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A PRELIMINARY RECORD OF HOLOCENE EOLIAN ACTIVITY
IN MICHIGAN"S EASTERN UPPER PENINSULA

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Scott D. Crozier

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A PRELIMINARY RECORD OF HOLOCENE EOLIAN ACTIVITY IN MICHIGAN'S EASTERN UPPER PENINSULA

By

Scott David Crozier

A THESIS

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ABSTRACT

A PRELIMINARY RECORD OF HOLOCENE EOLIAN ACTIVITY IN MICHIGAN'S EASTERN UPPER PENINSULA

By

Scott David Crozier

An extensive dune field exists in the east-central upper peninsula of Michigan, inland (>20 km) from the modern Great Lakes shorezone. The dunes are parabolic, have northwesterly-orientated limbs, and overlie post-glacial lacustrine deposits approximately 10,000 years old. In an effort to estimate when sand mobilization occurred, two radiocarbon ages were derived from organic-rich deposits at the base of dunes. A subsequent examination of 10 surficial soils (Entic and Typic Haplorthods) was conducted along a 62 km transect joining the sites, to determine whether relative differences exist in soil development.

Results suggest that dunes mobilized well after deglaciation. The Seney Sand Pit, located at the southwest end of the transect, yielded a maximum-limiting estimate of 4940 $\pm$ 60 yrs B.P. This age implies eolian mobilization occurred during the middle Holocene, a period correlative with regional warming in the continental interior. At the other end of the transect, near Tahquamenon Falls, a radiocarbon date of 710 $\pm$ 100 yrs B.P. on similar materials suggests more recent mobilization, although this date may be unreliable. An exploratory examination of surface soils indicates that development varies along the transect, suggesting that dunes may have episodically mobilized in time and space.

ACKNOWLEDGEMENTS

If I have learned nothing else from this endeavor, I now realize that a thesis can not be completed alone. So many talented individuals have contributed to this work. Specifically, I would like to thank the Department of Geography for broadening my understanding and appreciation of the discipline and for providing me with lifelong friendships. I would also like to recognize the Michigan Soils Digitizing Team, who unselfishly offered the time and technical resources to make much of this thesis possible. Lastly, I reserve my utmost gratitude to my family for their constant encouragement and kind words. Thank you all for being a part of this study and a part of my life.

This thesis is dedicated in loving memory of Verna Tremont, my late grandmother and forever my best friend.

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CHAPTER I

INTRODUCTION AND PROBLEM DEFINITION

Landscapes composed of unconsolidated sand are susceptible to mobilization if vegetation is reduced or strong winds persist (McKee, 1979). Due to their sensitivity to climatic fluctuations, sand dunes can provide significant information about paleoenvironmental change. Extensive deposits of eolian sand occur in northeastern and central North America. Generally, forests mantle dunes in mesic and/or subarctic regions (e.g., Grigal *et al.*, 1976; Filion, 1987; Keen and Shane, 1990; Thorson and Schile, 1995; Arbogast *et al.*, 1997) while deposits in the semi-arid, Great Plains are anchored by grasses (e.g., Ahlbrandt *et al.*, 1983; Muhs, 1985; 1991; Forman and Maat, 1990; Madole, 1995; Arbogast, 1996). Although sand dunes and sand sheets within these regions are largely stable at present, geomorphic research indicates episodic and often extensive mobilization during the late-Quaternary. In this context, sand dunes currently anchored by forest are of particular interest because they presumably require significant environmental fluctuations for mobilization to occur.

Research conducted in forested regions of northeastern North America (e.g., Keen and Shane, 1990; Filion *et al.*, 1991; Thorson and Schile, 1995) has demonstrated conditions required to mobilize forested dunes (Figure 1:1). Results indicate that no unifying variable exists for the cause or timing of eolian sand mobilization. Rather, dune mobilization has occurred due to a variety of localized paleoenvironmental conditions.

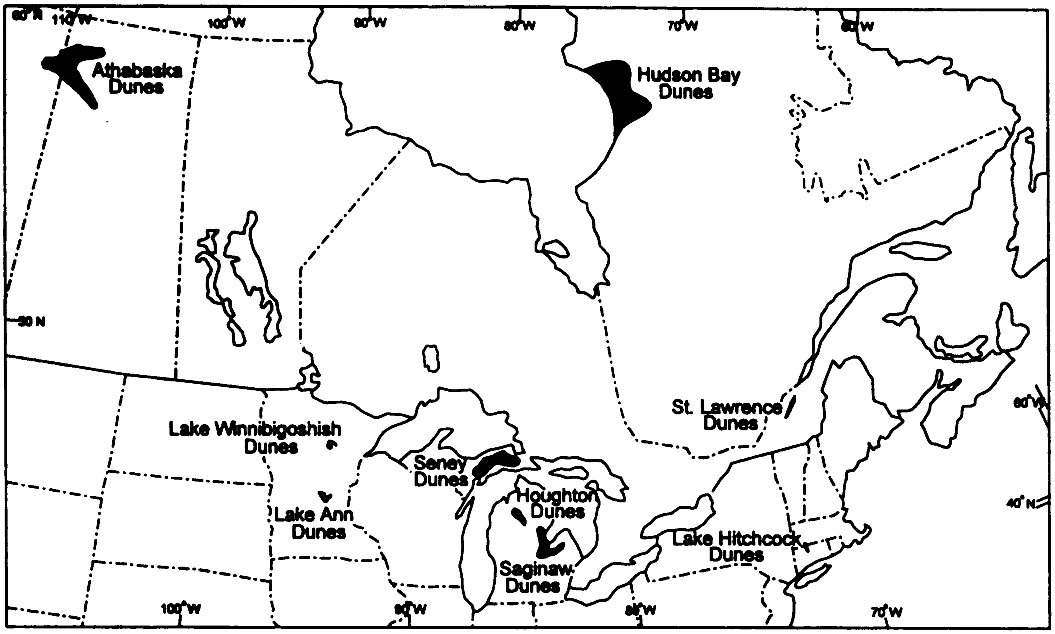


Figure 1:1 - Location of forested dune study areas within northeastern North America. Compiled from David (1981), Filion et al. (1991), Filion (1987), Thorson and Schile (1995), Keen and Shane (1990), Grigal et al. (1976) and Farrand and Bell (1982).

Broadly stated, the primary factors responsible for dune formation are: 1) deflation of newly deglaciated or subaerial surfaces, 2) devegetation aided by Holocene climate fluctuations, 3) destabilization as a result of increased wildfire frequencies, or 4) a combination thereof.

Eolian mobilization is common in deglacial environments as loose sediments are not yet stabilized by vegetation. Research conducted in Connecticut indicates dune development was aided by deglaciation. Thorson and Schile (1995) used dune orientation and sedimentary structures to suggest that mobilization initially occurred in newly deglaciated or subaerial sediments between 12,700 and 12,400 yrs B.P. Dunes were formed by predominantly northeasterly winds generated from a hypothesized glacial anticyclone.

Further evidence for deglacial dune formation was reported in the central St. Lawrence lowland of Quebec (Filion, 1987). Radiocarbon age determinations indicated parabolic dunes then had formed from about 10,000 to 7500 yrs B.P. The northeasterly orientation of dunes is consistent with those in Connecticut and stabilization corresponds with the collapse of the theorized anticyclone. Additional research, undertaken in northern Saskatchewan (David, 1981) used dune morphology and orientation to suggest large-scale dune mobilization initiated approximately 10,000 yrs B.P. and ceased about 8800 yrs B.P. David attributed dune activity to localized ice front oscillations and a theorized glacial anticyclone.

In contrast to deglacial eolian environments, Holocene climatic fluctuations have also been linked to the destabilization of forested dunes. For example, research conducted in north-central Minnesota suggests peak mobilization of dunes between about 8000 and

5000 yrs B.P. (Grigal *et al.*, 1976). This interval correlates favorably with the Altithermal, a warm/dry episode well-documented throughout the central United States (e.g., Wright, 1976; Baker *et al.*, 1992; Bradbury *et al.*, 1993). Subsequently, Keen and Shane (1990) conducted research in east-central Minnesota and reported eolian activity between about 8000 and 5000 yrs B.P., when Altithermal conditions induced both northwest winds and a localized shift from mixed coniferous/deciduous forest vegetation to a grass-dominated landscape.

The youngest dunes documented in northeastern North America are closely related to increased fire frequencies. Filion *et al.* (1991) utilized radiocarbon ages along a south-north transect to identify temporal and spatial mobilization patterns in northern Quebec, Canada. Results suggested that dunes are Late Holocene in age, with major periods of mobilization at 3650-2750 and after 700 yrs B.P. Given that a detailed fire chronology (Payette and Gagnon, 1985) correlates well with dune mobilization intervals, the authors concluded that catastrophic wildfires triggered widespread dune activation in northern Quebec.

While eolian research has been conducted in significant portions of northeastern North America, sizable spatial gaps remain in the paleoenvironmental record. For example, a large field of forested dunes exists in the eastern upper peninsula of Michigan that has yet to be investigated. These dunes are parabolic and occur mostly in swampy landscapes, which suggests they formed under drier, and potentially windier, paleoenvironmental conditions. Given this dune field lies between those studied to the west (Grigal *et al.*, 1976; Keen and Shane, 1990), the east (Filion, 1987; Thorson and Schile, 1995) and the northeast (Filion *et al.*, 1991), no single cause or period of eolian mobilization can be assumed. Thus, an investigation of interior dune landscapes within Michigan's upper peninsula would further clarify the causal and temporal variability of eolian mobilization in northeastern North America.

The purpose of this thesis is to reconstruct the geomorphic history of an interior dune field of Michigan's eastern upper peninsula. Three working hypotheses are considered for catalysts of eolian activity: 1) dune formation shortly after deglaciation, 2) devegetation and activation as a result of mid-Holocene warming and drying, and 3) episodic remobilization of dunes in the late Holocene. Absolute and relative dating techniques will be used to generate data regarding spatial and temporal patterns of regional dune activity. Potentially, these pilot data will better define late Quaternary paleoenvironmental conditions in the Great Lakes region, further refining the regional geomorphic record.

CHAPTER II

RELEVANT RESEARCH

Eolian Processes and Dune Formation

A basic understanding of eolian processes is fundamental to sand dune research. Like water, air is a fluid. Air is 1000 times less dense than water, however, which limits both the size of particles transported by wind and the extent of movement (Bagnold, 1941). Translocation of sand is resisted largely by gravity, friction, and cohesion forces (McKee, 1979). Thus, a minimum wind speed, or fluid threshold velocity, is required to overcome these forces and initiate particle transport. Threshold velocities typically increase as particle size increases. Generally, threshold velocity for medium sand (0.5 mm-0.25 mm) is 6 ms^{-1}, but individual rates vary with particle shape and sorting, surface roughness and moisture characteristics (Lettau and Lettau, 1978).

Particle transport by wind occurs by three main processes: suspension, saltation, and creep (McKee, 1979). Although fine sediment (i.e. silt and clay) can be carried in suspension, this process rarely transports particles ≥100 μm in diameter (Nalpanis, 1985). As a result, suspension is a relatively minor agent in sand dune formation. The majority of eolian sand transport occurs by saltation (Anderson, 1989). Saltation occurs when particles skip or bounce along the surface after the minimum threshold velocity is exceeded (Figure 2:1). Once airborne, saltating particles are transported along asymmetrical trajectories for relatively short distances. Although sediment size and wind speed ultimately determine the precise degree of movement, most particles travel within 10 mm of the surface and are displaced 12-15 times more horizontally than vertically. As

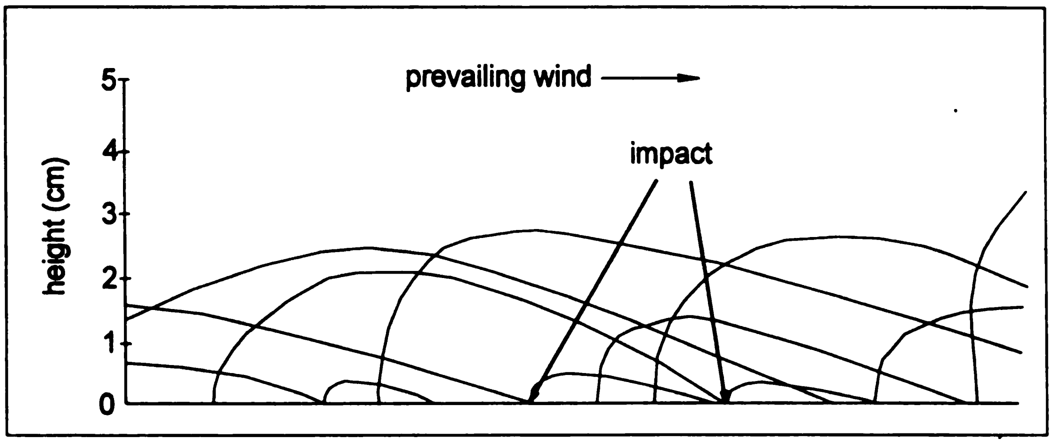


Figure 2:1 - Generalized Saltation Patterns (Modified from Ritter, 1986).

sand grains return to the surface, the resultant impact entrains additional particles on the ground, forcing them into the air. Consequently, the initial saltation of particles induces further saltation.

Particles too large for airborne translocation are moved primarily by creep. Creep involves the rolling of coarse sediments along the surface and is initiated by the impact of finer, saltating particles (Bagnold, 1973). Additional movement occurs when coarse particles roll into small depressions or "micro-craters" left by saltating particles.

Accumulation of eolian sand commonly occurs with changes in surface roughness, the presence of slight depressions, or due to small obstacles (e.g., vegetation; Kocurek *et al.*, 1992). Once the deposit reaches a height of roughly 30 cm, eolian sands exhibit a characteristic dune profile consisting of three components: the ***backslope***, ***crest***, and ***slip-face*** (Figure 2:2). The backslope is the windward surface where erosion is the dominant process and slope declivity is generally between 10° and 15° (Ritter, 1986). In contrast, the slip-face or lee slope is inclined between 30° and 35°, near the angle of repose for most sands. This surface represents a depositional surface where particles avalanche from the dune crest and are deposited once they are sheltered from the wind. The crest is the relatively flat summit or cap of a dune where erosion roughly equals deposition.

Given that sand dunes are dynamic features, they commonly migrate in accordance with prevailing winds. McKee (1979) observed that unrestricted dunes can migrate over 1 m per day under windy conditions. In order for dunes to remain in equilibrium, or maintain their shape while migrating, deposition on the slip-face must equal erosion along the backslope. In general, erosional surfaces contain more coarse sediments while finer particles dominate depositional structures (Bagnold, 1941).

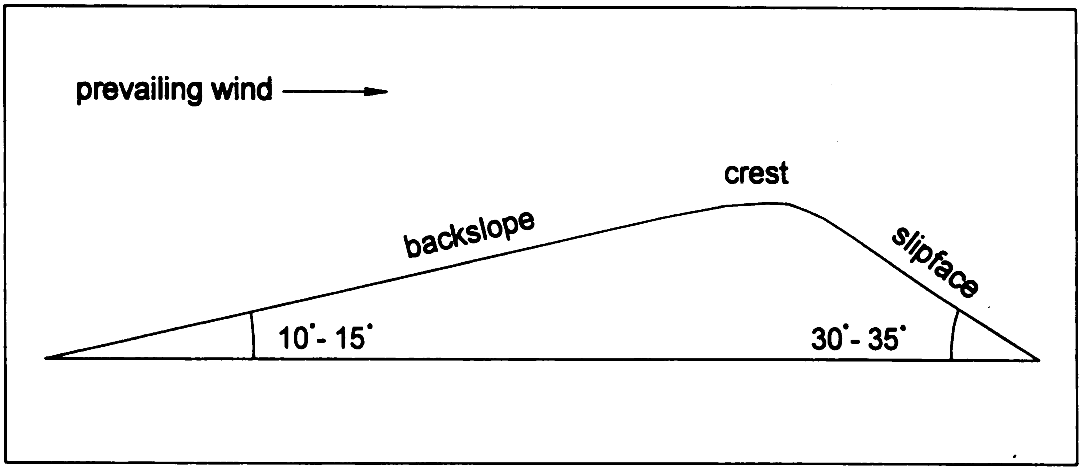


Figure 2:2 - Characteristic Dune Profile (Modified from Ritter, 1986).

While dunes are difficult to classify based on formative processes, they have been categorized by elemental shapes. McKee (1979) recognized nine broad classifications of dune morphology (Figure 2:3), with each type found in compound and complex forms. Typically, wind regime is the primary factor governing active dune configurations (Fryberger and Dean, 1979). *Transverse dunes* are continuous ridges of sand with crests perpendicular to dominant winds (McKee, 1979). Slip-faces are all orientated roughly in the same direction and net sand transport is normal to the dune crest. *Barchanoid ridges* have a sinuous, asymmetrical crest that is transitional to barchan dunes. In contrast to transverse dunes, which develop with large sand supplies, *Barchan dunes* evolve with more limited amounts of sand. Barchans are crescentic in plan-view with dune limbs projecting downwind. Where strong, multi-modal winds persist, the slip-face of barchan dunes may be truncated and rounded *dome dunes* develop. *Linear dunes* are visually similar to transverse ridges, but evolve when net sand transport is parallel to the crest. Slip-faces frequently exist on either side of the dune crest, although only one is active at a given time. *Star dunes* develop in multidirectional winds and have a number of slip-faces radiating from a central peak. As a result, star dune migration rates are relatively low (McKee, 1979). *Reversing dunes* form where two winds from nearly opposite directions are balanced with respect to strength and duration. Consequently, a second slip-face commonly develops.

In contrast to active dune morphologies, the migration of anchored dunes is restricted by the presence of topographic barriers and/or vegetative cover. McKee (1979) recognizes two primary forms of anchored dunes: *blowout* and *parabolic*. Blowouts are bowl-shaped hollows that are eroded into the loose sediments of an otherwise vegetated

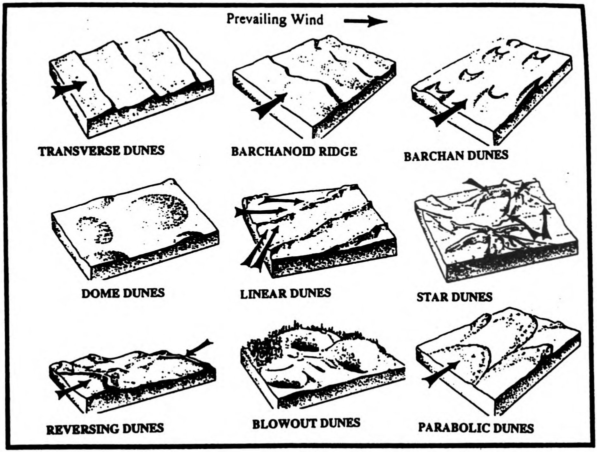


Figure 2:3 Generalized Dune Forms (Modified from McKee, 1979).

dune. These features typically concentrate near the crest of a dune due to higher wind velocities and increased desiccation. Although these features are largely deflational, deposition does occur where margins migrate downwind.

Parabolic dunes develop from blowout features (Summerfield, 1991). Prolonged devegetation and erosion of the central portion of a dune allows it to migrate downwind while vegetative cover protects the limbs from wind shear. As a result, a U- or V-shaped parabolic dune develops with limbs pointing upwind. The shape of parabolic dunes is influenced by the type of vegetation (i.e. trees or grasses) among which it develops (Filion and Morisset, 1983). Typically, parabolic dunes of a given area migrate in similar paths, creating a series of concentric dunes. These configurations indicate the dominant wind direction during mobilization.

Eolian Activity in Central North America

Dune landscapes are found in a variety of locations in central and northeastern North America. Episodes of late Quaternary (Pleistocene and Holocene epochs) eolian sand mobilization have been reported within these regions. Stabilized dune fields within the mid-continent are distinguished most readily by the type of vegetation that anchors them. Broadly stated, dunes in the semi-arid Great Plains are stabilized by grasses (e.g.,Wright *et al.*, 1985; Forman and Maat, 1990; Muhs and Maat, 1993; Arbogast, 1996), while dune fields in the more humid regions to the east and north are anchored by forest (e.g., Grigal *et al.*, 1976; David, 1981; Filion, 1987; Thorson and Schile, 1995; Arbogast *et al.*, 1997).

