying Unit I at Exposure 2 is about 26 m of eolian sand that extends to the top of the exposure. As at Exposure 1, this contact is marked by an abrupt shift in the

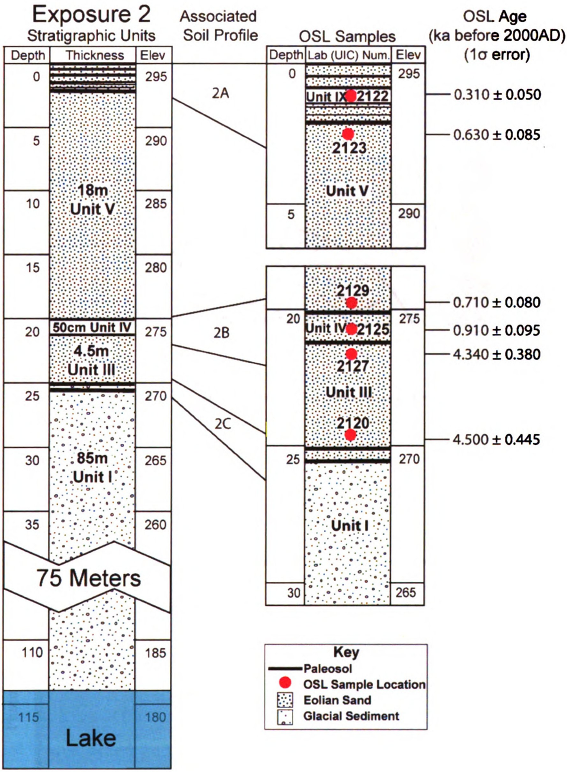

Figure 5:8 – Generalized stratigraphy of Exposure 2. Major eolian units and paleosols are shown in relation to their associated soil profile. Generalized OSL sample locations are labeled with UIC sample number and associated age. Depth and elevation in meters.

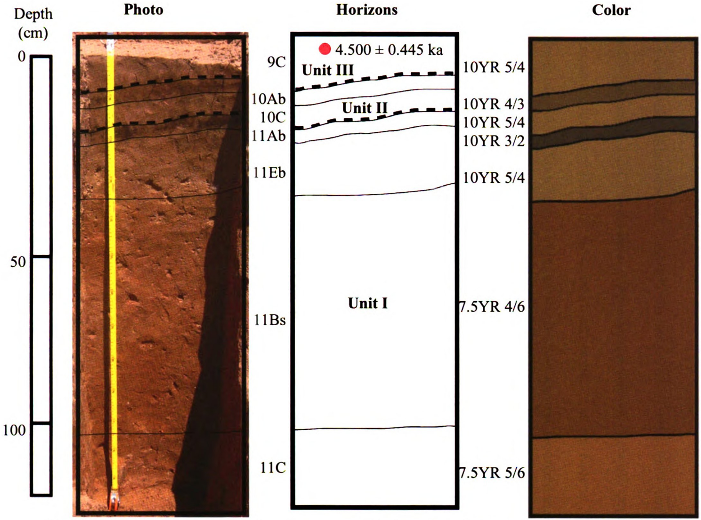

Figure 5:9 – The lowermost soils of Exposure 2 (Soil profile 2C). Dashed lines represent the boundary between stratigraphic units. Soils 10 formed in fine eolian sand. Soil 11 formed in glacial sediment. Red dots are representative locations of OSL samples.

particle size distribution from coarse sand and gravel of the glacial sediment to fine, well-sorted eolian sand (Figure 5:10). The eolian sand ranges from 89% to 99% pure sand, with 1% to 8% silt and 0.3% to 2% clay. The breakdown of the sand fraction is approximately: very coarse sand (<1%), coarse sand (<1%), medium sand (10.7%), fine sand (43.0%), and very fine sand (39.2%).

Contained within the eolian sand deposits are ten stratigraphic units, which are bounded by paleosols. Unit II is the basal eolian unit and consists of about 7 cm of fine sand. Formed at the top of Unit II is an Entisol which occurs at an elevation of 269 masl, and a depth of 25.98 m from the top of the exposure. This soil was described in soil profile 2C (Figure 5:9, Table 7:5) and has A-C horizonation with thickness and color values of (respectively): 10Ab (3 cm thick; 10YR 4/3), and 10C (5 cm thick; 10YR 5/4).

Overlying Unit II is Unit III, which is a 4.5 meter thick deposit of fine sand. Two OSL ages were obtained in this unit. The lowermost sample was collected near the base of the deposit from a depth of 25.58 m and returned an age of 4.5 ± 0.445 ka (UIC 2120). The uppermost sample was obtained near the top of the unit from a depth of 21.67 m and provided an age of 4.34 ± 0.380 ka (UIC 2127). Formed at the top of Unit III is a weakly developed Spodosol trending towards an Entisol, which occurs at an elevation of 274 masl, and a depth of 20.94 m in the exposure. This soil was described in soil profile 2B (Figure 5:11, Table 7:6) and exhibits A-E-Bs-C horizonation with the following thicknesses and colors, respectively: 9Ab (11 cm thick; 10YR 2/2), 9Eb (19 cm thick; 10YR 7/4), 9Bsb (48 cm thick; 10YR 5/8), 9C (>20 cm thick; 10YR 6/6).

Above Unit III is Unit IV, a thin (~1 m) unit of fine sand. An OSL sample was collected from this unit at a depth of 20.55 m and returned an age of 0.91 ± 0.095 ka

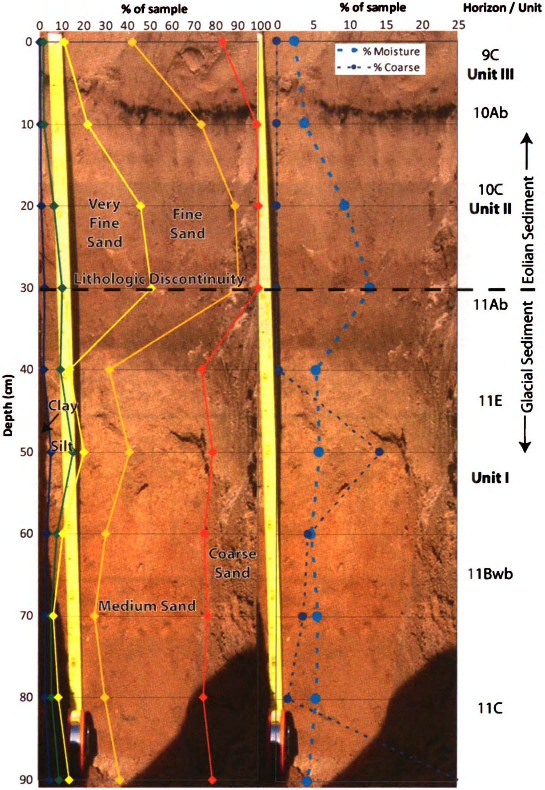

Figure 5:10 – Depth plot of texture, moisture, and coarse fraction content across the eolian sand/glacial sediment contact of Exposure 2.