Central Great Plains Dune Fields

The Great Plains region is an extensive semi-arid to sub-humid grassland that extends from the Alaskan arctic to southern Texas. The Rocky Mountains serve as the western boundary, while a more diffuse eastern boundary stretches from Texas to south-central Canada. Considerable portions of the Great Plains are mantled by eolian sand, forming sand sheets and dunes. The Nebraska Sand Hills, for example, is the largest sand sea in the western hemisphere (Ahlbrandt *et al.*, 1983) and extensive sand deposits are also found in Colorado (Muhs, 1985), Texas (Holliday, 1995) and Kansas (Arbogast, 1996).

Investigations of Great Plains dune fields indicate mobilization occurred episodically during the late Quaternary. Research conducted in northern Colorado (Forman and Maat, 1990) and central Nebraska (Wright *et al.*, 1985), for example, indicates activity during the cool, dry late Wisconsin. Subsequently, it appears that some mobilization occurred during the warm, dry mid-Holocene (Ahlbrandt *et al.*, 1983).

While some deposits of eolian sand may date to the late Pleistocene or middle Holocene, most Great Plains dunes appear to be late Holocene landforms. Ahlbrandt *et al.* (1983) used radiocarbon dating and soil characteristics to conclude that dunes in the Nebraska Sand Hills stabilized after 3000 yrs B.P. Additionally, dune soils that had morphological and textural similarities to those described by Ahlbrandt *et al.* (1983) were identified in northeastern Colorado (Muhs, 1985). Using soil morphology and archeological evidence, Muhs concluded that dune mobilization probably occurred between 3000-1500 yrs B.P.

Forman and Maat (1990) also recognized late Holocene dune activity in northeastern Colorado. Using radiocarbon and thermoluminescence age estimates, the authors concluded that dunes most recently stabilized < 3000 yrs B.P. and were active during both cold-arid and warm-arid regimes. Additional research in northern Colorado (Madole, 1994) indicates that dunes mobilized within the past 1000 yrs and under conditions only slightly warmer and drier than at present.

A study conducted on the Great Bend Sand Prairie of southern Kansas suggests that five brief periods of late-Holocene stability and soil formation occurred (Arbogast, 1996). Radiocarbon samples extracted from a series of weakly developed (A/C profiles), buried soils suggest that brief periods of stability occurred at approximately 2300, 1400, 1000, 700 and 500 yrs B.P. Pedogenesis likely occurred when the climate was more humid. In summary, evidence of extensive, late Holocene dune activity has been recorded throughout the Great Plains region (e.g., Ahlbrandt *et al.*, 1983; Madole, 1994; Arbogast, 1996). Research conducted throughout the area clearly indicates that dune activation readily occurs given a decrease in effective moisture (e.g. Muhs and Maat, 1993; Forman and Maat, 1990). Given future greenhouse warming scenarios, many of the sand dunes and sand sheets of the Great Plains are likely to become reactivated (Muhs and Maat, 1993).

Forested Dune Fields

In addition to dunes of the semi-arid Great Plains, a variety of forested dune fields are located within the more humid regions of eastern and north-central North America (Figure 1:1). While dunes mantled by grasses are especially susceptible to climatic fluctuations, forest-anchored dunes presumably require more significant environmental change to initiate mobilization. Geomorphic research indicates that dunes in these regions evolved due to a variety of paleoenvironmental conditions.

The oldest dunes in forested regions evolved in freshly deglaciated landscapes. Such environments are favorable for dune formation as unconsolidated sediments, not yet anchored by vegetation, are easily reworked by wind. Forested dune fields have been identified in various deglacial environments, including Connecticut (Thorson and Schile, 1995), southern Quebec (Filion, 1987), and northern Saskatchewan (David, 1981).

The earliest period of eolian sand mobilization following the recession of the Laurentide ice sheet was recorded by Thorson and Schile (1995). They theorized that deglaciation began approximately 20,000-22,000 yrs B.P. throughout the New England region. Glacial Lake Hitchcock subsequently formed (13,000 yr. B.P.) along the ice margin and continually expanded northward to become the largest of New England's ice-recessional lakes. A large, elongate ridge of eolian sand, referred to as the Longmeadow dune, then formed along the eastern shoreline of Glacial Lake Hitchcock between 12,700 and 12,400 yrs B.P. Approximately 10 m high and 1 km in length, this transverse dune is characteristic of strong, unimodal winds and abundant sand supply. Several primary sedimentary structures (e.g., grain-fall and grain-flow strata, planar lamination, ripple-form strata, and dry avalanching), similar to those found in modern dunes of arid environments, are present in the Longmeadow dune. However, secondary structures are lacking, suggesting that the dune accreted rapidly. Additionally, bedding orientation indicated that dune-forming winds were predominantly from the northeast. The authors concluded that these winds were attributed to katabatic outflow from the Laurentide ice sheet and driven by a hypothesized glacial anticyclone (COHMAP Members, 1988).

Using variations in sediment composition and the thickness of varves from the bottom of Glacial Lake Hitchcock, Thorson and Schile (1995) concluded that the lake drained rapidly about 12,400 yrs B.P. Subsequently, parabolic dunes formed on exposed lacustrine sands. Dune orientations indicate that dune-forming winds originated from the west-northwest, suggesting a prevailing regime similar to that of today. Existing pollen records also demonstrate an increase in regional temperatures (Webb, 1987). The authors speculate that the sudden exposure of an unvegetated lowland, coupled with warmer conditions, led to localized pulses of eolian activity.

Following this warm regime, near-glacial conditions returned to the region from about 11,000 to 10,000 yrs B.P. and reactivated the Longmeadow dunes. A cool, moist regime dominated this period, causing a reversal to more boreal taxa throughout the area. Eolian mobilization was limited, however, to the reworking of pre-existing dunes and sand transport during this period ended rapidly. The authors found no evidence of dune activity at Longmeadow in the Holocene.

Stabilized, forested dunes also exist in the central St. Lawrence lowland of southern Quebec (Filion, 1987). The dunes lie on the periphery of the St. Lawrence River, are parabolic, and are orientated northeasterly. Glacial chronologies indicate that the region became subaerial about 9500 yrs B.P. following the regression of the Champlain Sea (Gadd, 1971). In order to establish the history of dune development, Filion (1987) examined a series of stratigraphic sequences and obtained 25 radiocarbon dates from dune and sub-dune strata. She concluded that dune formation occurred immediately following the withdrawal of the Champlain Sea. Radiocarbon ages indicate eolian activity from about 9500 to 7500 yrs B.P. Northeasterly-orientated dunes are consistent with the theorized anticyclonic winds induced by the retreating Laurentide ice sheet during this interval. According to Filion, dune activity ceased following the breakup of the anticyclone (about 7500 yrs B.P.) when a more humid climatic regime was established in the region. Dunes have been largely stable since that time, with only very localized activity being provoked by wildfire and anthropogenic perturbations.

While dune formation has been linked to Laurentide ice retreat in the southeast (e.g., Connecticut and southern Quebec), deglacial dunes also developed along the northwestern side of the ice margin. For example, a study by David (1981) indicates that dune formation in newly deglaciated sediments occurred in northern Saskatchewan. The Athabaska dunefield contains a variety of eolian features that were previously identified both as "ice-crack moraines" and beach ridges. David used dune morphology, structure, and sedimentology, as well as glacial chronologies, to reconstruct the history of these dunes.

Complex deposits of eolian sand, referred to as "Cree Lake type dune ridges" by David (1981), exhibited primary sedimentary structures. These structures are similar to those found in parabolic dunes. David theorized that dune ridges evolved from simple and composite parabolic dunes through the process of dune elongation. Additionally, bedding planes within the dunes indicated paleowinds from the southeast, which is essentially opposite from the present-day wind regime. Given these orientations, David attributed eolian activity to anticyclonic winds driven by glacial ice to the east. These cool, dry winds sufficiently hindered stabilizing vegetation, allowing dune mobilization to occur from about 10,000 to 8800 yrs B.P. Immediately after this interval, a sudden climatic change caused a southwesterly shift in predominant winds and the rapid stabilization of dunes. Subsequently, no large-scale eolian activity has been recorded for this dune field.

While eolian activity has been demonstrated in newly deglaciated environments, forested dune mobilization has also been linked to Holocene climatic fluctuations. Research conducted by Grigal *et al.* (1976), for example, used radiocarbon control and soil morphology to date dune activity in north-central Minnesota. Lake Winnibigoshish (Figure 1:1), one of the largest lakes in the state, is bordered by an extensive dune field to the southeast. These dunes are parabolic with northwesterly orientations. When the construction of a dam raised the lake level by 3 m, subsequent wave action eroded dunes adjacent to the lake. As a result, a sequence of five buried soils was exposed in dunes along the lakeshore, prompting a study of eolian activity.

The basal soil of the sequence is continuous throughout the exposure and expresses a A/Bs/C profile. Charcoal fragments recovered from the A1b horizon of this soil yielded a radiocarbon age of about 7910 yrs B.P. The overlying buried soils all have weakly developed, discontinuous A/C horizonation. Organic residues recovered from the top of the third buried soil rendered a radiocarbon date of about 5040 yrs B.P. Measurable amounts of Fe were found only in the basal and surface soils, indicating soil development under an acidifying vegetative cover. Particle-size distributions for all buried soils are similar to the modern C horizon, suggesting a uniform parent material.

Grigal *et al.* concluded that climatic fluctuations were responsible for dune activity in the region. The establishment of conifers soon followed deglaciation (approximately 11,600 yrs B.P.) and led to the development of the basal, spodic soil. Warmer and drier conditions occurred from about 8000 to 5000 yrs B.P. and caused a shift of regional vegetation from forest to a grassland-dominated landscape. As a result, dunes mobilized episodically during periods of prolonged drought. After about 5000 yrs B.P., a more humid climate returned to the region, leading to the re-establishment of forest cover and the stabilization of dunes.

In addition to the dunes at Lake Winnibigoshish, other forested dunes exist in the Anoka sand plain of east-central Minnesota. These dunes are also parabolic, with northwesterly orientations. Keen and Shane (1990) used magnetic susceptibility, pollen stratigraphy, and radiocarbon dating of lake-bottom sediments to reconstruct dune activity along the northwest margin of Lake Ann.

Results indicate that heightened eolian activity occurred in the area during the early to mid-Holocene. Specifically, Keen and Shane (1990) recognized three distinct episodes of dune mobilization, with peak activity transpiring at about 7400, 5800, and 4900 yrs B.P., respectively. Each episode involved a period of drought and subsequent loss of vegetation, an increase in eolian flux, and finally a return of more mesic conditions, forest cover, and dune stability.

Generally stated, the periods of climatic warming in east-central Minnesota (Keen and Shane, 1990) correlate well with those documented by Grigal *et al.* (1976) in the north-central portion of the state. Additionally, both studies correlate favorably with the Altithermal, an interval of mid-Holocene (7500-4000 yrs B.P.) warm, dry conditions (e.g., Wright, 1976; Baker *et al.*, 1992; Bradbury *et al.*, 1993). The Altithermal, or "Long Drought", has become an increased area of research in the Great Plains region. Although much is known of the Altithermal in western North America, the degree and extent of this climatic interval to the northeast is still unclear.

While eolian activity has been linked to Holocene climate fluctuations and newly deglaciated landscapes, a notable relationship also exists between forested dune mobilization and wildfire events. Research of this nature was conducted by Filion (1984,1991) along the eastern shore of Hudson Bay in Quebec, Canada. In both of her studies, Filion used radiocarbon dating to reconstruct a regional eolian activity chronology.

In the initial study, Filion (1984) examined a large field of parabolic dunes along a transect that extended from shrub tundra to boreal forest. Several sequences of buried paleosols were documented throughout the study area, indicating multiple periods of mobilization and stabilization. Filion obtained 101 radiocarbon dates from charcoal and wood fragments within the paleosols to establish the chronology of activity. Specifically, three major periods of dune mobilization were recognized: 3250-2750 yrs B.P., 1650-1050 yrs B.P., and 750 yrs B.P. to the present.

Filion concluded that dune activity in the boreal zone was driven by a periglacial cycle. The cycle was initiated by cool, dry conditions that promoted large-scale forest fires. Dune mobilization followed, as existing vegetation was stripped and cool, dry conditions delayed post-fire plant recolonization. Eventually, a shift to a warmer, more humid regime resulted in revegetation, dune stability, and soil formation. The cycle repeated whenever cool, dry conditions were persistent enough to initiate fire.

In the following study, Filion *et al.* (1991) focused on the eolian chronology within the forest and shrub tundra regions. A set of 196 radiocarbon dates was used to demonstrate diachronous dune mobilization throughout the region. The authors linked temporal differences in dune mobilization to vegetative cover. Generally, boreal forest dunes started to mobilize around 6000 yrs B.P., while forested/shrub tundra dune activity began about 4250 yrs B.P. and shrub tundra dunes followed around 3900 yrs B.P. Results agree favorably with Filion's previous study (1984), as catastrophic fire frequencies apparently triggered major episodes of eolian activity.

Relative Age Dating in Soils

In the absence of absolute age control, relative age dating has aided in timing geomorphic events. Soil development has proved to be an effective tool in relative age determination. The degree of soil development is dependent upon five soil-forming factors: climate, parent material, relief, organisms, and time. Jenny (1941) developed a functional-factorial model that used these soil-forming factors to explain soil development. Generally, the model states that by holding four of the factors constant, variations between soils are explained by the remaining factor. Consequently, by selecting soils that are consistent in climate, parent material, relief and organisms, one can attribute variations in soil development to the age of the soil. Sequences of such soils are referred to as soil chronofunctions.

Jenny (1941) defined a chronosequence as a collection of related soils in a geographic region that differ primarily as a result of the soil-forming factor of time. Bockheim (1980) selected 32 chronosequences from the literature to construct chronofunctions. Soils were studied over seven different climatic regions and within seven different parent materials. Further, chronosequence ages ranged from 0-500 yrs B.P. to over 1,000,000 yrs B.P. Bockheim used these highly variable settings to compare soil properties in different climatic regions and parent materials and to determine what soil properties are consistent throughout the chronosequences. Results indicated that most soil properties continued to change with time, even with the passage of over 1 million years. Soil development was shown to be heavily dependent on climate as the rates of increase in solum thickness, oxidation depth, and clay content in the B horizon were positively correlated with mean annual temperature. Parent material also influenced soil development rates as solum thickness and B horizon clay content were positively correlated with parent material clay content. Base saturation and soil pH decreased with time independently of climate, parent material, and organic factors. Bockheim suggested that, since soils continue to develop with time, rather than reaching a steady state, relative age dating in soils is valid, regardless of soil age.

Podzolization and Spodosols

Podzolization is a dominant soil-forming process in northern Michigan (Gardner and Whiteside, 1952; Brewer, 1982). This process involves the translocation of organic matter (OM), aluminum (Al) and sometimes iron (Fe) which may eventually lead to the formation of Spodosols. Spodosols, by definition, must have a spodic (Bs) horizon that meets predetermined chemical and morphological criteria (Soil Survey Staff, 1975). These illuvial (Bs) horizons are generally located beneath an eluvial mineral horizon, frequently a light-colored, albic (E) horizon.

In order for podzolization to take place, movement and accumulation of sesquioxides and organic matter must take place in the spodic horizon. For translocation to occur, exchangeable bases (Ca^+, Mg^+, Na^+) must first be leached from the upper part of the solum and displaced by hydrogen (H^+) ions, resulting in a lowered pH. This is best accomplished in sandy parent materials, which can be quickly acidified. A moist, cool environment also enhances podzolization as low temperatures inhibit the microbal decomposition of organic matter, allowing it to translocate. Such conditions are found in areas of coniferous forest. Therefore, Spodosols often develop in northern Michigan where acidic, sandy parent material and coniferous forest cover are commonplace.

Schaetzl and Isard (1991) used vegetative and climatic patterns to study the geographic distribution of Spodosols in the southern peninsula of Michigan (Figure 2:4). A floristic tension zone runs diagonally across this peninsula, separating the predominately deciduous forests of the south from the mixed coniferous-deciduous forests to the north (Brewer, 1982). This boundary also separates two broad types of parent material, with sandy sediment dominating in the northern zone and finer-textured, loamy materials prevailing to the south. Climatic conditions also vary north-to-south with snowfall amounts notably higher in the northern and northwest portions of the peninsula.

The authors concluded that a strong correlation exists between forest type, cold-season climate and podzolization. The floristic tension zone roughly coincides with the southern limit of Spodosols (Brewer, 1982), further suggesting that coniferous and mixed forest assemblages are favorable for podolization. Strong spodic development was also related to areas of thick winter snowpacks. Early winter snows act as an insulator for the soil, inhibiting soil frost. This allows for increased infiltration of meltwater in the spring, which heightens the podzolization process. Consequently, Spodosol development was found to increase northwesterly through the peninsula. Given that coniferous forest cover and snowfall frequency increase to the north, even stronger spodic development may exist within the northern peninsula of Michigan.

Although many pedons exhibit spodic morphology in the field, often these soils are

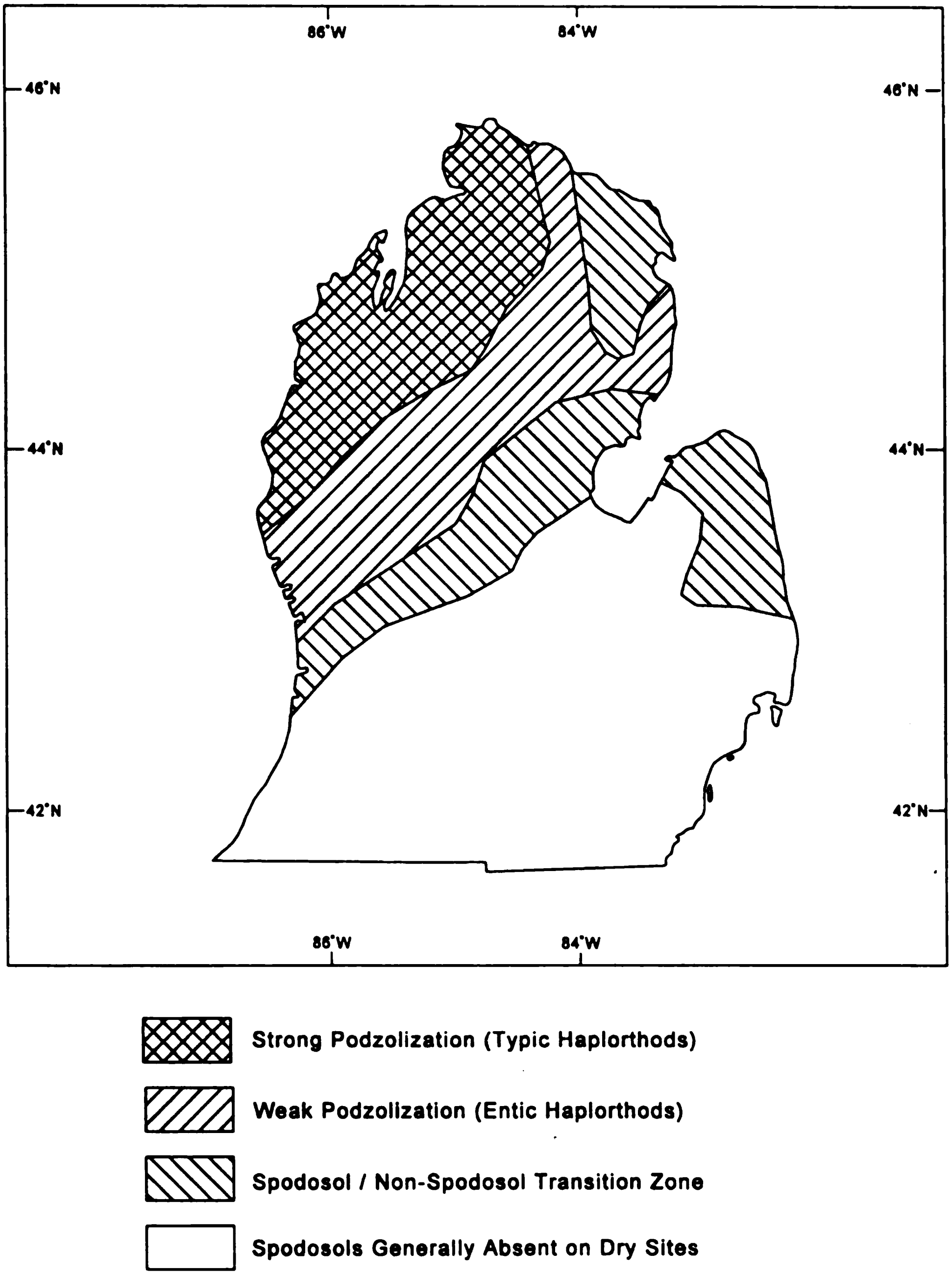


Figure 2:4 - Degree of Podzolization in lower Michigan. (Modified from Schaetzl and Isard, 1991).

not considered Spodosols based on the chemical criteria outlined by the Soil Survey Staff (Schaetzl and Mokma, 1988). The disparity between morphologic and chemically-determined class boundaries has created difficulties for field mappers trying to accurately identify Spodic morphology in the absence of chemical data. To better address this problem, Schaetzl and Mokma (1988) developed a numerical index of soil development, the POD index. Index values are determined solely from morphologic criteria in the field and offer a method of measuring spodic development without chemical data. The index assumes that as soil development increases: eluvial (E) horizons become "whiter", illuvial (B) horizons become "darker" and "redder" and, the number of B subhorizons increases. Schaetzl and Mokma applied the index to 723 pedons throughout the United States that either exhibited, or were developing, Podzol morphology. Results indicated that the index was effective in differentiating between Podzols and non-Podzols and also between Entic and Typic subgroups of Spodosols. Generally, results suggested POD index ranges of 0-2 for non-Spodosols, 2-6 for Entic Haplorthods, and >6 for Typic Haplorthods. The authors also suggested that the index is useful for chronosequence studies in Podzolic soils.