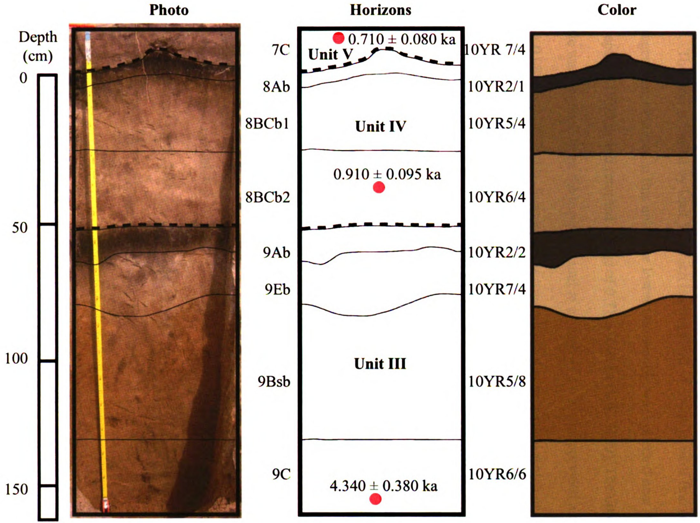

Figure 5:11 – The middle soils of Exposure 2 (Soil profile 2B). Dashed lines represent the boundary between stratigraphic units. These soils formed in fine eolian sand. Red dots are representative locations of OSL samples.

(UIC 2125). Formed at the top of Unit IV is an Entisol which occurs at an elevation of 275 masl and a depth of 20 m in the exposure. This soil was described in soil profile 2B (Figure 5:11, Table 7:6) and has A-BC1-BC2 horizonation, with thickness and colors of (respectively): 8Ab (8 cm thick; 10YR 2/1), 8BCb (22 cm thick; 10YR 5/4), and 8BCb2 (>27 cm thick; 10YR 6/4).

That soil is buried by Unit V, which is comprised of about 18 m of fine sand. Two OSL ages were obtained in this unit. The lowermost sample was collected near the base at a depth of 19.92 m and returned an age of 0.71 ± 0.08 ka (UIC 2129). The uppermost sample was obtained from the top of the unit at a depth of 2.6 m and provided a date of 0.63 ± 0.085 ka (UIC 2123). Formed at the top of Unit V is an Entisol which occurs at an elevation of 293 masl, and a depth of two meters in the exposure. This soil was described in soil profile 2A (Figure 5:12, Table 7:7) and has A-C horizonation with thicknesses and colors of (respectively): 7Ab (12 cm thick; 10YR 6/5), and 7C (>5 cm thick; 10YR 7/4).

Overlying Unit V is a sequence of about seven closely-spaced deposits of eolian sand that are separated by weakly developed soils (Figure 5:12, Table 7:7). The first of these soils is an Entisol formed in Unit VI (24 cm thick), with A-C horizonation and thicknesses and colors of (respectively): 6Ab (12 cm thick; 10YR 5/1), and 6C (10 cm thick; 10YR 7/3). Above this soil is an Entisol formed in Unit VII (30 cm thick), with A-A2-C horizonation and thicknesses and colors of (respectively): 5Ab (9 cm thick; 10YR 5/2), 5Ab2 (12 cm thick; 10YR 4/1), and 5C (9 cm thick; 10YR 6/3). Next upward is an Entisol formed in Unit VIII (9 cm thick), with A-C horizonation and thicknesses and colors of (respectively): 4Ab (6 cm thick; 10YR 3/2), and 4C (3 cm thick; 10YR 7/3).

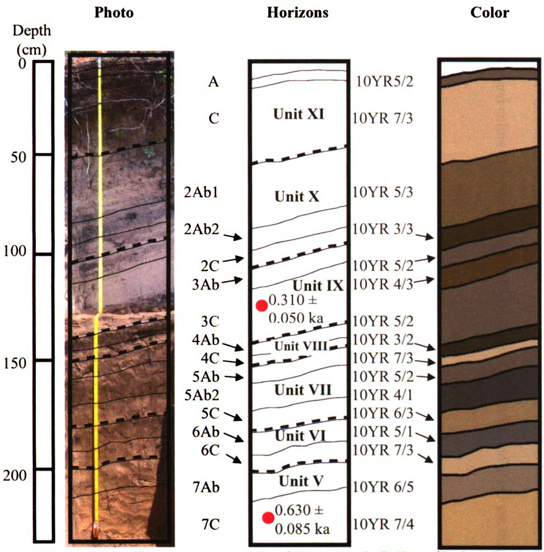

Figure 5:12 – The uppermost soils of Exposure 2 (Soil profile 2A). Dashed lines represent the boundary between stratigraphic units. These soils formed in fine eolian sand. Red dots are representative locations of OSL samples.

Above Unit VIII is Unit IX, which is comprised of 36 cm of fine sand. An OSL sample was collected in this unit at a depth of 1.15 m and provided a date of 0.31 ± 0.05 ka (UIC 2122). Formed in Unit IX is an Entisol with an A-C horizonation with thicknesses and colors of (respectively): 3Ab (8 cm thick, 10YR 4/3), and 3C (28 cm thick; 10YR 5/2). Next upward in the stratigraphic sequence is Unit X, which is comprised of 53 cm of fine sand. An Entisol has formed in this unit, and exhibits A1-A2-C horizonation and thicknesses and colors of (respectively): 2Ab1 (31 cm thick; 10YR 5/3), 2Ab2 (12 cm thick; 10YR 3/3), and 2C (10 cm thick; 10YR 5/2). Finally, the modern soil is an Entisol formed in Unit XI (50 cm thick), with an A-C horizonation and thicknesses and colors of (respectively): A (3 cm thick; 10YR 5/2), and C (47 cm thick; 10YR 7/3).

Exposure Three

Exposure 3 is located in the central part of the study area, approximately 150 m inland of the bluff exposures. This site is southwest of the high ridge which contains Exposure 2 (Figure 4:1). Given the inland location of Exposure 3, it could not be as accurately surveyed for elevation in reference to the lake. According to the USGS topographic map, the elevation at the top of Exposure 3 is approximately 295 masl, which is about the same as Exposure 2.

Visible in Exposure 3 are five stratigraphic units formed in eolian sand (Figure 5:13), which was 97.3% sand, with 2.0% silt and <1% clay. The breakdown of the sand fraction is: very coarse sand (<1%), coarse sand (3.5%), medium sand (43.6%), fine sand

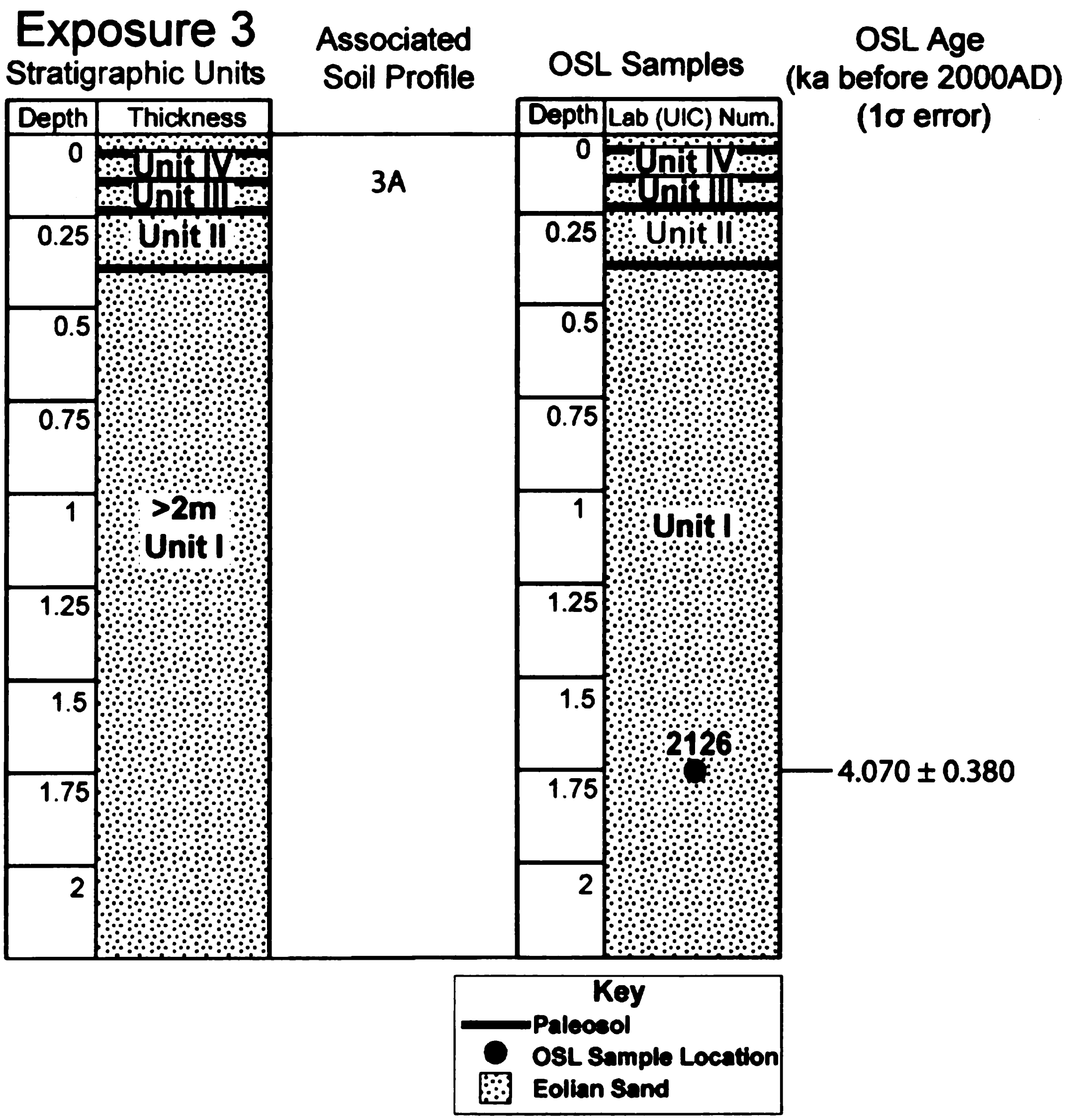

Figure 5:13 – Generalized stratigraphy of Exposure 3. Major eolian units and paleosols are shown in relation to their associated soil profile. The OSL sample location is labeled with UIC sample number and associated age. Depth and elevation in meters.

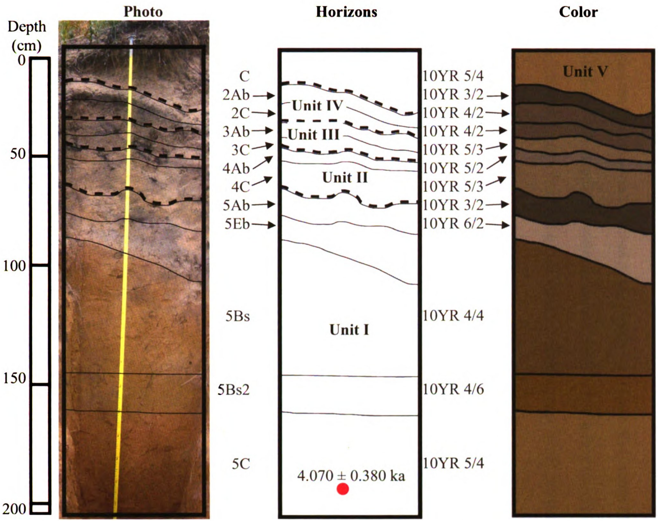

Figure 5:14 – The soils of Exposure 3 (Soil profile 3A). Dashed lines represent the boundary between stratigraphic units. They formed in fine eolian sand. The exposure is not extensive, and did not reveal any soils below this profile. Red dots are representative locations of OSL samples.

(44.8%), and very fine sand (5.4%). The exposure does not contain any lithologic discontinuities, as it outcrops in a blowout of an inland dune.

The basal unit (Unit I) consists of fine sand, and is greater than 1.39 meters thick. One OSL age was obtained from this unit. The sample was collected at a depth of 1.75 m and returned an age of 4.07 ± 0.38 ka (UIC 2126). Formed in the uppermost part of Unit I is a weakly developed Spodosol, which occurs at which at a depth of depth of 0.66 meters in the exposure. This soil was described in soil profile 3A (Figure 5:14, Table 7:8) and has A-E-Bs1-Bs2-C horizonation, with thicknesses and colors of (respectively): 5Ab (13 cm thick; 10YR 3/2), 5Eb (16 cm thick; 10YR 6/2), 5Bsb1 (20 cm thick; 10YR 4/4), 5Bsb2 (45 cm thick; 10YR 4/6), and 5C (>45 cm thick; 10YR 5/4).

Above Unit I is Unit II, a thin (19 cm) deposit of fine sand. Formed in this unit is an Entisol which occurs at a depth 0.47 m. The soil was described in soil profile 3A (Figure 5:14, Table 7:8) and exhibits A-C horizonation with thicknesses and colors of (respectively): 4Ab (5 cm thick; 10YR 4/2), and 4C (14 cm thick; 10YR 5/3).

That soil is covered by Unit III, a thin (14 cm) deposit of fine sand. Formed in this unit is an Entisol which occurs at a depth of 0.33 m. The soil was described in soil profile 3A (Figure 5:14, Table 7:8) and exhibits A-C horizonation with thicknesses and colors of (respectively): 3Ab (7 cm thick; 10YR 4/2), and 3C (7 cm thick; 10YR 5/3).

Overlying Unit III is Unit IV, a thin (17 cm) deposit of fine sand. Formed in this unit is an Entisol which occurs at a depth of 0.16 m in the exposure. The soil which was described in soil profile 3A (Figure 5:14, Table 7:8) and consists of A-C horizonation with thicknesses and colors of (respectively): 2Ab (7 cm thick; 10YR 3/2), and 2C (10 cm thick; 10YR 4/2).

Covering Unit IV is Unit V, which is comprised of 16 cm of fine sand. Formed at the top of this unit is an Entisol which occurs at the land surface (Figure 5:14). The soil was described in soil profile 3A (Figure 5:14, Table 7:8) and exhibits no discernable horizonation; rather one C horizon with a color of 10YR 5/4.