Spodosols as a Relative Age Dating Tool

A number of studies have used spodic development as an indicator of landform age. In northwestern lower Michigan, Franzmeier and Whiteside (1963) studied a chronosequence of spodic soils found on three lake terraces and a moraine of known age (Figure 2:5). The terraces were formed by fluctuating water levels in Lake Michigan during the Late Wisconsin and Holocene. The authors used soils at the crest of each

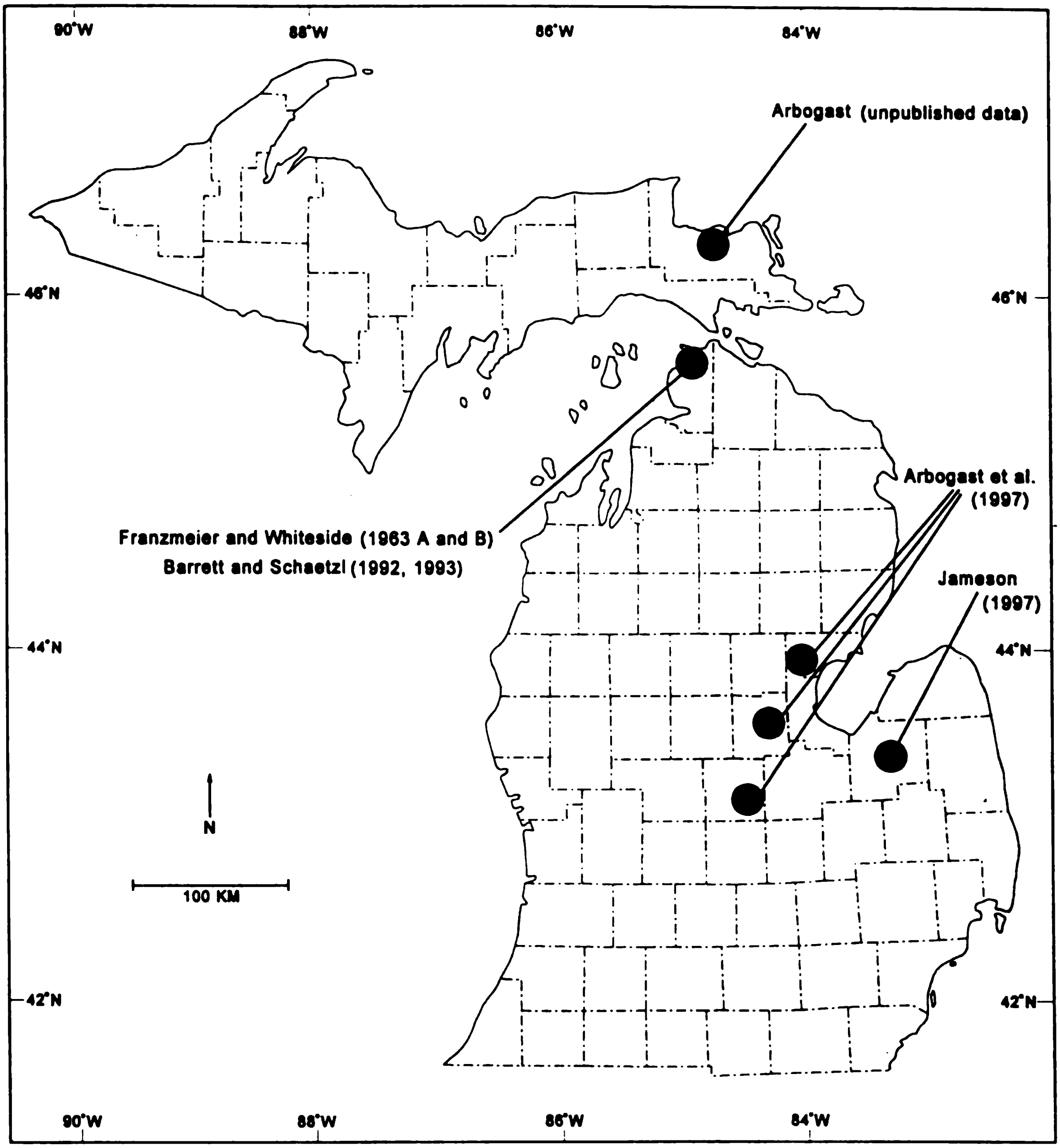


Figure 2:5 - Location of Research Areas within Michigan
(Modified from MIRIS base map, 1998).

surface to examine relationships on the Algoma (2250 yrs B.P.), Nipissing (4000 yrs B.P.), and Algonquin (8000 yrs B.P.) terraces, and the Valders moraine (11,800 yrs B.P.).

Fe and Al were extracted from all sampled horizons using sodium dithionite-citrate-bicarbonate. Results showed that as the age of the surface increased, Fe and Al measurements in the B horizons increased accordingly. Using the chemically determined class boundaries for spodic horizons (Soil Survey Staff, 1960), the authors suggested that a minimum of 3,000 years are needed for a Bs horizon to form while 8,000 years are necessary for Bhs development.

Following this study, Barrett and Schaetzl (1992, 1993) re-examined the lake terrace chronosequence. Four lake terraces were studied, with ages based on the center of age ranges cited in the literature: Algoma (3000 yrs B.P.), Nipissing (4000 yrs B.P.), Battlefield (10,000 yrs B.P.), and Main Algonquin: (11,000 yrs B.P.). Fe and Al were extracted from each horizon using sodium citrate-dithionite and acid ammonium-oxalate. Data derived from the extractions confirmed that Fe and Al were increasingly accumulating in the B horizons, with time. Mean weighted B horizon values, calculated to reflect soil development and soil development ratios, were regressed against terrace surface age to produce numerical chronofunctions. As with Franzmeier and Whiteside (1963a,b), Barrett and Schaetzl concluded that at least 4000, but less than 10,000 years, are required for spodic horizon development in northwestern lower Michigan.

A similar study by Arbogast *et al.* (1997) used soil development and a radiocarbon date to estimate the age of three inland dune fields in east-central Michigan (Figure 2:5). Overlying glaciolacustrine and outwash sediments, the dunes are parabolic with northwesterly-orientated limbs. They mantle wet landscapes. Potentially, these dunes could have formed directly after deglaciation, during a shift in Holocene climate, or at an intermediate period.

Three distinct dune fields were identified within three counties: Midland, Gratiot, and Arenac, with ten dunes being examined in each county. Soil pedons were described and sampled at each dune crest for further laboratory analysis. Results showed that soils and parent materials were similar in morphology, both within and between dune fields. No statistically significant differences were found in Fe and Al measurements, solum thickness and POD indices between dune fields. Mean POD indices for the dune fields were notably low (Arenac 0.4, Midland 0.4, Gratiot 0.0) which further suggested that dunes stabilized concurrently. Chemical data from the soils were then compared to soils of known age in the northern lower peninsula of Michigan (Franzmeier and Whiteside, 1963b; Barrett and Schaetzl, 1992). Results suggested that the dunes in east-central Michigan stabilized before 4,000 yrs B.P. but probably after 10,000 yrs B.P.

A closely related study by Jameson (1997) examined a stabilized dune field (Tuscola dune field) in east-central lower Michigan, to the southeast of the Arbogast *et al.* (1997) study (Figure 2:5). Dunes are parabolic, have a northwesterly orientation, and overlie glacio-lacustrine sediments that were deposited about 12,000 yrs B.P. Soil pedons at 19 dune crests were morphologically and chemically analyzed to suggest an eolian record within the Tuscola dune field. Resultant data were also compared to the dunes to the northwest to suggest if stabilization was concurrent between dune fields.

Soil horizonization, POD index values, and textural data indicate concurrent stabilization within the dune field. However, by applying statistical measures to chemical and morphological data, Jameson concluded that the Tuscola dunefield stabilized shortly before the dunes to the northwest. Results indicated that Tuscola dune soils had a significantly higher accumulation of Al and Fe in their Bs horizons than did the dunes studied by Arbogast *et al.* (1997). Further, cluster analyses clearly demonstrated the existence of two dune groups within the region, a Tuscola group and a Gratiot, Midland, and Arenac group.

A study by Arbogast (unpublished data) examined soils in a stabilized dune field perched on a hypothesized Nipissing terrace (Farrand and Drexler, 1985) in the eastern upper peninsula of Michigan (Figure 2:5). Dunes are parabolic and orientated to the northwest. Surficial soils were studied on 13 dune crests to provide an estimate of dune ages. Soil development was remarkably consistent between dunes as all pedons expressed A-E-Bhs-Bs-BC-C horizonation. POD indices were also consistent, ranging from 4 to 5 throughout the dune field. Fe and Al extractions from spodic horizons were compared to extraction data from lake terraces of known age in northwest lower Michigan (Barrett and Schaetzl, 1992). Arbogast concluded that the perched dunes stabilized at the end of the Nipissing transgression, about 5000 yrs B.P.

In summary, a great deal of research has been conducted in dune fields, both in grasslands and forest, in North America. Significant gaps remain, however, especially along the boundaries of major vegetation zones. A variety of well developed dunes exist in Michigan that have yet to be studied. Given the central location of Michigan dune fields, no single cause or period of eolian mobilization can be assumed.

CHAPTER III

STUDY AREA

The Seney Dune Field, located within the eastern upper peninsula of Michigan (Figure 3:1), contains extensive tracts of eolian sand. This thesis investigates a portion (approximately 2000 km^2) of this dune field, essentially encompassing all of Luce County and the extreme east-central portion of Schoolcraft County. Work was focused within Hiawatha National Forest, with additional investigations in Tahquamenon Falls State Park. The study area is bordered by the village of Seney to the west, Manistique Lake to the south, and Betsy Lake to the north and east. Regional drainage is provided primarily by the Tahquamenon River, which flows roughly southwest to northeast and empties into Whitefish Bay of Lake Superior.

Geology

The eastern upper peninsula of Michigan is predominantly a glacial landscape. Pre-Quaternary geology of the study area consists largely of Precambrian sandstone and Paleozoic limestone and dolomite bedrock units (Mickelson *et al.*, 1983). These bedrock units are mantled by glacial deposits of Pleistocene age that typically range from 5 to 20 meters in thickness (Leverett, 1929; Blewett and Rieck, 1987). An early, major advance of the Laurentide Ice Sheet took place during the Pre-Illinoian ($\leq$ 730,000 yrs B.P.) and extended to southern Indiana and Ohio (Eschman, 1985). Michigan was subsequently reworked by a series of glacial advances and retreats through the Wisconsin Stage (110,000-10,000 yrs B.P.; Flint, 1971; Mickelson *et al.*,1983; Eschman, 1985).

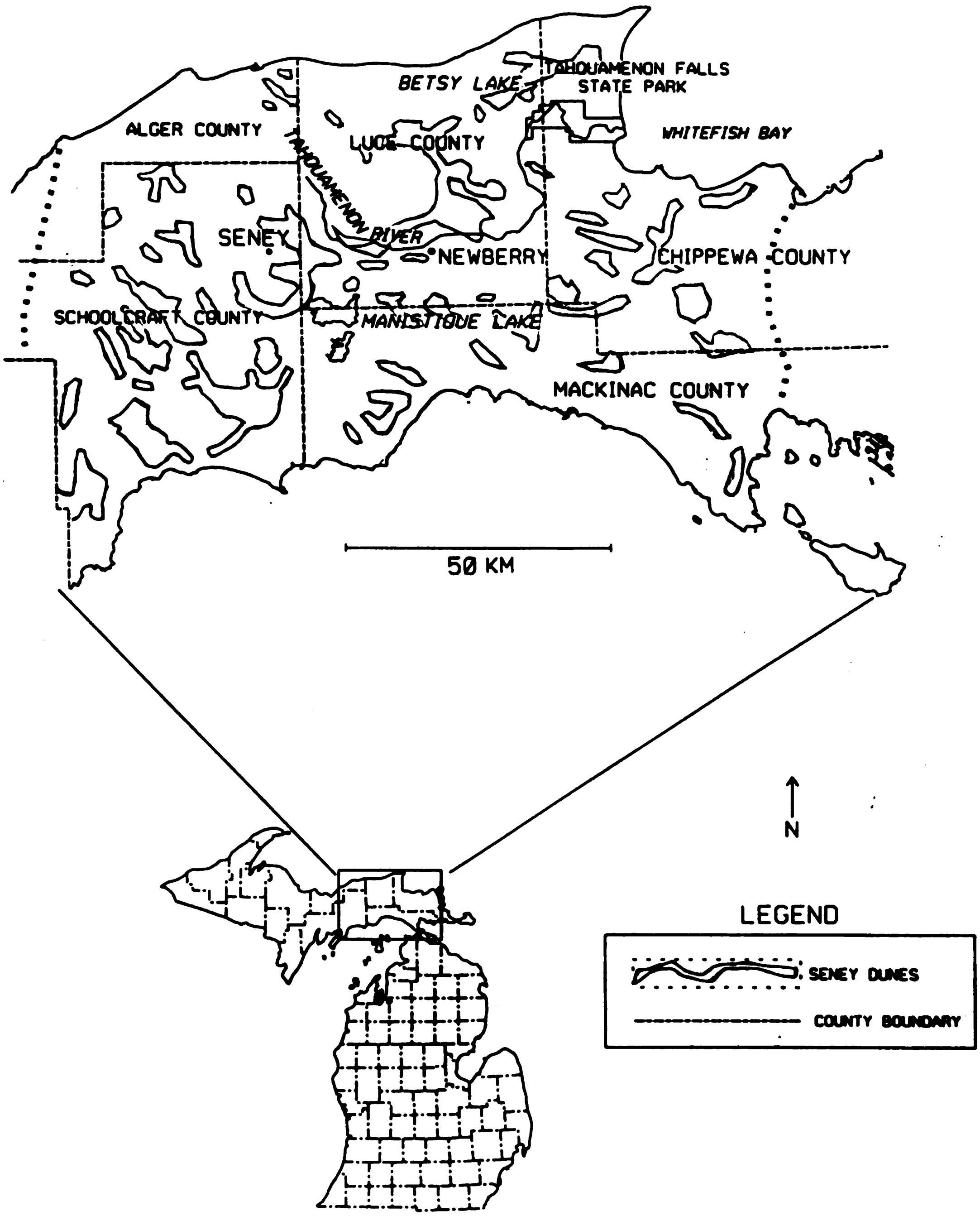


Figure 3:1 - Study Area (Modified from MIRIS base map, 1998).

The most recent glaciation influencing the study area took place during the Greatlakean stadial (11,850 yrs B.P.) of the Late Wisconsin, when the Superior lobe flowed southeasterly over the region (Figure 3:2). Final deglaciation took place around 11,500 yrs B.P. (Saarnisto, 1974; Drexler *et al.*, 1983). A minor readvance during the Marquette Stadial (9900 yrs B.P.) extended to the northernmost margin of the study area, marking the last known glacial activity within the region (Futyma, 1981).

As deglaciation progressed, a series of proglacial lakes developed in the region. Glacial Lake Algonquin represents the bodies of water that occupied all or part of the Michigan, Huron, Superior and Erie basins. Eschman and Karrow (1985) recognized four major phases of Lake Algonquin history, with the study area directly influenced by the latter two. Early Lake Algonquin entailed a high-level (184 m) phase with southeasterly drainage via the Port Huron outlet around 12,000-11,500 yrs B.P. During the low-water, Kirkfield phase (11,500-11,200 yrs B.P.), Lake Algonquin drained eastward to the Ontario basin through the Kirkfield outlet (Karrow, 1987). Increased meltwater brought about the Main Algonquin phase (11,200-10,600 yrs B.P.) where rapid retreat of the ice front extended the lake northward, at an elevation of 184 meters. During this interval, much of the study area was inundated (Figure 3:3; Futyma, 1981). Finally, the deglaciation of drainage outlets, coupled with regional isostatic uplift, resulted in a step-wise lowering of water levels during the Algonquin-Stanley phase (10,600-10,000 yrs B.P.). Several studies indicate that the study area has remained subaerial following the terminal drainage of Lake Algonquin (10,000 yrs B.P.; e.g., Drexler *et al.*, 1983; Eschman and Karrow, 1985; Karrow, 1987).

Current surficial geology reflects the influence of Glacial Lake Algonquin, with

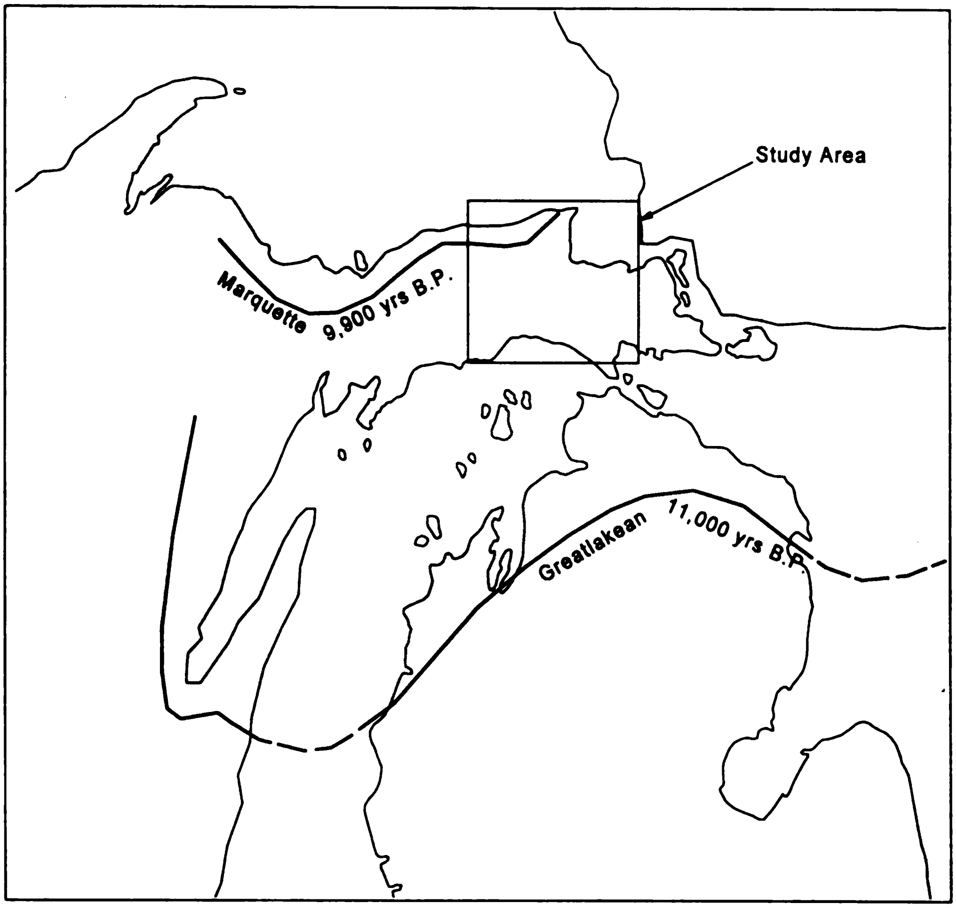


Figure 3:2 - Maximum limits of Greatlakean and Marquette Substages (Modified from Eshman and Karrow, 1985).