Discussion

This section discusses the results of this thesis and reconstructs the history of eolian sand deposition and dune growth at Arcadia Dunes. This reconstructed history is assessed in the context of the perched-dune model and compared with the results of other Lake Michigan coastal dune studies (e.g., Dow, 1937; Snyder, 1985; Arbogast et al., 2002a). Finally, the chapter concludes with a discussion of other possible explanations for sand dune mobilization at Arcadia Dunes.

Eolian Activity at Arcadia Dunes

This section summarizes the major periods of eolian activity at Arcadia Dunes as they are reconstructed from stratigraphic and OSL dating of sand bodies in the site exposures. The presence of paleosols was used as an indication of stable episodes when little or no eolian sand was deposited. It is assumed that little or no erosion of previously buried soils occurred, as no evidence of truncated stratigraphic units was observed. OSL samples were collected from pedogenically unaltered eolian sand units (C horizons) that were selected on the basis of their confidence for yielding accurate age estimates. When two OSL samples were collected from thick eolian units, the lower date is interpreted as the onset of deposition, whereas the upper date is interpreted to essentially represent the end of sand deposition for that unit.

Results of this study indicate that the lower ~ 75% of the lake-facing bluff exposures at Arcadia Dunes consist of undifferentiated glacial sediment. This sediment was most likely deposited during the Greatlakean advance at 14,000 cal. yr. B.P. (Larson and Schaetzl, 2001; Evenson et al., 1976; Dow, 1937). A similar stratigraphic setting has been described at other locations where high-perched dunes occur. In the first study of high-perched dunes in Michigan, Dow (1937) recognized that these dunes were perched upon tall coastal bluffs composed of glacial sediment of the Manistee Moraine. According to Snyder (1985), the bluff at Sleeping Bear dunes (approximately 40 km north-northwest of Arcadia Dunes) is also composed of glacial sediment of the Manistee Moraine deposited during the Greatlakean advance.

Evidence indicates that following deposition of the glacial sediment, the landscape at Arcadia Dunes was stable for a relatively long period of time, resulting in the development of soils with spodic characteristics (A-E-Bs-C horizonation). This period of stability has also been recognized nearby at Sleeping Bear dunes (Snyder, 1985) and apparently occurred during the Chippewa low stage of ancestral Lake Michigan, which began when the North Bay outlet was uncovered after the retreat of Greatlakean ice about 10,000 years ago (Larsen, 1987). Similarities in soil development between Arcadia and Sleeping Bear dunes during the Chippewa low phase suggest a stable period of consistent length between the sites. According to Snyder (1985), this period of stability occurred because the active lake shore was far to the west and this portion of the landscape was stabilized by vegetation.

This relationship between low lake levels and soil formation has also been observed along Lake Superior, where a similar period of stability occurred at the Grand Sable dune field. This interval resulted in the formation of the Sable Creek Soil in the uppermost part of Late-Wisconsin glacial sediments (Anderton and Loope, 1995). Arbogast (2000) also noted the presence of a paleosol formed at the top of glacial sediment at Nodaway Point along Lake Superior. These periods of stability along Lake Superior appear to have occurred during the Houghton low phase, which was roughly contemporaneous with the Chippewa low phase of Lake Michigan (e.g., Larson and Schaetzl, 2001).

OSL dating indicates that following the Chippewa interval, mobilization of eolian sand began at Arcadia Dunes and buried the glacial materials. The onset of this depositional interval did not result in immediate, massive dune construction. Rather the beginning of dune growth is marked by a thin deposit of eolian sand at Exposure 2, which was probably quickly deposited. After this episode, a period of stability occurred in which a weakly developed Entisol formed. Evidence from soil development during other studies (e.g., Loope and Arbogast, 2000; Arbogast et al., 2002a) suggests that this Entisol probably developed for ~ 150 years before it was buried by Unit III during the onset of major dune construction. The earliest OSL age was acquired from the base of this latter unit, and returned a date of approximately 4.5 ka. This exposure was active until ~ 4.34 ka (with a relatively large error; Figure 5:15), at which time the top of the unit was deposited and a soil began to form during a period of stability.

When the OSL ages are compared to Baedke and Thompson's (2000) lake-level curve (Figure 5:15), it appears that the activation of Exposure 2 correlates to the

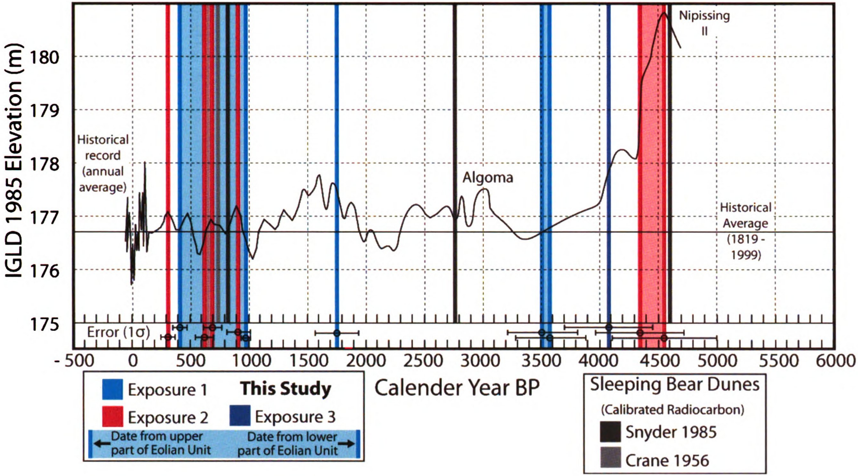

Figure 5:15 – Dates of eolian activity correlated to a lake-level curve (Modified from Baedke and Thompson, 2000). Dates from this study are OSL ages of eolian sand units. Also shown for reference are the calibrated radiocarbon ages from Sleeping Bear Dunes, as discussed in Snyder (1985).

Nipissing II high stand at ~ 4.5 ka and later. An episode of rapid dune construction began ~ 4.5 ka at Exposure 2 which resulted in the deposition of ~ 4.5 m of eolian sand by ~ 4.34 ka. Deposition of eolian sand apparently ended when lake levels fell following the Nipissing II phase. These results mirror those found by Snyder (1985) at Sleeping Bear dunes, where the burial of the basal paleosol was dated to 4,559 cal. yr. B.P. (Figure 5:15). Anderton and Loope (1995) and Arbogast (2000) found that Lake Superior's high-perched dunes activated early in the Nipissing high stand over 5,000 years ago. The presence of a thin eolian unit between the glacial sediment and the basal date of 4.5 ka at Arcadia Dunes indicates excellent correlation with the results from these other studies (Snyder, 1985; Anderton and Loope, 1995; Arbogast, 2000).

Initial growth of sand dunes during the Nipissing high stand has also been reported in low-perched dunes along much of Lake Michigan's southeastern shoreline. These low-perched dunes do not mantle sediment of glacial origin, but rather cover topographically low lake plains formed during early proglacial lake phases (e.g., Arbogast and Loope, 1999; Arbogast et al., 2002). Arbogast and Loope (1999) dated a paleosol formed in the basal eolian unit at Nordhouse Dunes, and estimated that construction began during the Nipissing high stand between 4,900 and 4,500 cal. yr. B.P. The initial construction of Lake Michigan's low-perched dunes during the Nipissing high stand has also been reported by Arbogast et al. (2002), Van Oort at al. (2001), Hansen et al. (2002), Lepczyk and Arbogast (2005), and Fisher and Loope (2005).

As the growth of dunes at Arcadia progressed in its initial phases, eolian sand apparently spread inland at least 150 m. This conclusion is supported by an OSL age of ~ 4.07 ka from the dune ridge which contains Exposure 3. This period of mobilization has not been commonly reported in previous studies of Lake Michigan coastal dunes. Deposition of the dune in which Exposure 3 crops out probably lagged behind the onset of construction during the Nipissing, as it is located in the limb of a parabolic dune. The dune may have originated as a blowout in the lake-ward dune ridge, and then propagated inland ~ 150 m as a parabolic dune by ~4.07 ka.

It is unknown if Exposure 1 was active as early as Exposures 2 and 3, but the two thin units (II and III) just above the basal paleosol suggest two short periods of eolian activity at this exposure that probably occurred at about the same time as the initial deposition of sand at Exposure 2. These earliest depositional episodes at Exposure 1 were punctuated by two periods of stability during which a soil formed, the first being an Entisol with spodic characteristics and the other a weakly developed Entisol. The earliest OSL evidence of activity at Exposure 1 is recorded in Unit IV, from ~ 3.53 to ~ 3.49 ka. Interestingly, this general period of activity correlates with a regression in Lake Michigan that followed the Nipissing high stand (Baedke and Thompson, 2000) and thus does not fit within the perched-dune model (Dow, 1937; Snyder, 1985; Anderton and Loope, 1995). However, many other studies on both high and low-perched dunes on the west coast of Lower Michigan have found evidence of dune construction at this time, as near Holland (Arbogast et al., 2002a), South Haven (Van Oort et al., 2001) and Empire (Arbogast and Loope, 2000). Eolian activity at high-perched dunes along Lake Superior was also found between 3,800 and 3,000 years ago (Anderton and Loope, 1995; Arbogast, 2000). These ages suggest that sand can be supplied to high-perched dunes even during lowering lake phases.

Following the period of dune growth at ~ 3.5 ka at Exposure 1, the landscape here stabilized sufficiently for a well-developed Entisol with spodic characteristics to form. This soil apparently formed over a period of ~ 1,700 years, as it was buried by Unit V during the next period of activity at this site ~ 1.76 ka. The formation of this soil, coupled with the absence of eolian units between the dates of ~ 3.49 and ~ 1.76 ka at any of the exposures, suggests that little or no eolian activity occurred at Arcadia Dunes during the Algoma high stand from 3,200 – 2,300 cal. yr. B.P. (Baedke and Thompson, 2000). The lack of dune construction at this time seems odd because intensified bluff erosion was presumably occurring (e.g., the perched-dune model). The paucity in eolian activity at this time may have been caused by climate changes, such that prevailing winds or storm tracks were not conducive to dune erosion. It is also possible that foredunes had formed during the post-Nipissing regression and prevented erosion of coastal bluffs during the Algoma high stand.

Based on data from OSL ages, the majority of eolian activity at Arcadia Dunes has apparently occurred within the last 1,000 years. Eolian sand was supplied at Exposure 1 again at approximately 0.97 ka, with the onset of deposition of Unit VI. Exposure 1 remained active while the thin Unit IV of Exposure 2 was deposited for about 0.91 ka. Deposition of Unit IV then ceased at this site, and Exposure 2 remained stable long enough for an Entisol to form, which was then buried at ~ 0.71 ka by the 18 meters of sand comprising Unit V. This was a large and rapid period of dune building which lasted until ~ 0.63 ka when deposition of Unit V ceased. Apparently Unit VI of Exposure 1 continued to slowly thicken during these periods of activity at Exposure 2, and sand deposition did not cease until ~ 0.4 ka. In summary, within the last 1,000 years Exposure

1 activated, during which time Exposure 2 activated, stabilized to form a soil, reactivated to deposit 18 meters of sand, and again stabilized approximately 200 years before Exposure 1.

Exposure 1 shows little evidence of recent eolian activity, except for a thin mantle of bare sand above the uppermost paleosol. The paleosol has an overthickened A horizon which suggests it may be cumulic, forming concurrently with a slow influx of eolian sand. Exposure 2 shows evidence of many eolian events more recent than ~ 0.63 ka. An OSL sample within the seven upper paleosols dated to ~ 0.31 ka, which is buried by two thin sand units.

Widespread dune construction and mobility within the last 1,000 years is common among most dunes along the Lake Michigan shoreline (e.g., Loope and Arbogast, 2000; Arbogast et al., 2002; Van Oort et al., 2001; Hansen et al., 2002; Lepczyk and Arbogast, 2005; Fisher and Loope, 2005). Dune growth at 1.0 ka may be related to a period of widespread drought in the region which occurred at that time (Booth et al., 2006) and will be discussed later. Coastal dune studies have found that it is more likely that dune growth is related to lake-level cycles. Most studies have found that recent periods of considerable dune growth correlate to a ~ 150-year cycle in dune building (e.g., Loope and Arbogast, 2000), caused by fluctuating lake levels (e.g., Thompson, 1992; Thompson and Baedke, 1995; Thompson and Baedke, 1997). OSL samples at Arcadia Dunes did not resolve the eolian activity here at such a high temporal resolution. However, the youngest OSL date of ~ 0.31 ka was acquired from a unit of eolian sand buried by two units in which Entisols are formed. These soils, along with the other recent paleosols or cumulic A horizons at the study area, may correlate to this 150-year cycle.

Geomorphic evidence indicates that the dunes at Arcadia continue to be active, as several blowouts and active dunes (e.g., Old Baldy) are present at the study area. These features tend to be oriented to the southwest, which suggests that the most important mobilizing winds at Arcadia are from that direction. This pattern suggests that the current wind and storm regime (e.g., Delcourt et al., 2002) is sufficient for these dunes to remain active and migrate, and that some factor other than lake-level fluctuations may have caused their initiation. It is difficult to associate these features with lake level events, since they are topographically removed from any immediate effects of wave erosion.

Discrepancies in the Timing of Eolian Activity

The hypothesis of this thesis is that deposition of eolian sand at Arcadia Dunes is linked with periods of high lake levels as per the perched-dune model (Dow, 1937; Anderton and Loope, 1995). Although general agreement exists between the lake-level curve and dune construction at Arcadia Dunes, several periods of dune activity and/or stability occurred that are not readily explainable by the perched-dune model. This section discusses some of the other explanations for the periods of eolian activity at Arcadia Dunes.