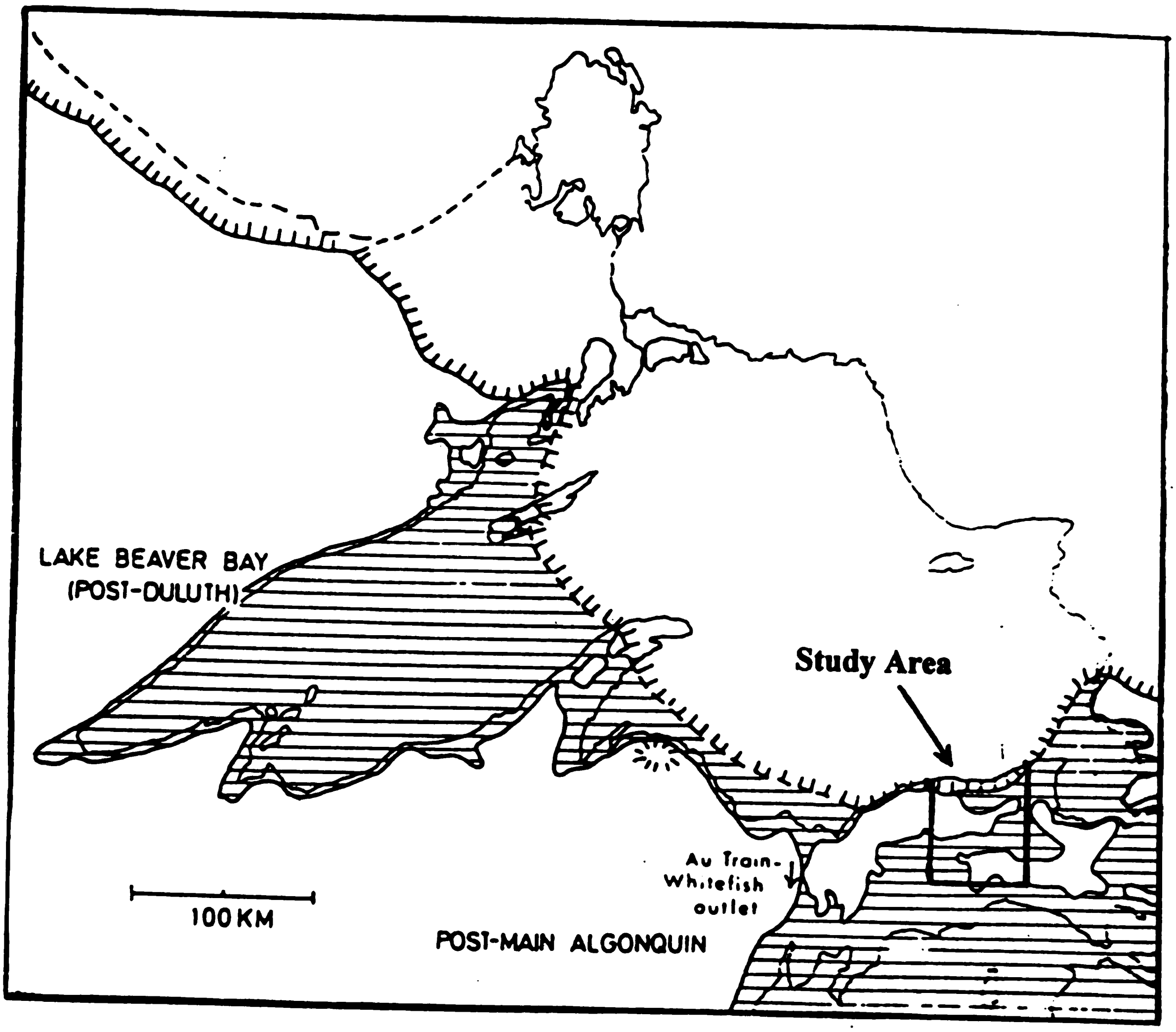


Figure 3:3 – Glacial Lake Algonquin (Modified from Eschman and Karrow, 1985).

extensive lacustrine deposits dominating the region. Dunes are largely parabolic in form with low-lying swampland typically occupying inter-dune areas (Figure 3:4). A northwesterly dune orientation is predominant throughout the study area (Figure 3:5). Dune deposits vary from less than 3 to over 8 meters in height and have maximum slopes of 6 to 12 percent (Berndt, 1977).

Soils

Podzolization is the dominant process influencing soil morphology of dunes in the study area. Dune soils are commonly represented by two series: Rousseu and Rubicon (Whitney, personal comm.). Although both series are classified as sandy, mixed, frigid Entic Haplorthods, Rousseu soils occur in finer sands and exhibit slightly more developed profiles (Oa/A/E/Bs1/Bs2/BC/C) than do Rubicon soils (Oe/A/E/Bs/BC/C; Berndt, 1977). Inter-dune, swampland areas often contain Carbondale soils, which are classified as euic Hemic Borosaprists and have Oa1/Oa2/Oe1/Oe2 horizonation (Whitney, personal comm.).

Climate

Newberry, located within the core of the study area, has maintained temperature and precipitation records since 1951. Regional climate is broadly characterized as semi-marine, humid continental (National Oceanic and Atmospheric Administration, 1997). While annual temperature variations in the region are great, the lakes moderate temperature relative to western locations (e.g., Wisconsin, Minnesota) of similar latitude. Winter typically extends from December to March with an average low temperature of

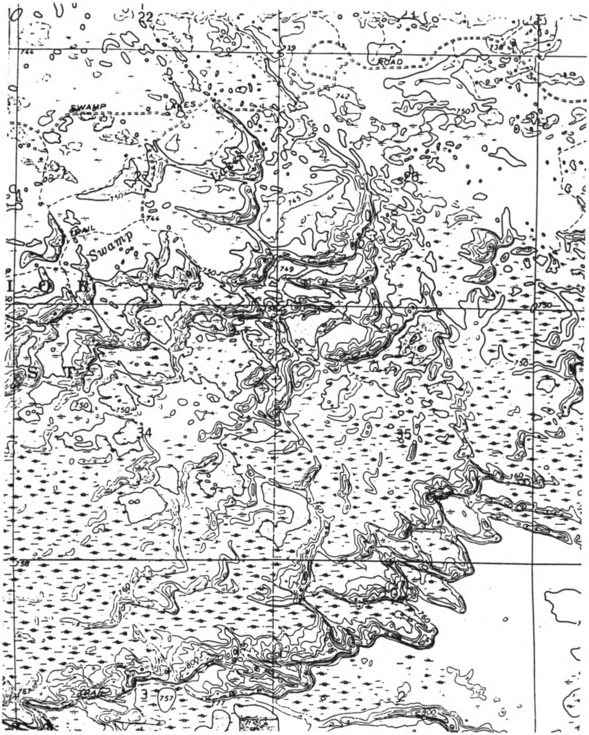


Figure 3:4 Parabolic Dunes Within Study Area (Modified from U.S.G.S. Betsy Lake SW, Mich., 1:24,000, 1968).

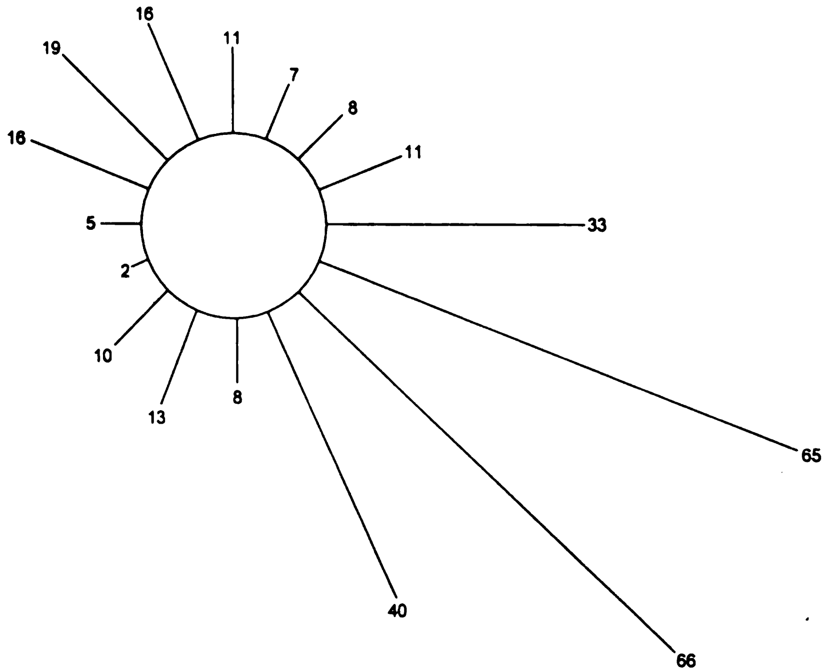


Figure 3:5 - Dune Orientation Rose (Seney dune field). Diagram depicts the orientation of the limbs of all identifiable parabolic dunes within the study area. Lengths of the arms are proportional to the frequency of occurance. Long arms in the SE direction imply that the majority of dunes have northwesterly orientation.

about -11.4°C. The summer growing season usually lasts from late May to September with an average high temperature of approximately 20.1°C.

Precipitation accumulates consistently throughout the year. Lake-effect winds, derived from both Lakes Michigan and Superior, commonly converge near the center of the peninsula during the spring and summer months and induce late afternoon showers (Harman and Hehr, 1972). About 58% (489 mm) of the annual precipitation total accumulates during the summer growing season (Table 3:1). Throughout the fall and winter, lake-effect processes lead to increased cloudiness and snowfall, with total snowfall averaging 2.75 m annually. The prevailing winds are west-northwesterly, averaging 14 km/hr (National Oceanic and Atmospheric Administration, 1997).

Table 3:1 - Mean Annual Temperature and Precipitation (1951-1980); Newberry, MI (National Oceanic and Atmospheric Administration, 1997)

	Temperature (C°) Mean Daily Max.	Mean Daily Min.	Precipitation (mm)
J	-4.9	-13.4	52.8
F	-3.5	-13.5	42.9
M	1.4	-8.7	53.9
A	9.3	-1.6	60.5
M	16.7	3.6	77.2
J	21.8	8.5	86.6
J	24.6	11.5	78.0
A	23.4	11.3	93.2
S	18.1	7.3	93.2
O	12.2	2.7	70.6
N	4.1	-3.3	69.6
D	-2.2	-9.9	58.2

Vegetation

Pre-settlement forest vegetation of the study area consisted primarily of white pine and red pine in well-drained, upland areas, while sugar maple and northern white-cedar were common, lowland swamp species (Taylor, 1991). During the late-nineteenth to early-twentieth centuries, widespread logging and associated wildfires extensively cleared the region (Findell and Chase, 1957). Since that time much land has been reclaimed by the government and established as state forest land. Currently, over 90% of the study area is classified as mixed, second growth forest. Common tree species include jack pine, red pine, sugar maple, northern white-cedar, yellow birch, black spruce, balsam fir, and aspen.

Cultural History

The earliest recorded humans inhabiting the study area were present during the late Holocene. Archeological research conducted in lower Whitefish Bay indicates the study area contains one of the most significant sites of Middle Woodland culture (2200-1200 yrs B.P.) in the Great Lakes region (Taylor, 1991). Several shards of pottery, flint scrapers and projectile points were recovered adjacent to the bay shorezone. Taylor (1991) attributed these artifacts to the Laurel culture, who favored lakes and rivers for habitat.

Just prior to European settlement, the Ojibwa (or Chippewa) tribe dominated the region, mainly occupying the banks of the Tahquamenon River (Taylor, 1991). The Ojibwa were nomadic and subsisted largely on whitefish and sturgeon during the ice-free season and deer during the winter months. With the arrival of European settlers to the east, Native Americans became heavily involved in the fur trade which greatly depleted natural wildlife within the region (Taylor, 1991).

By the 1820's, the Ojibwa faced European occupation of their tribal land. A series of seven treaties were signed from 1820 to 1855, leading to the first major cut of Tahquamenon valley pine around 1870. Construction of large sawmills ensued during the early 1880's and Newberry, McMillan and Emerson were established as major lumbering centers soon after (Taylor, 1991). Extensive logging of white and red pine continued until the mid-1920's. When large stands of pine had been depleted, northern hardwood logging spurred a regional pulpwood industry. Settlers also attempted to drain lowland swamp regions for agriculture, but were largely unsuccessful.

During the mid-nineteenth century, significant efforts were made to preserve the natural beauty in and around the study area. The Seney Wildlife Refuge, located near the town of Seney in east-central Schoolcraft county, was established in 1935 (Anderson, 1985). The Hiawatha National Forest and Tahquamenon State Park were also instituted in central and northeastern Luce County. While significant logging continues in the study area today, forest management practices have been adopted to ensure forest conservation. Newberry remains the center of employment, with work provided primarily by the Newberry State Hospital, Newberry State Correctional Facility, and various logging and limestone quarry operations throughout the area (Lamb, 1997).

CHAPTER IV

METHODS

The overall design of this study involves the application of both absolute and relative age-dating techniques to better understand the temporal and spatial variability in eolian activity in the eastern upper peninsula of Michigan. As a result of similar research, a generally accepted methodology has emerged which includes several field and laboratory procedures. The following discussion is a summary of the methods employed in this study.

Field Methods

In an effort to identify sites most suitable for analysis, the study area was initially surveyed using 7.5-minute U.S. Geological Survey topographic maps. Landforms expressing parabolic configurations and relatively high relief (>3m) were noted as potential study sites. Sand quarries were of particular interest because they often provide excellent stratigraphic exposures of eolian deposits. Following the preliminary map examination, an extensive field survey of the study area was conducted. General site characteristics (*i.e.* vegetative cover, drainage, slope, and disturbance) were noted and a multiple-extension bucket auger was used to examine sub-dune strata for organic-rich sediments suitable for radiometric dating. When identified, several sediment cores of the deposit were inspected along the dune base to insure contiguity.

Two sites yielded organic material adequate for radiocarbon dating and, consequently, detailed stratigraphic sections were completed at both sites. Detailed descriptions were compiled and each section was photographed. Horizon-based soil samples (~400 g) were described and collected using standard terminology and procedures (Soil Survey Staff, 1993) and additional samples were collected systematically at 20-cm intervals down the section profile for sedimentological analysis. Bulk radiocarbon samples were retained and wrapped in aluminum foil to protect against contamination. Both samples were sent to the INSTAAR laboratory in Boulder, CO for age determination.

A subsequent, surficial examination of 10 dune soil pedons was performed along an east-northeasterly-trending transect. Selected study sites were delineated at roughly regular (5-10 km) intervals and located on flat, stable dune crests. Physical characteristics, such as slope, vegetative cover, and potential disturbance, were carefully examined to ensure uniformity across sample sites. Horizon-based soil samples were then collected from hand-dug pits (>1.5 m) or available exposures.

Laboratory Methods

Munsell colors were determined on field-moist soils under fluorescent light by two persons working independently. Samples were then oven dried for three hours at 60°C and passed through a 2 mm sieve to remove large organic detritus. The remaining sample was repeatedly halved, using a sample splitter, to insure homogeneity. A sub-sample of approximately 30 g was retained for all subsequent analysis (Soil Survey Staff, 1993).

Particle size determinations were made by dispersing samples in a $Na_2CO_3[NaPO_3]_6$ solution, shaking for 10 hours, and wet-sieving through a 53 μm sieve. Silt and clay content determinations were made by hydrometer analysis of sediment that passed through the sieve. Sand contents were then dry-sieved to separate very coarse (vcs), coarse (cs), medium (ms), fine (fs), and very fine sand (vfs) fractions. Solum-weighted amounts for each particle size separate were determined by dividing each horizon-weighted sum total by solum thickness. Organic matter content of radiocarbon samples was determined by loss-on-ignition at 450 °C (Ball, 1964). Reaction of samples in 2:1 water-soil mixtures was analyzed using an Orion pH meter (model #720A).

Two chemical extractions were run for all Bs and C soil horizons to identify the forms of Fe and Al in each sample. Sodium-citrate-dithionite extraction is generally accepted to remove all "free" Fe (Fe_d) and Al (Al_d) - i.e., that which is not included in silicate mineral structures. Fe_d and Al_d can exist as amorphous and crystalline oxides, as well as when bound to organic complexes. Ammonium-oxalate extraction is thought to isolate active Fe (Fe_o) and Al (Al_o) which can be present as amorphous oxides and organic complexes. Both sodium-citrate-dithionite and ammonium-oxalate extractions were performed using standard procedures (Soil Conservation Service, 1990) and Fe and Al concentrations were measured by a DCP spectrophotometer (Daly, 1982) at Michigan State University. Horizon-weighted measures of Fe_d, Fe_o, Al_d, and Al_o were calculated for all Bs horizons by multiplying the amount of each extract by the thickness (cm) of each horizon.

Additionally, in an effort to determine prevailing, dune-forming paleowinds within the study area, a sand rose histogram was constructed (Figure 3:5). A sand rose graphically displays both the directional variability and degree of potential sand drift based on the sixteen directions of the compass. Conforming to the methods outlined by Arbogast *et al.* (1997), the azimuth of the axis of all clearly identifiable dunes within the study area was determined, directional counts summed, and a sand rose diagram constructed.

CHAPTER V

RESULTS AND DISCUSSION

The purpose of this study is to estimate the age of sand dunes within the eastern upper peninsula of Michigan. Toward that end, radiocarbon ages were obtained at two widely-spaced sites. In addition, surface soils were examined along a study transect connecting the radiocarbon dates. The resultant chemical and morphological data provide relative age control. Lastly, physical signals, such as dune orientation and landscape characteristics, are used to reconstruct paleoenvironmental conditions. This chapter describes the data derived from this study.

Seney Sand Pit

The Seney Sand Pit is an active sand quarry located 1 km south of the town of Seney (Figure 5:1). Approximately 8 m high, this site consists of eolian sands that overly another depositional unit (Figure 5:2). The dune is parabolic with northwesterly-orientated limbs. The deposit is anchored by secondary forest, while the surrounding landscape is lowland swamp. This site is rare because the eolian sands overlie sediments sufficiently fine and organic-rich so as to be radiocarbon datable.

The lowermost unit (Unit I) at the Seney Sand Pit extends from about 8 m to 5.59 m below to surface (Figure 5:3). Further investigation shows that the deposit is traceable and generally uniform throughout the entire quarry. Textural analysis indicates that the lower portion of the unit is comprised almost entirely of sand, with individual fractions as follows: coarse sand (0.3%), medium sand (37.3%), fine sand (55.9%), very fine sand

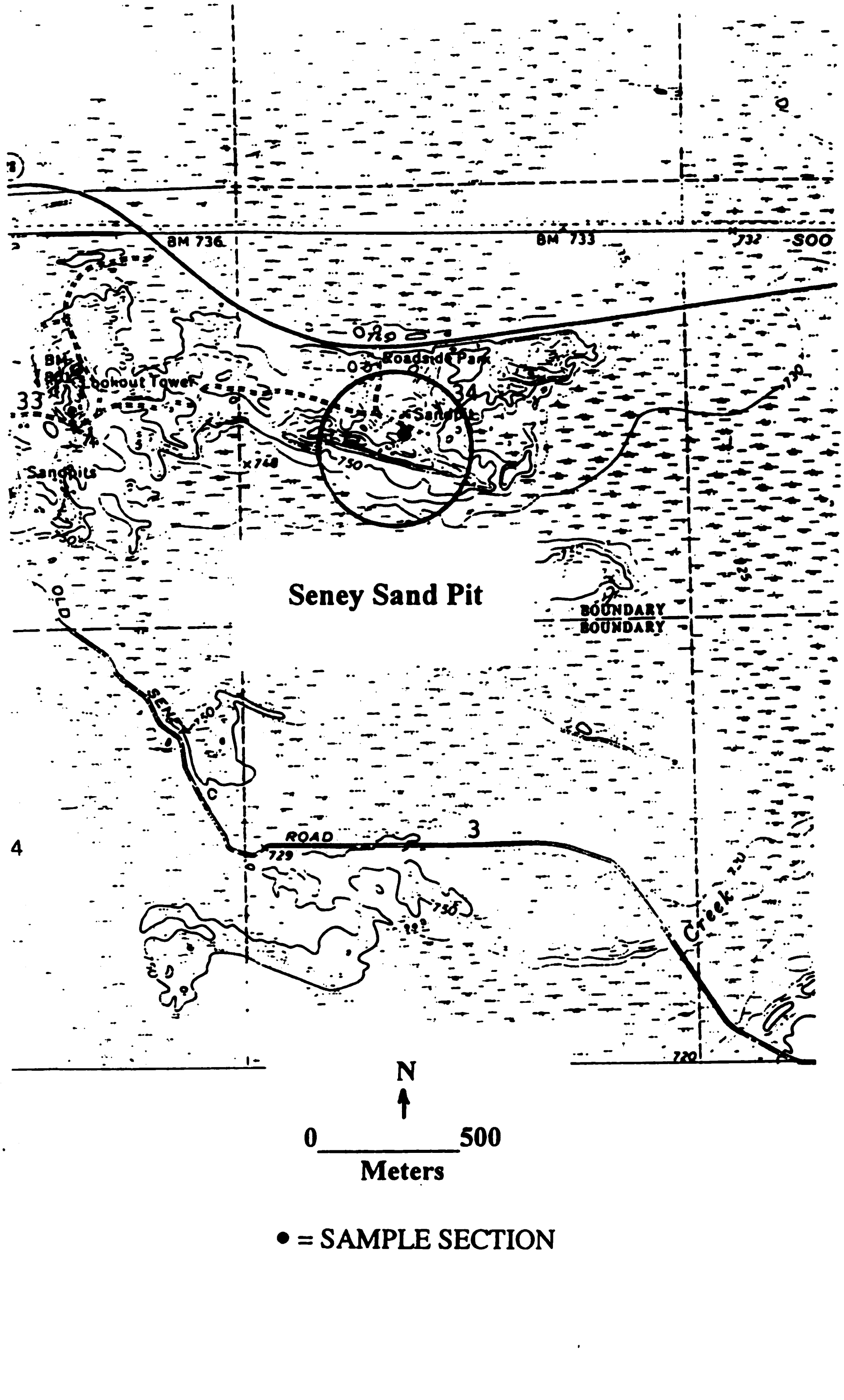


Figure 5:1 - Topographic Map of the Seney Sand Pit, located in the SW 1/4, sec. 34, T 46, R 13, Seney, MI Quadrangle, 1: 24,000 (1972).

Figure 5:2 – View of the Seney Sand Pit that shows the lowermost lacustrine deposit (Unit I) mantled by about 5.6 m of eolian sand (Unit II). Radiocarbon age determination on organic material at the contact surface yielded an age of 4940 ± 60 yrs B.P.

Unit II

4940 ± 60 yrs B.P.

Unit I

Figure 5:3 – View of Unit I. This deposit extends from about 8 to 5.59 m.

(6.1%), and silt (0.4%), respectively. Sediments become progressively finer upward within the unit, with the upper portion comprised of mostly silt (50.2 %) and very fine (15.7%), fine (17.7%), and medium sands (12.7%). Organic material is concentrated in the top 10 cm of the deposit, characterized by a darker color (10 YR 4/6) within this sub-layer. Organic carbon constitutes 1.86 percent of the sediment. Humic acids within the sediment yielded an accelerator radiocarbon date of 4940 $\pm$ 60 yrs B.P. (CAMS-31810). The uppermost boundary of Unit I is wavy and capped by a thin layer of coarse sediments (Figure 5:4).

Unit I is probably lacustrine in origin. Sedimentology of the deposit exhibits a classic fining-upward sequence, a common depositional progression in lacustrine settings. The concentration of organic material at the upper portion of the deposit is also consistent with lake deposition patterns. The uppermost portion of the unit is likely a gravel lag deposit. The wavy upper boundary, coupled with the presence of coarse sediments, suggests that it is an erosional surface. Following drainage of the lacustrine environment, subaerial processes probably truncated the upper surface from Unit I, leaving a gravel cap. Eolian deposition subsequently mantled the deposit with the overlying dune sands.

The surface unit (Unit II) at the Seney Sand Pit consists of 5.59 m of eolian sand (Figure 5:5A). Texture is consistent throughout the eolian profile, with the majority of sediment being medium and fine sand (Table 5:1; Figure 5:5B). Contained within Unit II is a well-developed surface soil (Figure 5:6). This soil demonstrates A-E-Bhs-Bs1-Bs2-BC-C horizonation. The C horizon has a yellowish brown color (10 YR 5/4) and a pH value of 5.6. The BC horizon is 22 cm thick and extends from 1.99-1.77 m within the exposure. This horizon has a light yellowish brown color (10 YR 6/4) and a pH value of

Unit II

4940 ± 60 yrs B.P.

Unit II

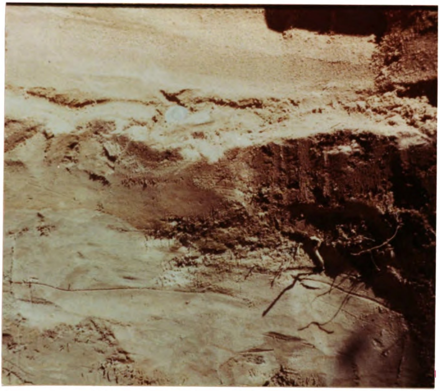


Figure 5:4 – View of wavy, organic-rich upper boundary of Unit I. This deposit is capped by a gravel lag.

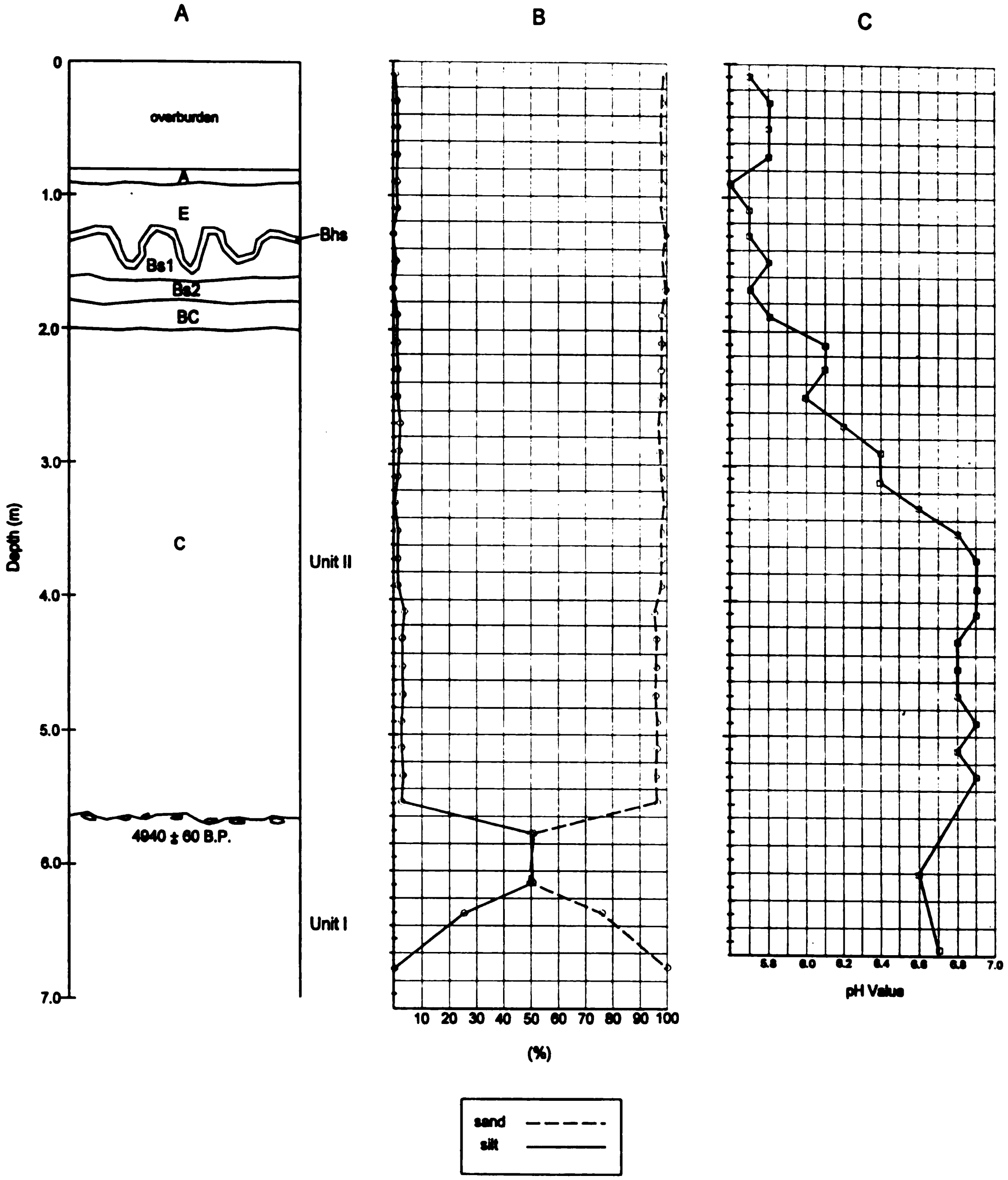


Figure 5:5 (A). Section diagram of Seney Sand Pit; (B). Sedimentology distribution of silt and sand; (C). pH values down the Seney Section.

Table 5:1 - Soil and Sediment data for the Seney Sand Pit

	color	thickness (cm)	pH	vcs	cs	ms	fs	vfs	silt	clay
A horizon	10 YR 4/1	10	4.8	0.1	1	42.2	46.9	4.5	5.3	0
E horizon	10 YR 6/3	38	5.4	0	1.5	44.5	47.7	3.9	2.4	0
Bhs horizon	5 YR 3/3	5	5.1	0	1.3	36.5	51.7	6	4.5	0
Bs1 horizon	10 YR 5/6	27	5.6	0	1.4	37	57.9	3.1	0.6	0
Bs2 horizon	10 YR 5/6	17	5.7	0	1.8	48.4	48	0.8	1	0
BC horizon	10 YR 6/4	22	5.6	0	2.4	69.5	27.1	0.7	0.3	0
C horizon	10 YR 5/4	440	5.6	0	1.9	42.2	50.9	4.1	0.9	0
Unit I (upper)	10 YR 4/6		6.6	1.4	2.3	12.7	17.7	15.7	50.2	0
Unit I (lower)	10 YR 5/3		6.7	0	0.3	37.3	55.9	6.1	0.4	0

Figure 5:6 – View of the surface soil at the Seney Sand Pit.

5.6. The Bs2 horizon is 17 cm thick and extends from 1.77-1.60 m. Horizon color is yellowish brown (10 YR 5/6) and pH is 5.7. The Bs1 horizon is 27 cm thick and extends from 1.60-1.33 m, with a yellowish brown color (10 YR 5/6) and a pH of 5.6. The Bhs horizon is 5 cm thick and extends from 1.33-1.28 m. Horizon color is dark reddish brown (5 YR 3/3) and pH is 5.1. The E horizon is 38 cm thick, ranges from 1.28-.90 m, and has a pale brown color (10 YR 6/3) and pH of 5.4. Finally, the A horizon is 10 cm thick and ranges from 0.90-0.80 m. Horizon color gray (10 YR 5/1) and pH is 4.8. A resultant POD index value of 14 indicates that the soil expresses characteristics consistent with Typic Haplorthod morphology (Schaetzl and Mokma, 1988). While buried by up to 80 cm of overburden in some areas, the surface soil is largely intact and uniform across the dune.

In summary, a detailed stratigraphic investigation was conducted at the Seney Sand Pit. This examination revealed that two, distinct depositional units exist at the site. The lower unit (Unit I) consisits of a fining-upward sequence of lacustrine sands and silts. Overlying Unit I is an 8-m thick unit (Unit II) of eolian sand, one that is contained within a parabolic dune that has a northwesterly orientation. Formed within the uppermost part of Unit II is a Typic Haplorthod that is traceable throughout the quarry.

Data derived from the Seney Sand Pit provide a record of late-Quaternary landscape evolution at the site. The sedimentology and lateral continuity of Unit I strongly suggests that the deposit is lacustrine in origin, and probably accumulated during the Nipissing stage of the ancestral Great Lakes (Lewis, 1970; Karrow, 1987). Sometime after the lake receded, Unit I was apparently truncated, as indicated by the gravel lag at the top of the deposit. A radiocarbon age of about 5000 yrs B.P. was derived from the silts within the lag matrix, and provides a minimum-limiting age for the erosion of Unit I; this date also evaluates the maximum-limiting date for the deposition of the overlying eolian sand contained within Unit II.

In the context of estimating the age of the dune at the Seney Sand Pit, a primary concern is the validity of the radiocarbon date. Although sediments have been accurately dated (by radiocarbon) in non-leaching environments such as the Great Plains (e.g., Forman and Maat, 1990; Arbogast, 1996), little is known about dating organic deposits in highly weathered environs such as Michigan. Radiocarbon dating of organics within Spodosols has been reported elsewhere (e.g., Matthews and Dresser, 1981), but some results suggest that ages are greatly underestimated. This underestimation probably occurs because organic turnover within acidic soils such as Spodosols is very rapid (e.g., Geyh *et al.*, 1983; Matthews, 1980). Nonetheless, the radiocarbon date derived from the Seney Sand Pit is believed to be an accurate maximum-limiting estimate for the development of the dune. This conclusion seems logical for two reasons: 1) about 5.6 m of eolian sand overlies the radiocarbon sample location, suggesting that contamination of the underlying silt by "younger" organics is a remote possibility; 2) sediment pH is remarkably uniform throughout Unit II, and is near neutral at the base of the dune (Figure 5:5C). Thus, translocation of "younger" organics deep within the profile seems unlikely.

Assuming that the radiocarbon age is valid, the date correlates reasonably well with the terminal part of the Altithermal, a well-documented (e.g., Baker *et al.*, 1992; Bradbury *et al.*, 1993) warm and dry interval that occurred in central North America from about 7500-4000 yrs B.P. In east-central Minnesota, Keen and Shane (1990) recognized three periods of mid- Holocene dune activity with the most recent period of mobilization transpiring at about 4900 yrs B.P. Prolonged warming and drying during this period theoretically could have destabilized the landscape at the Seney Site, causing extensive eolian activity. The radiocarbon date suggests that Unit I at the Seney Site was truncated during the earlier part of the Altithermal. Subsequently, the silts contained within Unit I were exposed for a sufficiently long period of time such that organics were incorporated within the sediment. This sediment was then buried by eolian sands, and soil formation occurred.

The northwesterly orientation of the dune limbs at the Seney Site correlates with dominant wind direction during the Altithermal, as well as other parabolic dune fields within northeastern North America (Figure 1:1). For example, Grigal *et al.* (1976) reported that a parabolic dune field in north-central Minnesota, which has northwestern orientation, formed during the Altithermal. Overall, the radiocarbon control and physical landscapes in Minnesota are remarkably consistent with the Seney Sand Pit. These data suggest a more regional scale of eolian mobilization, rather than isolated periods of dune activity.

Tahquamenon Falls Exposure

The Tahquamenon Falls site is a parabolic dune located about 2 km north of the upper falls of the Tahquamenon River and directly west of Highway 123 (Figure 5:7). Approximately 7m high, this site consists of eolian sands that overlie another depositional unit. The dune mantles a swampy environment and is anchored by primarily red pine, yellow birch, and sugar maple. An exposure along a dune limb revealed a representative sequence of deposits. Organic material located in a sub-dune deposit along the exposure

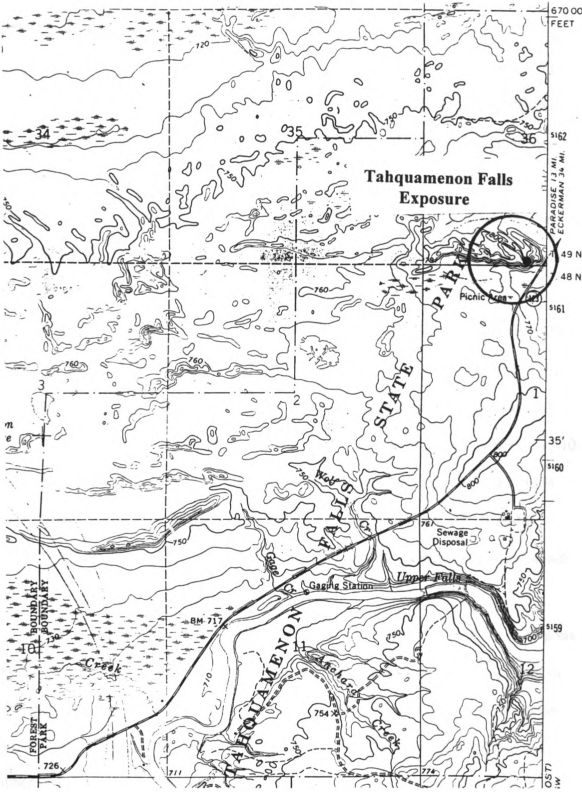


Figure 5:7 – Location of the Tahquamenon Falls Exposure (Modified from U.S.G.S. Betsy Lake South, Mich., 1:24,000, 1968).

made this site favorable for investigation (Figure 5:8).

The lowermost unit (Unit I) at the Tahquamenon Falls site extends from about 3.5 to 2.74m at the dune exposure (Figure 5:9A). Further investigation demonstrated that the basal unit is generally uniform beneath the entire dune. Particle size analysis indicates that the lower portion of the unit is composed almost entirely of sand, with individual fractions as follows: coarse sand (0.3%), medium sand (68.3%), fine sand (30.1%), very fine sand (0.9%), and silt (0.4%), respectively (Table 5:2). As with the lower unit of the Seney Sand Pit, sediments exhibit a fining-upward sequence in Unit I of the Tahquamenon Falls site. The upper portion of this unit is comprised of mostly silt (79%), with other sediment fractions as follows: coarse sand (1.0%), medium sand (10.7%), fine sand (3.6%), very fine sand (5.3%), and clay (0.4%). Organic material is concentrated within the upper 20 cm of the deposit; this sub-layer is characterized by a darker color (10 YR 4/4). Organic carbon constitutes 2.93 percent of the sediment. Humic acids within the sediment yielded an accelerator radiocarbon date of 710 $\pm$ 100 yrs B.P. (CAMS-31809). The upper boundary of Unit I is wavy, truncated, and appears to be an erosional surface.

The depositional history of Unit I at the Tahquamenon Falls site is probably analogous with that of Unit I at the Seney Sand Pit. Again, a fining-upward sequence is present in both deposits. Sediment texture and color, coupled with the presence of organic matter, is quite similar between the two sites. While a gravel lag deposit is not evident at the upper surface of Unit I, a similar sequence of erosion, brief stabilization, and eolian deposition probably occurred at the Tahquamenon Falls site.

Overlying Unit I is the surficial unit (Unit II), which consists of 2.54 m of eolian sand

Unit II

710 ± 100 yrs B.P.

Unit I

Figure 5:8 – View of the Tahquamenon Falls Exposure that shows the lowermost lacustrine deposit (Unit I) mantled by about 2.6 m of eolian sand (Unit II). Radiocarbon age determination on organic material at the contact surface yielded an age of 710 ± 100 yrs B.P.