I am confident that the OSL ages in this study are correctly interpreted, because of several age controls. First, all of the ages are in correct stratigraphic order, with progressively younger ages toward the top of exposures. An excellent example of this order can be seen in Unit IV at Exposure 1. This unit is 12 m thick and provided an OSL ages of 3.53 ka and 3.49 ka from the bottom and top of the unit, respectively. In addition to this correct age/stratigraphic ordering, none of the dates are unexpectedly old or young in reference to the relative soil development of the associated paleosols. Most importantly, dates from within the same eolian unit were very similar (within 1σ error of each other) or dated correctly in terms of superposition.

There are, however, relatively large standard deviations (300-400 years) on samples returning ages of several thousand years. This is due, in part, to the age of the samples, as the accumulation of electrons asymptotically reduces due to partial saturation of the electron traps (Duller, 2004). Generally this is corrected for by during the calculating a regeneration curve. Other possible explanations include incomplete reset of the electron traps, anomalous fading (Duller, 2004), and the dampening effect of moisture on the dose rate. Because of these issues, it is understood that older ages estimate a broader possible range of periods of eolian activity.

Assuming that the OSL dates are reasonably accurate, Exposure 1 was active from about 3.53 to 3.49 ka. The lake-level curve (Figure 5:15; Baedke and Thompson, 2000) indicates that this is a period of lowering lake levels after the Nipissing high stand. According to the perched-dune model, Arcadia Dunes should have been stable during this time. Apparently portions of the dune field were active, along with several other dune fields across the region, along both Lakes Michigan (Arbogast and Loope, 1999; Arbogast et al., 2002; Van Oort et al., 2001; Arbogast and Loope, 2000) and Superior (Anderton and Loope, 1995; Arbogast, 2000).

Explanations of perched dune activity unrelated to lake levels are varied. Dow (1937) hypothesized that, given time, perched dunes may be supplied with a considerable amount of sand from beaches. His observations indicated that the majority of sand being supplied to the perched dunes was from glacial material. However it is conceivable that widening beaches formed during the post-Nipissing regression may have been able to supply a substantial amount of sand to the perched dunes at approximately ~3.5 ka (Dow, 1937). Therefore, the perched dunes may be *composite* in the context of the Dow (1937) model, in that they are supplied with sand by varying sources.

It is also possible that regional climate changes had some impact on the evolution of coastal dune systems at Arcadia. Such changes have been documented with respect to dunes in the interior of Michigan. For example, inland dunes of Michigan's Upper Peninsula were active between ~ 7.0 and 5.5 ka (Arbogast et al., 2002b; Arbogast and Packman, 2004). This time period correlates to the Altithermal (hypsithermal) climate interval, which was a warmer and drier period throughout the region (Delcourt et al., 2002). During this time lake levels fell and soil moisture decreased (Davies et al., 2000), which may have lead to increased eolian activity. Arbogast et al. (2002b) did not believe that the direct impact of the Altithermal was enough to stimulate activity in Michigan's interior. Rather, they associated activation to a simultaneous drop in groundwater occurring in conjunction to a slight increase in arid conditions. Interior dunes of the Upper Peninsula may have then stabilized in response to a rise in the water table at approximately 5.5 ka. The high elevations, sandy sediments, and lack of redoximorphic features at Arcadia Dunes make this explanation unlikely.

The Midwest became cool and moist beginning at about 5,300 cal. yr. B.P. (Delcourt et al., 2002). The influx of moisture during this time period may have been partly responsible for the rising lake levels of the Nipissing transgression (Booth et al., 2002), which in turn appears to have stimulated dune activity along the Great Lakes shorelines. This period was apparently interrupted by a short-lived but intense regional drought that occurred about 4,200 cal. yr. B.P. (Booth et. al., 2006), which resulted in changes in vegetative cover, as well as an increase in forest fire and interior dune activity throughout the region. It is possible that this drought contributed to the mobilization of eolian sand at Arcadia from about 4.5 to 3.5 ka.

By 3,000 cal. yr. B.P. the region had returned to a general trend of cool and moist conditions. Since 3,000 cal. yr. B.P. lake-effect precipitation has slowly increased due to a southward shift in the polar jet stream during the winter (Delcourt et al., 2002). This time frame coincides with the onset of a relatively long period of dune stability at Arcadia. Frequent periods of dune activity apparently returned approximately 1,000 years ago and coincided with the Medieval Warm Period (MWP), which lasted from about 800 to 1,300 A.D. (Booth et al., 2006; Bradley et al., 2003). Droughts associated with the MWP were the main factor responsible for increased wildfires and dune activation throughout the Midwest and Great Plains of North America (Booth et al., 2006; Foreman et al., 2001).

Following the MWP the region became cooler and wetter during the Little Ice Age (LIA), which occurred from ~1300 to 1900 A.D. (Soon and Baliunas, 2003; Crowley, 2000). This shift in climate has been attributed to a retraction of the polar jet stream (Soon and Baliunas, 2003), and presumably reduced droughts and wildfires in the Great Lakes region. Temperatures then increased in the 20th century, and projections for accelerated warming in the future (Crowley, 2000) may cause added stress to dune systems throughout the mid-continent. These shifting periods of drought during the Holocene may have affected sand dunes throughout the region (Booth et al., 2006), including those at Arcadia. However, the direct impact of large-scale climate changes on coastal dunes is largely unstudied.

Intense wind is another possible explanation of eolian activity during relatively low lake phases. The transportation of dune sand in Michigan's coastal dunes is enhanced during the late fall and winter, when conditions are drier and vegetative cover is reduced (Van Dijk, 2004). This timing leaves the dunes particularly susceptible to intense winds which can potentially increase wave erosion, and further compromise the vegetative cover. Arbogast (2000) cited strong coastal winds as a possible reason that dunes at Nodaway Point were slow to stabilize following high lake levels. He argued that intense winds may have re-worked the dune sand, slowing the recovery of vegetation and subsequent stabilization of the dune. Even though wind and storm tracks are related to large-scale climate factors (e.g., Delcourt et al., 2002), the effect of any given wind or storm regime is ultimately an unpredictable site-specific factor. Arcadia Dunes, along with all coastal dunes, are vulnerable to unpredictable and highly localized meteorological conditions.

All of the aforementioned causes of eolian activity probably work in conjunction with site-specific factors (slope, aspect, vegetation, etc.) and complicate the response of coastal dunes to lake-level fluctuations (Sommerset et al., 2007; Bailey et al, 2001; Murray and Clemmensen, 2001). The activation of Arcadia Dunes during relatively low lake levels indicates that, as other researchers have found, their activation may be a result of localized variables that work in conjunction with fluctuating lake levels (Arbogast, 2000; Wilson and Braley, 1997). Perhaps the effects of regional droughts (e.g., Booth et al., 2006) may have a more significant direct impact on vegetation cover and associated sand mobilization than previously recognized.

CHAPTER 6

SUMMARY AND CONCLUSIONS

The goal of this study was to reconstruct the evolution of the Arcadia Dune field and to test the validity of the perched-dune model using optically stimulated luminescence (OSL) dating. Previous investigators (e.g., Anderton and Loope, 1995; Arbogast and Loope, 1999, Loope and Arbogast, 2000, Arbogast et al., 2002) have employed stratigraphic analysis and near exclusive use of radiocarbon (^{14}C) dating to interpret the depositional history of coastal dunes along the Great Lakes. The intent of these studies has been to relate dune activity to paleoclimatic conditions, specifically lake-level fluctuations. However, interpretation of eolian activity using radiocarbon analysis may lead to erroneous conclusions due to uncertainties with the method.