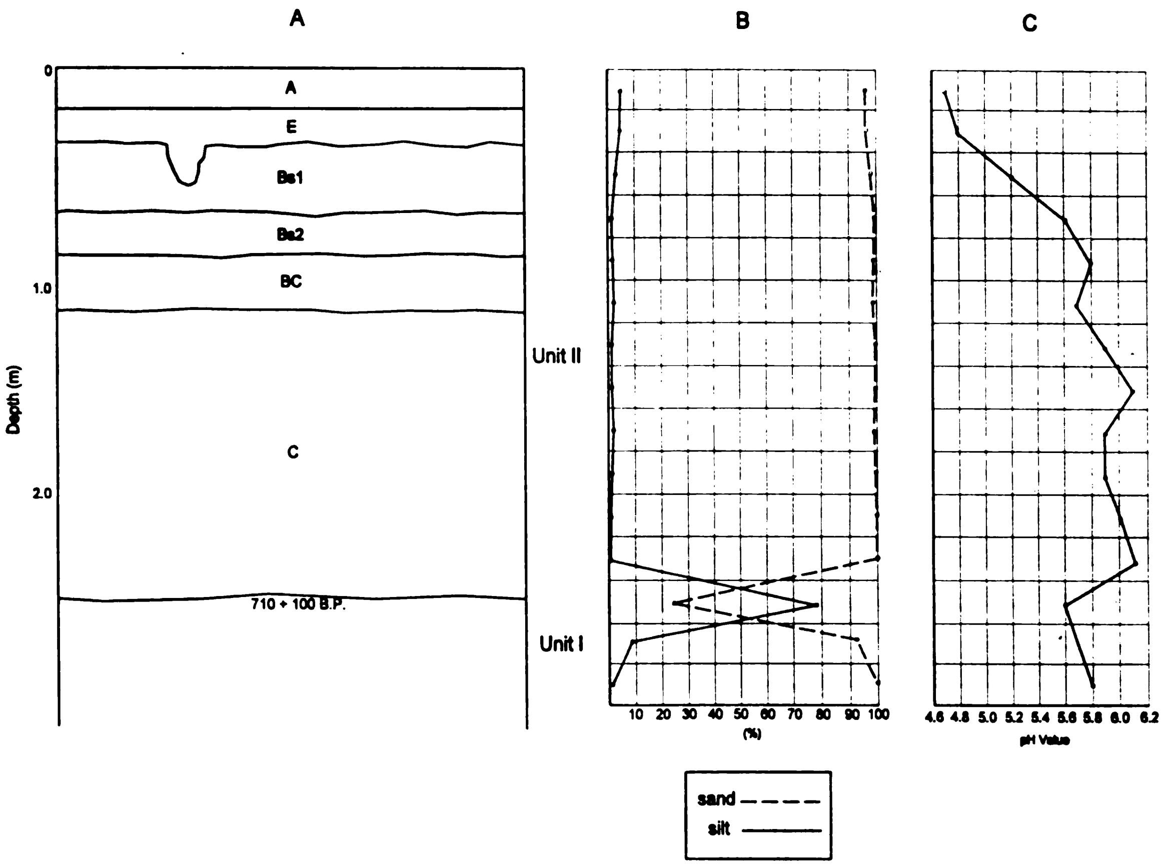


Figure 5:9 (A). Section diagram of Tahquamenon Falls Exposure; (B). Sediment distribution of silt and sand; (C). pH values down the Tahquamenon Falls Section

Table 5:2 - Soil and Sediment data for the Tahquamenon Falls Exposure

	color	thickness (cm)	pH	vcs	cs	ms	fs	vfs	silt	clay
A horizon	10 YR 3/1	20	4.3	0	1.9	54	37.3	3.3	3.5	0
E horizon	10 YR 6/2	16	4.9	0	1.6	53.3	37.9	4.3	2.9	0
Bs1 horizon	7.5 YR 4/6	33	5.2	0	0.9	51.7	39.8	6.1	1.5	0
Bs2 horizon	7.5 YR 4/6	20	6.2	0	1.9	62	31.3	3.9	0.9	0
BC horizon	7.5 YR 5/4	25	6.5	0	2.3	62.6	30.3	4	0.8	0
C horizon	10 YR 4/4	140	6	0	1.1	39.5	49.9	9.1	0.4	0
Unit I (upper)	10 YR 4/6	10	5.8	0	1	10.7	3.6	5.3	79	0.4
Unit I (middle)	10 YR 4/6	10	5.6	0	0.3	51.2	19.8	21.7	7	0
Unit I (lower)	10 YR 4/4		6	0	0.3	68.3	30.1	0.9	0.4	0.9

at the dune exposure (Figure 5:9A). Texture is consistent throughout the eolian profile, with the majority of sediment being medium and fine sands (Figure 5:9B). Contained within Unit II is a moderately developed surface soil. This soil is undisturbed and generally uniform across the dune. Soil morphology has A-E-Bs1-Bs2-BC-C horizonation (Table 5:2). The C horizon has a dark yellowish brown color (10 YR 4/4) and a pH value of 6.0. The BC horizon is 25 cm thick and extends from 1.14-0.89 m within the exposure. This horizon has a brown color (7.5 YR 5/4) and a pH of 6.5. The Bs2 horizon is 20 cm thick and ranges from 0.89-0.69 m, with a strong brown color (7.5 YR 4/6) and a pH of 6.2. The Bs1 horizon is 33 cm thick and extends from 0.69-0.36 m. Horizon color is strong brown (7.5 4/6) and pH is 5.2. The E horizon is 16 cm thick and ranges from 0.36-0.20 m, with a light brownish gray color (10 YR 6/2) and pH of 4.9. Finally, the A horizon is 19 cm thick and extends from 0.20-0.0 m, with a very dark gray color (10 YR 3/1) and pH of 4.3. These horizons and color designations render a POD index value of 8, which suggests a Typic Haplorthod classification.

In summary, a detailed stratigraphic investigation was conducted at the Tahquamenon Falls Exposure. This examination revealed that two, distinct depositional units exist at the site. The lower unit (Unit I) consists of a fining-upward sequence of sand and silt. Overlying Unit I is a 2.5 m thick unit (Unit II) of eolian sand, one that is contained within a parabolic dune that has a northwesterly orientation. Formed within the uppermost part of Unit II is a Typic Haplorthod that is traceable throughout the dune surface.

Data derived from the Tahquamenon Falls site provide a record of late Quaternary landscape evolution at this site. The sedimentology and lateral continuity of Unit I strongly suggests that the deposit is lacustrine in origin, and probably accumulated during the Nipissing stage of the ancestral Great Lakes (Lewis, 1970; Karrow, 1987). Sometime after the lake receded, Unit I was apparently truncated, as indicated by the wavy upper boundary of the deposit. A radiocarbon age of about 700 yrs B.P. was derived from silts within the lower deposit and suggests a minimum-limiting age for the erosion of Unit I. This date also provides the maximum-limiting date for the deposition of the overlying eolian sand contained within Unit II.

A date of about 700 yrs B.P. suggests that eolian stabilization at the Tahquamenon Falls site is not consistent with that proposed at the Seney site (4940 $\pm$ 60 yrs B.P.). Although spodic development *is* present at the Tahquamenon Falls site, the soil is notably less developed than at the Seney site, indicating more recent stabilization. A date of about 700 yrs B.P. coincides with a cooler, more mesic environment, which followed the proposed middle Holocene warming. A similar period (750 yrs B.P.) of eolian activity was reported in forested dune fields of Quebec, Canada (Figure 1:1; Filion, 1984). Eolian activity was attributed to increased wildfire frequencies that were triggered by a regional climatic cooling. "Catastrophic" fires within the region devegetated the landscape, allowing dunes to reactivate.

Wildfire occurrences have also been reported in the eastern upper peninsula of Michigan. Jenkins (1943) studied logging field notes to determine the natural cover of northern Michigan and found that nearly every discussion made some mention of fire in the mixed forest-marsh areas. He noted that fire frequencies were highest in "sizable areas of Grayling sand." In 1976, a lightning-induced wildfire burned 30,000 hectares of land, mostly within the Seney National Wildlife Refuge, about 20 km west of the study area (Figure 5:10). A subsequent study by Anderson (1982) showed that the fire burned

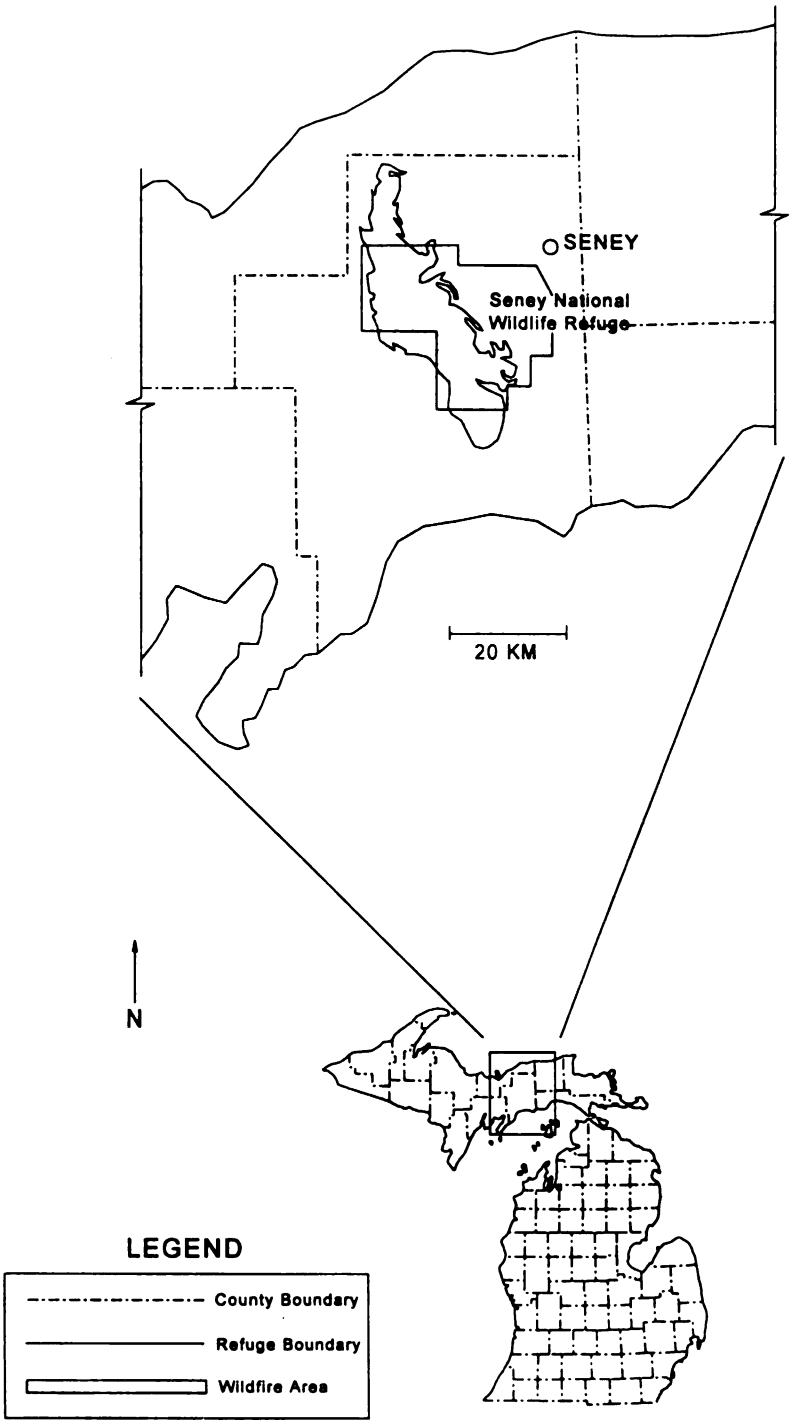


Figure 5:10 - Location and perimeter of 1976 Seney National Wildlife Refuge wildfire. (Modified from Anderson, 1982).

with varying degrees of intensity, leaving some areas with little vegetation effected, while all herbaceous and woody vegetation were consumed in other locales. Using documentation from the refuge, Anderson found that lightning fires occurred roughly once a decade, usually within similar locations and during periods of low water levels. Mineral soils sampled three years after the fire showed that virtually no differences existed between burned and unburned plots that remained stable.

Recent mobilization (750 yrs B.P.) of parabolic dunes due to wildfire has been suggested cold climates (Filion, 1984). Further, fire records demonstrate that wildfires were widespread in the upper peninsula of Michigan (Jenkins, 1943) and common just west of the study area (Anderson, 1982). As such, the reactivation of the Tahquamenon Falls dune due to wildfire seems possible.

As with the Seney Sand Pit, the validity of the radiocarbon date at the Tahquamenon Falls Site must be considered. Dune size, orientation and sedimentology, as well as the surrounding landscape, are reasonably consistent between the Seney and Tahquamenon Falls sites. The radiocarbon ages, however, do not correlate. The radiocarbon sample at the Seney site was collected roughly under the dune crest, where it was mantled by about 5.6 m of eolian sand. Conversely, the Tahquamenon Falls radiocarbon sample was collected along a dune limb, where it was buried by about 2.5 m of eolian sand. Consequently, the Tahquamenon Falls sample was more susceptible to contamination of modern carbon by water infiltration. Indeed, sediment pH at the Tahquamenon Falls site indicates that pH levels have not yet stabilized above the radiocarbon sample location and the sediment is still notably acidic (Figure 5:9C). Further, the standard deviation of the Tahquamenon Falls sample (700 $\pm$ 100 yrs B.P.) suggests a greater variation in the age of organics within the sample, as opposed to the Seney Site (4940 $\pm$ 60 yrs B.P.) which has a much smaller deviation.

Another factor that must be considered is the likelihood that a series of events occurred within the time frame that a date of about 700 yrs B.P. allows. If the date is valid, an erosional event truncated the upper surface of Unit I, roughly 2.5 m of eolian sand subsequently buried the deposit, the dune then stabilized, and the surface soil subsequently developed within roughly 700 years. While wildfire occurrence could explain dune reactivation, the degree of soil development, coupled with the absence of charcoal, make the probability of wildfire influence less likely.

While the uncertainty of radiocarbon control at the Tahquamenon Falls site may preclude absolute age control, some generalizations can be made given the physical characteristics of the dune. For example, the dune has a northwesterly orientation, is devoid of discontinuities within the upper unit (Unit II), and overlies lacustrine sediments. This evidence suggests that dune formation occurred over one, broad depositional event, when northwesterly winds prevailed. Also, given that Unit II overlies lacustrine sediments and is currently located in a swampy landscape, dune mobilization must have occurred during a drier environment than exists today. Although similarities exist between the sequence of deposits at the Seney and Tahquamenon Falls sites, a less developed surface soil at the Tahquamenon Falls site suggests that the timing of dune stabilization may not have been uniform throughout the study area.

Soil Study Transect

Differing radiocarbon age determinations and soil development at the Seney Sand Pit and Tahquamenon Falls sites suggest that eolian activity within the study area was not uniform. Due to inconsistencies between the two sites, estimations of exactly when dune activity occurred are difficult. In an effort to further clarify this issue, 10 pedons on dunes were examined along a transect connecting the two, end-member sites (Figure 5:11). Surface soils were used as measures of relative surface exposure. Morphologic and chemical tests for each pedon included: horizon thickness, color, texture, pH, and Fe and Al extractions (Table 5:3). POD index values were also calculated for each site and were used as indicators of soil development. Resultant data were used to suggest both the relative age of dunes and spatial patterns of eolian activity within the study area.

Transect Site 1

Site 1 is located on a large dune immediately north of Highway M-28 and approximately 2.5 km east of the Seney Sand Pit (Figure 5:12). The dune mantles an extensive wetland area and is stabilized by a mixed forest assemblage. Dune limbs are orientated to the northwest and the dune is about 9 m in relief.

The surface soil on the crest of Site 1 is a well-developed Spodosol that contains A-E-Bhs-Bs1-Bs2-BC-C horizonation (Figure 5:13). All horizons are texturally classified as sand, with fine sand being the dominant component. The C horizon begins at a depth of 1.17 m, is yellowish brown (10 YR 5/4) and has a pH of 6.5 (Table 5:3). Overlying the C horizon is a 28-cm thick (1.17-0.89 m) BC horizon, which is strong brown (7.5 YR 4/6) and has a pH of 5.1. The Bs2 horizon extends from 0.89-0.71 m. This horizon is also

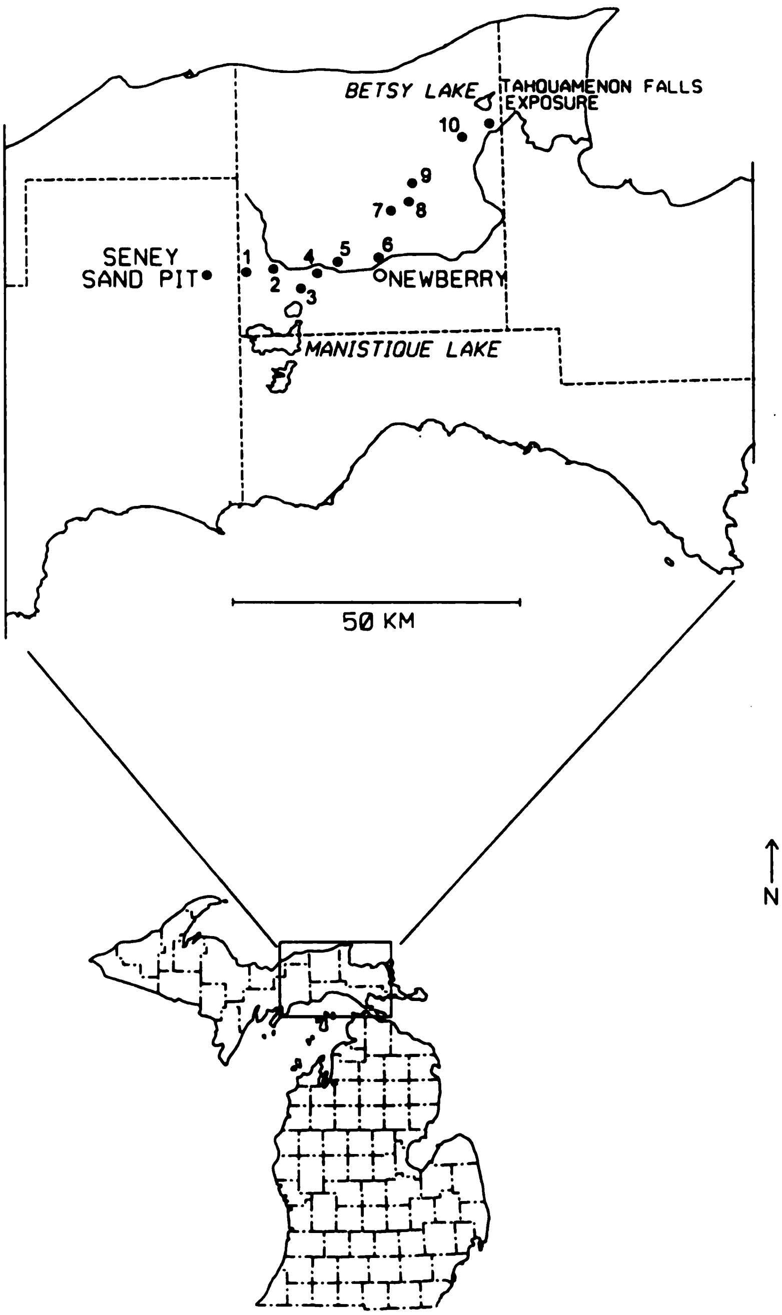


Figure 5:11 - Study transect site locations (Modified from MIRIS base map, 1998).