It is useful to place this study within the context of early assumptions about the evolution of coastal sand dunes along Lake Michigan. For many years the prevailing thinking was that that massive coastal dunes formed during the Nipissing high stand (Dorr and Eschman, 1970; Buckler, 1979), and were supplied by sand from wide beaches (Olson, 1958a, b). However, studies focusing on erosion and sediment supply of high-perched dunes proposed that they may actually activate when lake levels are high such that waves undercut coastal bluffs which, in turn, allow sand to mobilize (e.g. Dow, 1938; Marsh and Marsh, 1987). The application of dating technology to reconstruct lake levels (e.g., Thompson, 1992; Thompson and Baedke, 1995; Thompson and Baedke, 1997; Baedke and Thompson, 2000) has allowed researchers to test this dune model (Anderton and Loope, 1995; Arbogast and Loope, 2000). However, the uncertainties of relating radiocarbon dates of paleosols to periods of eolian deposition may have lead to flawed interpretations as to the precise relationship between lake levels and dune activity (e.g., Snyder, 1985).

In the context of reconstructing the geomorphic history of Arcadia Dunes, I sought to test the perched-dune model, specifically that eolian sand was largely supplied to the dunes high lake levels. I tested this hypothesis by dating a variety of eolian sand units with OSL; this method estimates the time of sand deposition. These dates were subsequently compared to established late-Holocene lake-level estimations for Lake Michigan (e.g., Thompson, 1992; Thompson and Baedke, 1995; Thompson and Baedke, 1997; Baedke and Thompson, 2000).

Results indicate that, following the retreat of Greatlakean ice, a period of stability occurred during which a well-developed paleosol formed in the glacial sediment. This basal paleosol was buried by dune sand approximately 4.5 ka during the onset of eolian sand deposition. Growth of the Arcadia dunes continued episodically with major periods of deposition at ~3.5, 1.7, 1-0.4, 0.9, 0.7 and 0.3 ka and exhibited high chronological variability between exposures. Most of the sand deposition occurred within the last 1,000 years. These periods of dune construction frequently correlated to high lake levels, although a noticeable exception occurred approximately 3.5 ka during the post-Nipissing regression. This period of dune construction suggests that Dow's (1937) hypothesis of sand being supplied to perched dunes from beaches may be viable under certain conditions.

In conclusion, OSL dates from this study show a mixed relationship between high lake levels and eolian activity at Arcadia Dunes. Results indicate that dune construction began during the Nipissing high stand. In general, the sand dunes were active when Lake Michigan was at high or rapidly fluctuating lake levels, supporting the hypothesis that the Arcadia Dunes were active during high lake phases. However, a major period of activity occurred approximately 3.5 ka which correlates to a distinct lake regression following the Nipissing, possibly supplied with sand from wide beaches. Conversely, the sand dunes were not always active when lake levels were high. These discrepancies suggest that the sand dunes are complex, and may have been responding unexpectedly to some other climate and site-specific variables.

Contributions of this Study

This study is the first to exclusively use OSL dating to reconstruct the geomorphic history of a coastal sand dune along Lake Michigan and to test the perched-dune model. It provides an excellent suite of dates which add to the rapidly growing literature of dune chronology throughout the region. This study is also the first to focus solely on establishing a detailed chronology of eolian activity on high-perched dunes in northwest Lower Michigan, which contains some of Michigan's most high profile dune fields (Snyder, 1985). It is hoped that this study will stimulate additional research at Arcadia Dunes and surrounding high-perched dune fields.

Further Research

Interpretations of radiometric dates would be aided by continued research into lake-level fluctuations and their causes. This will help identify additional elements that should be incorporated into the perched-dune model. It may also highlight unusual periods of eolian activity that may be tied to short, but intense lake-level fluctuations which were not recorded in lake-level estimations. Such a brief time scale for lake-level fluctuations would result in high uncertainty using radiocarbon dating of ridge/swale complexes, even if they are large enough to cause massive dune construction.

Region-wide OSL dating of coastal sand dunes should continue in conjunction with improvements in OSL methodology and sample collection techniques. No one study area provides a complete picture of activity along Lake Michigan, so each individual location is important in the formation of a region-wide suite of dates. Large exposures containing complex stratigraphy of eolian units should be closely studied and dated to construct a rigorous model of sand mobilization.

Furthermore, the apparent mobilization of Arcadia Dunes during a lake regression indicates that high-perched sand dunes may activate due to other causes. Quantitative studies of sand transportation on perched dunes may clarify the source of dune sand (e.g., Dow, 1937). Future studies should also focus on identifying regional climatic factors, such as precipitation, wind, etc., as well as studying local factors such as soil moisture, topography, and fire regime. It is likely that these factors work in concert with lake-level fluctuations.

CHAPTER 7 – APPENDIX

SOIL PROFILE DATA TABLES

Table 7:1 – Soil profile 1D description.

Horizon	Depth(cm)	Thickness(cm)	Color	Boundary		Structure	
5Ab	0	5	10YR5/4	**Distinctness**	Abrupt	**Type**	Granular
				Topography	Smooth	**Grade**	Weak
5C	5	5	10YR6/4	**Distinctness**	Abrupt	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
6Ab	10	6	10YR4/3	**Distinctness**	Abrupt	**Type**	Subangular blocky
				Topography	Smooth	**Grade**	Weak
6Eb	16	7	10YR5/3	**Distinctness**	Abrupt	**Type**	Granular
				Topography	Diffuse	**Grade**	Weak
6Bwb	23	10	10YR4/3	**Distinctness**	Abrupt	**Type**	Subangular blocky
				Topography	Diffuse	**Grade**	Weak
7Ab	33	4	10YR3/3	**Distinctness**	Gradual	**Type**	Subangular blocky
				Topography	Diffuse	**Grade**	Weak
7Eb	37	20	10YR5/4	**Distinctness**	Abrupt	**Type**	Granular
				Topography	Smooth	**Grade**	Weak
7Bsb	57	43	7.5YR4/6	**Distinctness**	Gradual	**Type**	Granular
				Topography	Smooth	**Grade**	Weak
7C	100	20+	10YR6/4	**Distinctness**	Gradual	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A

Table 7:2 – Soil profile 1C description.

Horizon	Depth(cm)	Thickness(cm)	Color	Boundary		Structure	
3C	0	7	10YR5/4	**Distinctness**	Abrupt	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
4Ab	7	10	10YR3/4	**Distinctness**	Abrupt	**Type**	Subangular blocky
				Topography	Smooth	**Grade**	Weak
4Eb	17	29	10YR6/4	**Distinctness**	Clear	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
4Bsb	46	30	10YR5/6	**Distinctness**	Clear	**Type**	Single Grain
				Topography	Wavy	**Grade**	N/A
4C	76	25+	10YR6/5	**Distinctness**	Diffuse	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A

Table 7:3 – Soil profile 1B description.

Horizon	Depth(cm)	Thickness(cm)	Color	Boundary		Structure	
3Ab1	0	7	5YR2.5/1	**Distinctness**	Abrupt	**Type**	Subangular blocky
				Topography	Smooth	**Grade**	Moderate
3Ab2	7	4	7.5YR3/2	**Distinctness**	Abrupt	**Type**	Subangular blocky
				Topography	Smooth	**Grade**	Weak
3C	11	70+	10YR5/4	**Distinctness**	Abrupt	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A

Table 7:4 – Soil profile 1A description.

Horizon	Depth(cm)	Thickness(cm)	Color	Boundary		Structure	
C	0	22	10YR4/3	**Distinctness**	N/A	**Type**	Single Grain
				Topography	N/A	**Grade**	N/A
2Ab	22	19	10YR3/1	**Distinctness**	Abrupt	**Type**	Subangular blocky
				Topography	Smooth	**Grade**	Weak
2Eb	41	6	10YR6/2	**Distinctness**	Clear	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
2Bsb	47	7	10YR6/3	**Distinctness**	Clear	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
2C	54	48+	10YR7/2	**Distinctness**	Gradual	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A

Table 7:5 – Soil profile 2C description.

Horizon	Depth(cm)	Thickness(cm)	Color	Boundary		Structure	
9C	0	8	10YR5/4	**Distinctness**	N/A	**Type**	Subangular blocky
				Topography	N/A	**Grade**	Weak
10Ab	8	3	10YR4/3	**Distinctness**	Abrupt	**Type**	Subangular blocky
				Topography	Smooth	**Grade**	Weak
10C	11	5	10YR5/4	**Distinctness**	Abrupt	**Type**	Subangular blocky
				Topography	Smooth	**Grade**	Weak
11Ab	16	4	10YR3/2	**Distinctness**	Abrupt	**Type**	Subangular blocky
				Topography	Diffuse	**Grade**	Weak
11Eb	20	15	10YR5/4	**Distinctness**	Abrupt	**Type**	Single Grain
				Topography	Diffuse	**Grade**	N/A
11Bsb	35	70	7.5YR4/6	**Distinctness**	Gradual	**Type**	Single Grain
				Topography	Diffuse	**Grade**	N/A
11C	105	10+	7.5YR4/6	**Distinctness**	Gradual	**Type**	Single Grain
				Topography	Diffuse	**Grade**	N/A

Table 7:6 – Soil profile 2B description.

Horizon	Depth(cm)	Thickness(cm)	Color	Boundary		Structure	
8Ab	0	8	10YR2/1	**Distinctness**	Clear	**Type**	Subangular blocky
				Topography	Smooth	**Grade**	Weak
8BCb1	8	22	10YR5/4	**Distinctness**	Diffuse	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
8BCb2	30	27	10YR6/4	**Distinctness**	Diffuse	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
9Ab	57	11	10YR2/2	**Distinctness**	Clear	**Type**	Subangular blocky
				Topography	Wavy	**Grade**	Weak
9Eb	68	19	10YR7/4	**Distinctness**	Diffuse	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
9Bsb	87	48	10YR5/8	**Distinctness**	Diffuse	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
9C	135	20+	10YR6/6	**Distinctness**	Diffuse	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A

Table 7:7 – Soil profile 2A description.

Horizon	Depth(cm)	Thickness(cm)	Color	Boundary		Structure	
A	0	3	10YR5/2	**Distinctness**	N/A	**Type**	Granular
				Topography	N/A	**Grade**	N/A
C	3	47	10YR7/3	**Distinctness**	Gradual	**Type**	Granular
				Topography	Smooth	**Grade**	N/A
2Ab1	50	31	10YR5/3	**Distinctness**	Diffuse	**Type**	Subangular blocky
				Topography	Broken	**Grade**	Weak
2Ab2	81	12	10YR3/3	**Distinctness**	Clear	**Type**	Subangular blocky
				Topography	Smooth	**Grade**	Weak
2C	93	10	10YR5/2	**Distinctness**	Clear	**Type**	Granular
				Topography	Smooth	**Grade**	N/A
3Ab	103	8	10YR4/3	**Distinctness**	Clear	**Type**	Granular
				Topography	Smooth	**Grade**	N/A
3C	111	28	10YR5/2	**Distinctness**	Clear	**Type**	Granular
				Topography	Smooth	**Grade**	N/A
4Ab	139	6	10YR3/2	**Distinctness**	Clear	**Type**	Subangular blocky
				Topography	Smooth	**Grade**	Weak
4C	145	3	10YR7/3	**Distinctness**	Clear	**Type**	Granular
				Topography	Smooth	**Grade**	N/A
5Ab1	148	9	10YR5/2	**Distinctness**	Clear	**Type**	Subangular blocky
				Topography	Smooth	**Grade**	Weak
5Ab2	157	12	10YR4/1	**Distinctness**	Gradual	**Type**	Subangular blocky
				Topography	Diffuse	**Grade**	Weak
5C	169	9	10YR6/3	**Distinctness**	Clear	**Type**	Subangular blocky
				Topography	Diffuse	**Grade**	Weak
6A	178	12	10YR5/1	**Distinctness**	Clear	**Type**	Subangular blocky
				Topography	Diffuse	**Grade**	Weak
6C	190	10	10YR7/3	**Distinctness**	Clear	**Type**	Granular
				Topography	Diffuse	**Grade**	N/A
7A	200	12	10YR6/5	**Distinctness**	Gradual	**Type**	Subangular blocky
				Topography	Diffuse	**Grade**	Weak
7C	212	5+	10YR7/4	**Distinctness**	Gradual	**Type**	Granular
				Topography	Diffuse	**Grade**	N/A

Table 7:8 – Soil profile 3A description.

Horizon	Depth(cm)	Thickness(cm)	Color	Boundary		Structure	
C	0	16	10YR5/4	**Distinctness**	N/A	**Type**	Single Grain
				Topography	N/A	**Grade**	N/A
2Ab	16	7	10YR3/2	**Distinctness**	Abrupt	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
2C	23	10	10YR4/2	**Distinctness**	Clear	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
3Ab	33	7	10YR4/2	**Distinctness**	Clear	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
3C	40	7	10YR5/3	**Distinctness**	Clear	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
4Ab	47	5	10YR5/2	**Distinctness**	Clear	**Type**	Subangular Blocky
				Topography	Smooth	**Grade**	Weak
4C	52	14	10YR5/3	**Distinctness**	Clear	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
5Ab	66	13	10YR3/2	**Distinctness**	Clear	**Type**	Single Grain
				Topography	Wavy	**Grade**	N/A
5Eb	79	16	10YR6/2	**Distinctness**	Abrupt	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
5Bsb1	95	20	10YR4/4	**Distinctness**	Gradual	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
5Bsb2	115	45	10YR4/6	**Distinctness**	Clear	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A
5C	160	45+	10YR5/4	**Distinctness**	Gradual	**Type**	Single Grain
				Topography	Smooth	**Grade**	N/A

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