Site	Horiz.	Color ID	POD	Thick	pH	%Feo	%Fed	Wtd. Feo	Wtd. Fed	%Alo	%Ald	Wtd. Alo	Wtd. Ald	Texture	vcs	cs	ms	fs	vfs	silt	clay
1	A	10 YR 3/1	22	19	5									fine sd	0	1.5	35.2	54.6	4.9	3.8	0
1	E	10 YR 6/3		22	5.1									fine sd	0.1	1.7	31.7	57.8	5.9	2.8	0
1	Bhs	5 YR 3/4		10	5.3	0.096	0.209	0.96	2.09	0.159	0.241	1.59	2.41	fine sd	0	0.9	27.9	61.3	6.3	3.6	0
1	Bs1	5 YR 4/6		20	5.1	0.053	0.133	1.06	2.66	0.251	0.382	5.02	7.64	fine sd	0.1	1.5	35.7	56.4	4.6	1.7	0
1	Bs2	7.5 YR 4/6		18	5.6	0.026	0.052	0.468	0.936	0.085	0.179	1.53	3.222	sand	0.3	8.1	50.7	37.9	1.7	1.3	0
1	BC	7.5 YR 4/6		28	6.1	0.015	0.031	0.42	0.868	0.045	0.114	1.26	3.192	fine sd	0.2	3.6	44.1	48.7	2.4	1	0
1	C	10 YR 5/4		0	6.5	0.013	0.018	0	0	0.038	0.064			fine sd	0.7	6.3	44.9	44.3	2.4	1.4	0
2	A	10 YR 3/1	6	15	4.3									sand	0.2	4.8	51	36.4	3.7	3.9	0
2	E	10 YR 6/2		21	5.1									sand	0	4.7	46.9	41.4	4.5	2.5	0
2	Bs1	7.5 YR 4/6		31	5.8	0.104	0.134	3.224	4.154	0.159	0.429	4.929	13.299	fine sd	0.1	2.9	38.5	50.1	6.6	1.8	0
2	Bs2	7.5 YR 5/4		22	5.9	0.048	0.077	1.056	1.694	0.077	0.238	1.694	5.236	fine sd	0	1.8	43.6	49.9	3.9	0.8	0
2	BC	10 YR 5/4		25	5.8	0.018	0.017	0.45	0.425	0.024	0.06	0.6	1.5	fine sd	0	3.1	40.4	47.6	8.3	0.6	0
2	C	10 YR 5/3		0	6	0.01	0.008	0	0	0.011	0.028			fine sd	0	2.5	42.5	47.7	6.8	0.5	0
3	A	10 YR 3/1	20	3	4.9									fine sd	0	1.3	35.6	52.5	7.7	2.9	0
3	E	10 YR 6/2		17	5.1									fine sd	0	1.6	39.3	49.8	6.4	2.9	0
3	Bhs	5 YR 3/3		26	4.9	0.075	0.2	1.95	5.2	0.182	0.279	4.732	7.254	sand	0	2.9	50	39.9	3.5	3.7	0
3	Bs1	7.5 YR 4/6		23	5.3	0.034	0.089	0.782	2.047	0.16	0.267	3.68	6.141	sand	0	2.2	47.9	43.5	4.5	1.9	0
3	Bs2	7.5 YR 4/6		15	5.2	0.038	0.063	0.57	0.945	0.195	0.234	2.925	3.51	fine sd	0	2.6	44.5	45.7	5.3	1.9	0
3	BC	10 YR 5/6		30	5.6	0.012	0.024	0.36	0.72	0.05	0.081	1.5	2.43	sand	0.1	4	54.7	36.9	3.3	1	0
3	C	10 YR 5/6		0	6.6	0.01	0.017	0	0	0.035	0.061			sand	0	2.1	55.7	37.7	3.6	0.9	0
4	A	10 YR 2/1	12	14	4.2									fine sd	0.1	1.4	40.1	46.2	4.8	7.4	0
4	E	10 YR 5/2		23	4.5									fine sd	0	1.4	41.1	49.1	6	2.4	0
4	Bhs	5 YR 3/2		15	4.7	0.066	0.17	0.99	2.55	0.111	0.167	1.665	2.505	fine sd	0	1.2	41.4	49.4	4.7	3.3	0
4	Bs1	7.5 YR 4/6		23	5.1	0.036	0.075	0.828	1.725	0.155	0.216	3.565	4.968	fine sd	0	1.5	43.7	47.1	4.7	3	0
4	Bs2	7.5 YR 4/6		15	5.3	0.024	0.056	0.36	0.84	0.109	0.166	1.635	2.49	fine sd	0	1.3	37.5	51.7	6.3	3.2	0
4	BC	10 YR 4/6		21	5.4	0.018	0.032	0.378	0.672	0.077	0.113	1.617	2.373	sand	0	3.4	50.5	41.5	3.2	1.4	0
4	C	10 YR 5/4		0	5.7	0.012	0.021	0	0	0.037	0.057			fine sd	0	1.2	40.1	53.7	3.9	1.1	0
5	A	10 YR 3/1	2	13	5									fine sd	0	0.2	14.5	65.9	15.5	3.9	0
5	E	10 YR 5/2		16	5.2									fine sd	0.1	1.1	18.3	63.7	13.3	3.5	0
5	Bs1	7.5 YR 4/6		26	5.8	0.052	0.136	1.352	3.536	0.089	0.226	2.314	5.876	fine sd	0	0.2	24.5	63.8	10.6	0.9	0
5	Bs2	10 YR 5/6		19	5.7	0.031	0.074	0.589	1.406	0.054	0.166	1.026	3.154	fine sd	0	0	18	70.6	10.7	0.7	0
5	BC	10 YR 5/6		23	6.1	0.016	0.023	0.368	0.529	0.037	0.088	0.851	2.024	fine sd	0	0	18.8	68.3	12.2	0.7	0
5	C	10 YR 6/4		0	5.9	0.024	0.02	0	0	0.03	0.061			fine sd	0	0	23.7	66.1	9.3	0.9	0

Table 5:3 - Soil and Sediment Data for Transect Sites

Site	Horiz.	Color ID	POD	Thick	pH	%Feo	%Fed	Wtd. Feo	Wtd. Fed	%Alo	%Ald	Wtd. Alo	Wtd. Ald	texture	vcs	cs	ms	fs	vfs	silt	clay
6	A	10 YR 2/1	4	10	4.8									sand	0	6.3	59.2	25.7	5	3.8	0
6	E	10 YR 4/2		18	4.8									sand	0	5.9	61	26.4	3.5	3.2	0
6	Bhs	5 YR 3/3		8	5.1	0.059	0.129	0.472	1.032	0.1	0.128	0.8	1.024	sand	0.2	12.2	61.5	20.3	2.7	3.1	0
6	Bs1	7.5 YR 4/6		28	5.4	0.017	0.041	0.476	1.148	0.085	0.167	2.38	4.676	sand	0	5.4	58.4	30.7	3.8	1.7	0
6	Bs2	10 YR 4/4		20	5.8	0.013	0.027	0.26	0.54	0.045	0.071	0.9	1.42	sand	0	1.7	56.5	35.4	5.3	1.1	0
6	BC	10 YR 5/4		18	6	0.009	0.018	0.162	0.324	0.038	0.07	0.684	1.26	sand	0	2.5	61.8	32.3	2.1	1.3	0
6	C	10 YR 5/4		0	6.8	0.012	0.016	0	0	0.025	0.035			sand	0	8.9	67.5	20.3	2.1	1.2	0
7	A	10 YR 3/1	2	8	4.8									fine sd	0	0.2	21.2	61.3	12.1	5.2	0
7	E	10 YR 5/3		16	5.2									fine sd	0	0.3	20	63.6	13	3.1	0
7	Bs1	7.5 YR 4/6		36	5.8	0.051	0.118	1.836	4.248	0.078	0.256	2.808	9.216	fine sd	0	0	21.5	65.1	12.7	0.7	0
7	BC	7.5 YR 4/6		33	5.9	0.022	0.048	0.726	1.584	0.033	0.149	1.089	4.917	fine sd	0	0	27.2	62	9.1	1.7	0
7	C	10 YR 5/4		0	6	0.017	0.023	0	0	0.02	0.054			fine sd	0	0.2	33.5	53.1	12.5	0.7	0
8	A	5 YR 2.5/1	14	13	4.5									fine sd	0	1.4	29.8	57.8	8.3	2.7	0
8	E	10 YR 5/2		23	4.6									fine sd	0	1.3	28.8	55.5	11.9	2.5	0
8	Bhs	5 YR 2.5/2		6	4.7	0.038	0.13	0.228	0.78	0.124	0.202	0.744	1.212	fine sd	0	1.9	38.4	46.5	8.5	4.7	0
8	Bs1	7.5 YR 4/6		23	4.9	0.055	0.151	1.265	3.473	0.298	0.422	6.854	9.706	fine sd	0	1.7	31.7	55.7	8.4	2.5	0
8	Bs2	7.5 YR 4/6		23	5	0.035	0.095	0.805	2.185	0.182	0.278	4.186	6.394	fine sd	0	0.7	19.3	60.9	15.8	3.3	0
8	BC	7.5 YR 5/4		25	5.7	0.012	0.022	0.3	0.55	0.051	0.079	1.275	1.975	fine sd	0	3.4	46.3	43.6	5.7	1	0
8	C	10 YR 5/4		0	6.5	0.008	0.016	0	0	0.032	0.051			fine sd	0	0.2	39.8	51.1	8.3	0.6	0
9	A	10 YR 4/1	18	4	4.8									fine sd	0	2.5	37	49.3	8.1	3.1	0
9	E	10 YR 6/2		23	4.8									fine sd	0	3.9	40.9	46.4	5.9	2.9	0
9	Bhs	5 YR 2.5/2		5	5.5	0.045	0.101	0.225	0.505	0.153	0.235	0.765	1.175	fine sd	0	2.7	44.5	44	4.8	4	0
9	Bs1	7.5 YR 5/6		23	5.3	0.035	0.071	0.805	1.633	0.082	0.349	1.886	8.027	fine sd	0	1	21.3	61.6	14.9	1.2	0
9	Bs2	7.5 YR 5/6		28	5.7	0.026	0.049	0.728	1.372	0.068	0.214	1.904	5.992	fine sd	0	1.9	36.5	53.7	7.2	0.7	0
9	BC	7.5 YR 5/6		20	6	0.014	0.035	0.28	0.7	0.03	0.143	0.6	2.86	fine sd	0	1.5	35.6	53.2	8.8	0.9	0
9	C	10 YR 5/6		0	6.5	0.015	0.028	0	0	0.041	0.09			sand	0	1.8	54.1	37.8	4.8	1.5	0
10	A	10 YR 2/1	10	28	4.3									fine sd	0.2	0.8	7.5	59.3	22.8	9.4	0
10	E	10 YR 5/2		31	4.3									fine sd	0	0.5	3.1	58.1	34.4	3.9	0
10	Bhs	5 YR 3/3		7	4.4	0.131	0.311	0.917	2.177	0.264	0.445	1.848	3.115	fine sd	0	1.3	6.8	67.6	17.7	6.6	0
10	Bs1	7.5 YR 4/6		33	5	0.037	0.074	1.221	2.442	0.185	0.332	6.105	10.956	fine sd	0	0	1.9	74.5	22.6	1	0
10	BC	7.5 YR 5/6		18	5.4	0.028	0.028	0.504	0.504	0.041	0.088	0.738	1.584	fine sd	0	0	1.3	71.7	26.7	0.3	0
10	C	10 YR 5/4		0	5.4	0.026	0.043	0	0	0.033	0.079			fine sd	0	0	1.4	72.8	24.1	1.7	0

Table 5:3 (cont.) - Soil and Sediment Data for Transect Sites

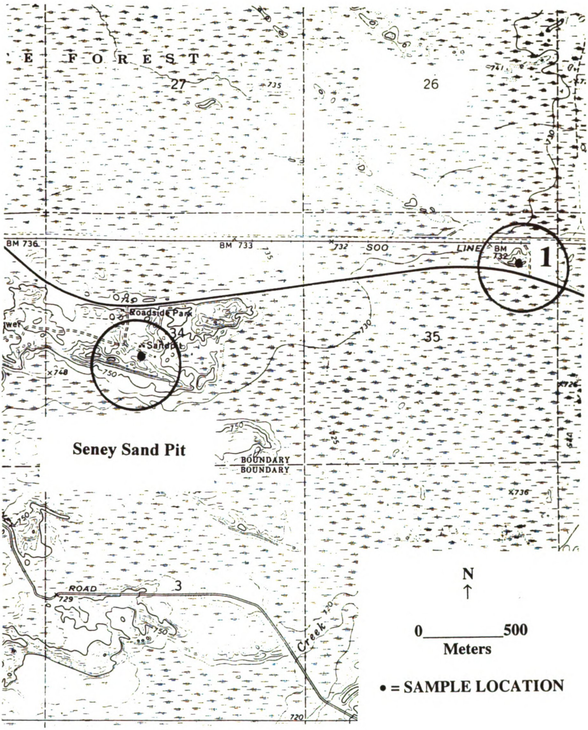


Figure 5:12 - Location of Transect Site 1 (Modified from U.S.G.S. Seney, Mich., 1:24,000, 1972).

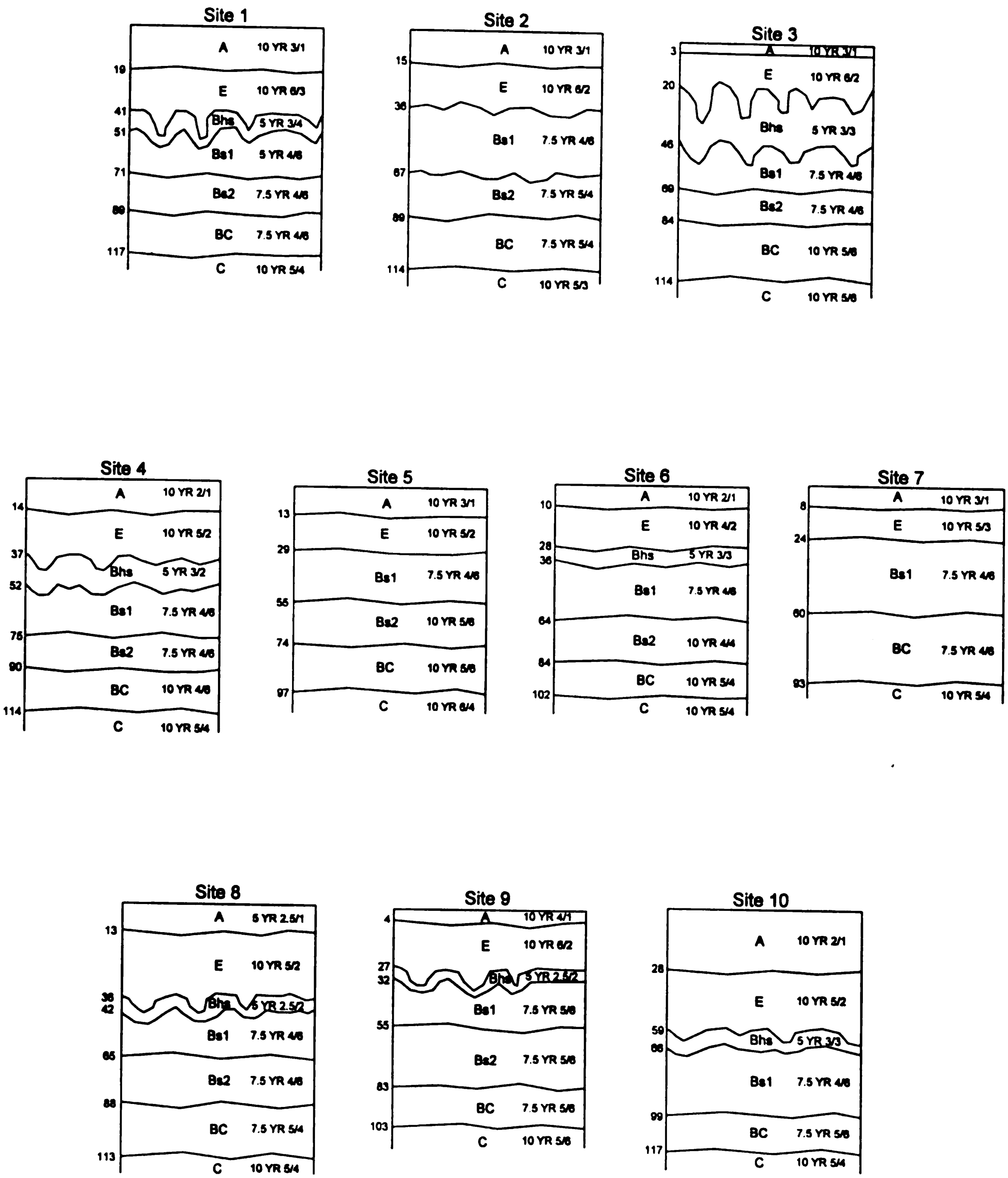


Figure 5:13 - Soil Profiles at Transect Site Locations

strong brown (7.5 YR 4/6) and is more acidic than the underlying horizons (pH = 5.6). Capping the Bs2 is the Bs1, which extends from 0.71-0.51 m. The Bs1 is yellowish red (5 YR 4/6) and is more acidic (pH = 5.1) than the underlying Bs2. The uppermost B horizon is a 10-cm thick Bhs (0.51-0.41 m), which is dark reddish brown (5 YR 3/4) and has a pH of 5.3. Each of the B horizons are well cemented with ortstein, which is reflected by the weighted Fe extractions of 5.69 (Fe_d) and 2.49 (Fe_o) derived from the group. The E horizon is 22-cm thick (0.41-0.19 m). The pale brown color (10 YR 6/3) of the horizon, as well as the acidic pH value (5.1), reflect the eluviation that has occurred from this part of the soil. Capping the soil is a 19-cm thick, very dark gray (10 YR 3/1) A horizon; this horizon is the most acidic of all the horizons at the site, with a pH of 5.0. Overall, this soil is extremely well developed. In that context, the calculated POD value for the soil is 22, which indicates a Spodosol of Typic Haplorthod morphology (Schaetzl and Mokma, 1988).

Transect Site 2

Site 2 is located on a dune 3.0 km north of Highway M-28 and approximately 1.5 km northeast of Wanamaker Lake (Figure 5:14). Mature forest anchors the dune, consisting of primarily red pine, spruce, and poplar. Dune limbs are orientated to the northeast and the deposit is about 3 m in relief.

The surface soil on the crest of Site 2 is a Spodosol that contains A-E-Bs1-Bs2-BC-C horizonation (Figure 5:13). All horizons are texturally classified as sand, with fine sand being the dominant component. The C horizon begins at a depth of 1.14 m, is brown (10 YR 5/3) and has a pH of 6.0 (Table 5:3). Overlying the C horizon is a 25-cm thick (1.14-

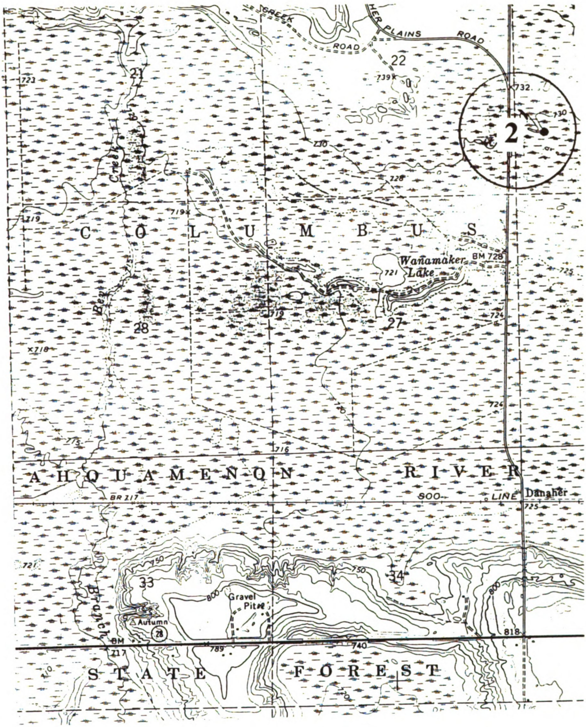


Figure 5:14 - Location of Transect Site 2 (Modified from U.S.G.S. Hardwood Island, Mich., 1:24,000, 1973).

0.89 m) BC horizon, which is brown (7.5 YR 5/4) and has a pH of 5.8. The Bs2 horizon extends from 0.89-0.67 m. This horizon is also brown (7.5 YR 5/4) and has a pH value of 5.9. The Bs1 horizon extends from 0.67-0.36 m. This horizon is strong brown (7.5 YR 4/6) and more acidic (pH = 5.8) than the underlying Bs2. All B horizons are well cemented, with weighted Fe extractions of 5.85 (Fe_d) and 4.28 (Fe_o). The E horizon is 21-cm thick (0.36-0.15 m), has a light brownish gray color (10 YR 6/2), and a pH value of 5.1. Capping the soil is a 15-cm thick, very dark gray (10 YR 3/1) A horizon. This horizon is the most acidic of all the horizons at the site, with a pH of 4.3. Overall, this soil is moderately developed. A POD index value of 6 suggests that the soil has Entic Haplorthod-like morphology.

Transect Site 3

Site 3 is located on a large dune 0.5 km south of Highway M-28 and approximately 0.5 km west of East Lake (Figure 5:15). The dune is situated on an upland area surrounded by pockets of low-lying wetlands. Forest cover consists of a mature, mixed assemblage of birch, sugar maple, spruce, and poplar. Dune limbs are orientated to the west and the deposit is about 6 m in thickness.

The surface soil on the crest of Site 3 is a well developed Spodosol that contains A-E-Bhs-Bs1-Bs2-BC-C horizonation (Figure 5:13). All horizons are texturally classified as sand, with fine sand being the dominant component. The C horizon begins at a depth of 1.14 m, is yellowish brown (10 YR 5/6) and has a pH of 6.6 (Table 5:3). Overlying the C horizon is a 30-cm thick (1.14-0.84 m) BC horizon, which is yellowish brown (10 YR 5/6) and has a pH of 5.6. The Bs2 horizon extends from 0.84-0.69 m. This horizon is

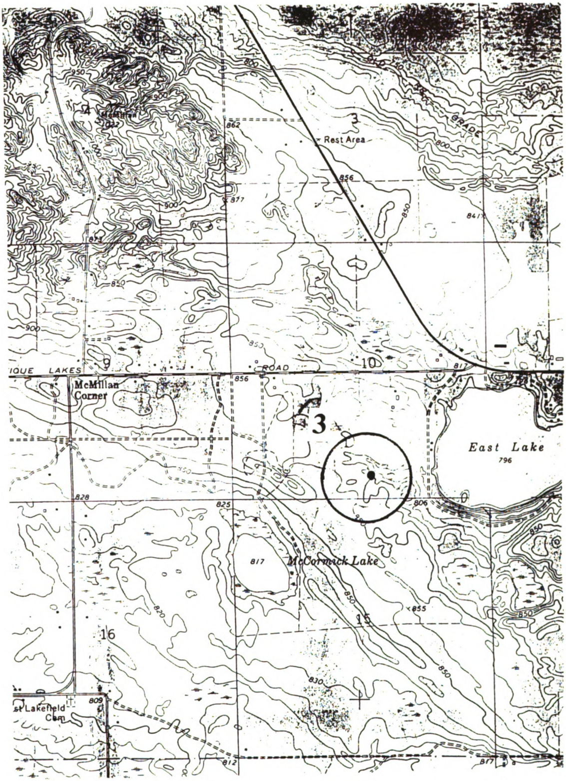


Figure 5:15 - Location of Transect Site 3 (Modified from U.S.G.S. McMillan, Mich., 1:24,000, 1973).

strong brown (7.5 YR 4/6) and is more acidic than the underlying horizons (pH = 5.2). The Bs1, which extends from 0.69-0.45 m, has a strong brown color (7.5 YR 4/6) and a pH value of 5.3. The uppermost B horizon is a 26-cm thick Bhs (0.46-0.20 m), which is dark reddish brown (5 YR 3/3) and has a pH of 4.9. Each of the B horizons was well cemented with ortstein concretions, which is reflected by the weighted Fe extractions of 8.19 (Fe_d) and 3.30 (Fe_o) derived from the horizons. The E horizon is 17-cm thick (0.20-0.03 m), has a light brownish gray color (10 YR 6/2), and a pH value of 5.1. Capping the soil is a thin, 3-cm thick A horizon. This horizon has a very dark gray (10 YR 3/1) color and a pH value of 4.9. As a whole, this soil is extremely well developed. In that context, the calculated POD value for the soil is 20, which suggests that it would classify as a Typic Haplorthod.

Transect Site 4

Site 4 is located on a large dune immediately west of County Road 33 and approximately 0.5 km east of North Manistique Lake (Figure 5:16). The dune mantles a patchy wetland environment and is stabilized by a sugar maple and red pine assemblage. Dune limbs are orientated to the west and the deposit is about 9 m in relief.

The surface soil on the crest of Site 4 is a well developed Spodosol that contains A-E-Bhs-Bs1-Bs2-BC-C horizonation (Figure 5:13). All horizons are texturally classified as sand, with fine sand being the dominant component. The C horizon begins at a depth of 1.14 m, is yellowish brown (10 YR 5/4) and has a pH of 5.7 (Table 5:3). Overlying the C horizon is a 21-cm thick (1.14-0.90 m) BC horizon, which is dark yellowish brown (10 YR 4/6) and has a pH of 5.4. The Bs2 horizon extends from 0.90-0.75 m. This horizon

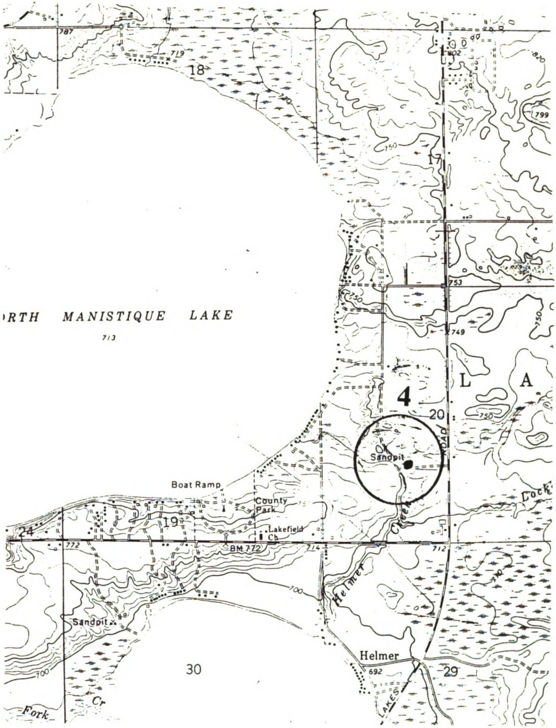


Figure 5:16 - Location of Transect Site 4 (Modified from U.S.G.S. McMillan, Mich., 1:24,000, 1973).

is strong brown (7.5 YR 4/6) and has a pH of 5.3. Overlying the Bs2 is the Bs1, which extends from 0.75-0.52 m. The Bs1 is also strong brown (7.5 YR 4/6) and is more acidic (pH = 5.1) than the underlying Bs2. The uppermost B horizon is a 15-cm thick Bhs (0.52-0.37 m), which is dark reddish brown (5 YR 3/2) and has a pH of 4.7. Each of the B horizons possesses moderate cementation. The total weighted Fe extractions for all B sub-horizons are 5.11 (Fe_d) and 2.18 (Fe_o). The E horizon is 23-cm thick (0.37-0.14 m), has a grayish brown color (10 YR 5/2), and a pH value of 4.5. Capping the soil is a 14-cm thick, black (10 YR 2/1) A horizon. This horizon is the most acidic of all the horizons at the site, with a pH of 4.2. Overall, this soil is well developed and has POD Index of 12.

Transect Site 5

Site 5 is located on a large dune 2.0 km northwest of Newberry and 1.5 km north of Dollarville (Figure 5:17). The dune mantles a low-lying wetland area and is stabilized by a mixed yellow birch, spruce, and sugar maple assemblage. Dune limbs are orientated to the west and the dune deposit is about 6 m in relief.

The surface soil on the crest of Site 5 is a Spodosol that contains A-E-Bs1-Bs2-BC-C horizonation (Figure 5:13). All horizons are texturally classified as sand, with fine sand being the dominant component. The C horizon begins at a depth of 0.97 m, is light yellowish brown (10 YR 6/4) and has a pH of 5.9 (Table 5:3). Overlying the C horizon is a 23-cm thick (0.97-0.74 m) BC horizon, which is dark yellowish brown (10 YR 4/6) and has a pH of 6.1. The Bs2 horizon extends from 0.97-0.74 m. This horizon is yellowish brown (10 YR 5/6) and is more acidic than the underlying horizons (pH = 5.7). The Bs1 horizon, which extends from 0.55-0.29 m, is strong brown (7.5 YR 4/6) and has a pH

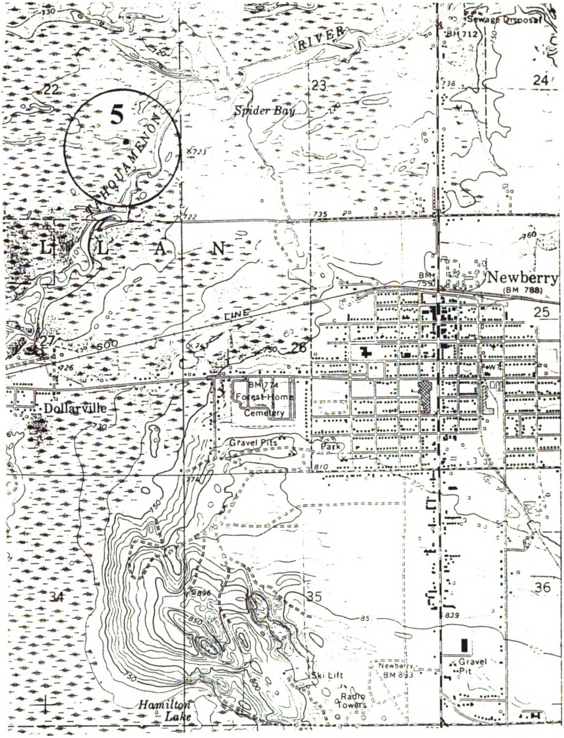


Figure 5:17 - Location of Transect Site 5 (Modified from U.S.G.S. Newberry, Mich., 1:24,000, 1973).

value of 5.8. No cementation was evident within the B horizons and weighted Fe extractions of 4.94 (Fe_d) and 1.94 (Fe_o) derived from the group. The E horizon is 16-cm thick (0.29-0.13 m), has a grayish brown color (10 YR 5/2), and a pH value of 5.2). Capping the soil is a 13-cm thick, very dark gray (10 YR 3/1) A horizon, which has a pH of 5.0. As a whole, this soil has weak development. The calculated POD value for the soil is 2, suggesting Entic Haplorthod morphology.

Transect Site 6

Site 6 is located on a large dune 1.5 km east of the Luce County Airport (Figure 5:18). The dune mantles an outwash deposit on the fringe of an extensive wetland landscape and is anchored by a dense cover of mature sugar maple and balsam fir. Dune limbs are orientated to the west and the deposit is about 6 m in relief.

The surface soil on the crest of Site 6 is a moderately developed Spodosol that contains A-E-Bhs-Bs1-Bs2-BC-C horizonation (Figure 5:13). All horizons are texturally classified as sand, with medium sand being the dominant component. The C horizon begins at a depth of 1.02 m, is yellowish brown (10 YR 5/4) and has a pH of 6.8 (Table 5:3). Overlying the C horizon is a 18-cm thick (1.02-0.84 m) BC horizon, which is also yellowish brown (10 YR 5/4) and has a pH of 6.0. The Bs2 horizon extends from 0.84-0.64 m. This horizon is dark yellowish brown (10 YR 4/4) and has a pH of 5.8. Overlying the Bs2 is the Bs1, which extends from 0.64-0.36 m. The Bs1 is strong brown (7.5 YR 4/6) and is more acidic (pH = 5.4) than the underlying Bs2. The uppermost B horizon is an 8-cm thick Bhs (0.36-0.28 m), which is dark reddish brown (5 YR 3/3) and has a pH of 5.1. Each of the B horizons lacked cementation and weighted Fe extractions

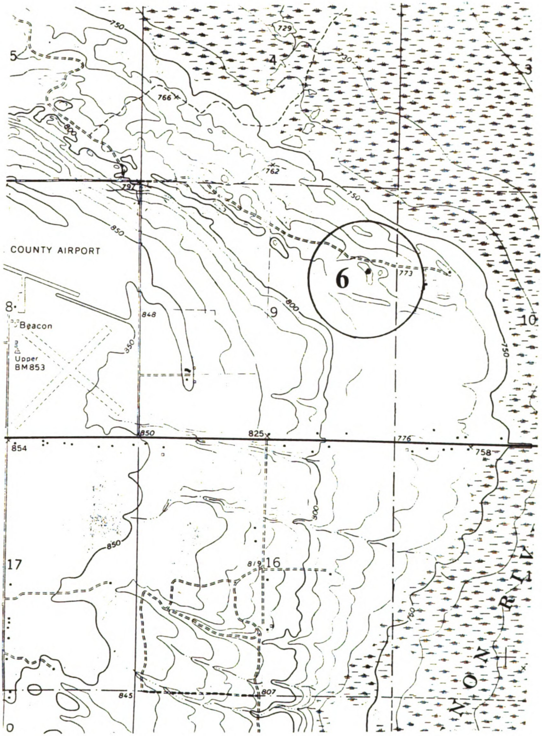


Figure 5:18 - Location of Transect Site 6 (Modified from U.S.G.S. Roberts Corner, Mich., 1:24,000, 1973).

of 2.72 (Fe_d) and 1.21 (Fe_o) are derived from the group. The E horizon is 18-cm thick (0.28-0.10 m) and has a dark grayish brown color (10 YR 4/2). The pH value for this horizon is 4.8. Capping the soil is a 10-cm thick, black (10 YR 2/1) A horizon, which also has a pH value of 4.8. Overall, this soil is weakly developed. Similarities in color between B horizons rendered a relatively low POD index value of 4, which indicates Entic Haplorthod morphology.

Transect Site 7

Site 7 is located on a large dune immediately north of Highway 123 and approximately 1.5 km south of Murphy Creek (Figure 5:19). The dune mantles an extensive wetland area and is stabilized by a mixed sugar maple and red pine assemblage. Dune limbs are orientated to the northwest and the dune has about 9 m of relief.

The surface soil on the crest of Site 7 is a weakly developed Spodosol that contains A-E-Bs1-BC-C horizonation (Figure 5:13). All horizons are texturally classified as sand, with fine sand being the dominant component. The C horizon begins at a depth of 0.93 m, is yellowish brown (10 YR 5/4) and has a pH of 6.0 (Table 5:3). Overlying the C horizon is a 33-cm thick (0.93-0.60 m) BC horizon, which is strong brown (7.5 YR 4/6) and has a pH of 5.9. Capping the BC is the Bs1, which extends from 0.60-0.24 m. The Bs1 is strong brown (7.5 YR 4/6) and has a pH value of 5.8. The Bs1 horizon lacks cementation, which is reflected by the weighted Fe extractions of 4.248 (Fe_d) and 1.836 (Fe_o) derived from the horizon. The E horizon is 16-cm thick (0.24-0.08 m), has a brown color (10 YR 5/3) of the horizon, and a pH value of 5.2. Capping the soil is an 8-cm thick, very dark gray (10 YR 3/1) A horizon. This horizon is the most acidic of all the

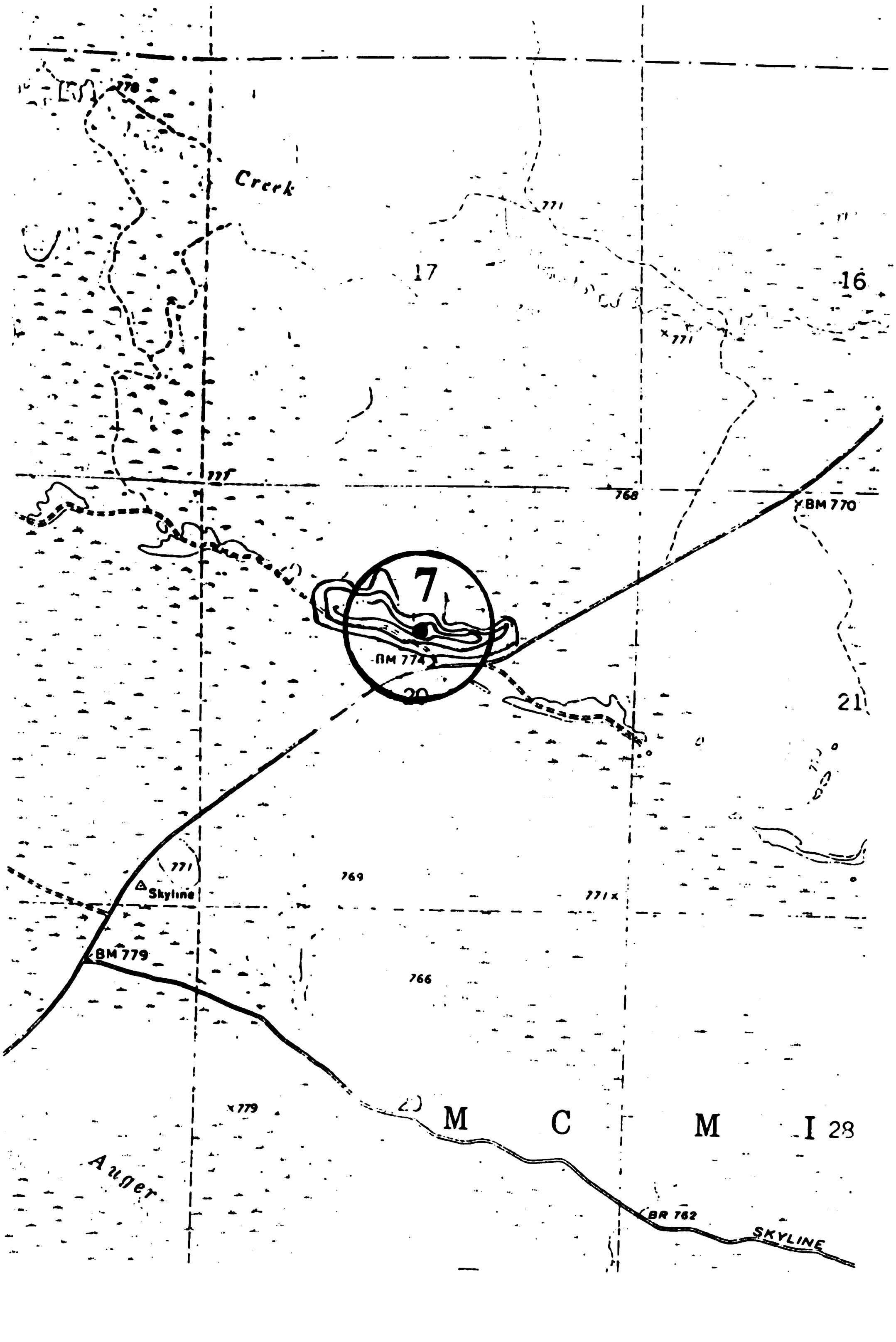


Figure 5:19 - Location of Transect Site 7 (Modified from U.S.G.S. Roy Lake, Mich., 1:24,000, 1973).

horizons at the site, with a pH of 4.8. This soil expresses the weakest development in the study transect. A POD index value of 2 suggests that the soil is an Entic Haplorthod.

Transect Site 8

Site 8 is located on a large dune immediately west of Highway 123 and approximately 5 km north of Murphy Creek (Figure 5:20). The dune is situated on an upland area and is stabilized by a red pine stand. Dune limbs are orientated to the west and the deposit is about 6 m in relief.

The surface soil on the crest of Site 8 is a well developed Spodosol that contains A-E-Bhs-Bs1-Bs2-BC-C horizonation (Figure 5:13). All horizons are texturally classified as sand, with fine sand being the dominant component. The C horizon begins at a depth of 1.13 m, is yellowish brown (10 YR 5/4) and has a pH of 6.5 (Table 5:3). Overlying the C horizon is a 25-cm thick (1.13-0.88 m) BC horizon, which is brown (7.5 YR 5/4) and has a pH of 5.7. The Bs2 horizon extends from 0.88-0.65 m. This horizon is strong brown (7.5 YR 4/6) and is more acidic than the underlying horizons (pH = 5.0). Overlying the Bs2 is the Bs1, which extends from 0.65-0.42 m. The Bs1 is also strong brown (7.5 YR 4/6) and has a pH value of 5.1. The uppermost B horizon is a 6-cm thick Bhs (0.42-0.36 m), which is dark reddish brown (5 YR 2.5/2) and has a pH of 4.7. Each of the B horizons were well cemented with distinct ortstein concretions, which is reflected by the weighted Fe extractions of 6.44 (Fe_d) and 2.30 (Fe_o) derived from the group. The E horizon is 23-cm thick (0.36-0.13 m), has a grayish brown color (10 YR 5/2), and a pH value of 4.6. Capping the soil is a 13-cm thick, black (5 YR 2.5/1) A horizon. This horizon is the most acidic of all the horizons at the site, with a pH of 4.5. Overall, this

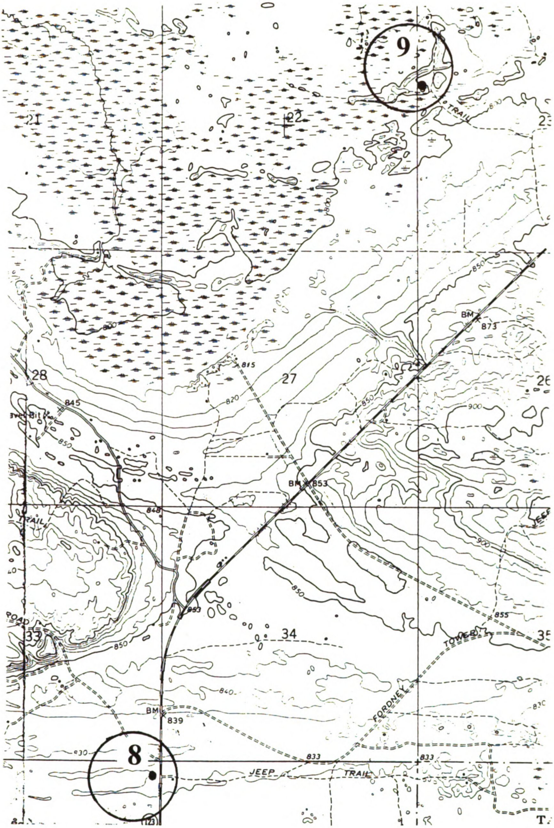


Figure 5:20 - Location of Transect Sites 8 and 9 (Modified from U.S.G.S. Betsy Lake SW, Mich., 1:24,000, 1968).

soil is well developed and has a POD index value of 14, suggesting Typic Haplorthod morphology.

Transect Site 9

Site 9 is located on a large dune 1.5 km north of Highway 123 and approximately 4.5 km north-northwest of Site 8 (Figure 5:20). Surrounded by lowland swamp, the dune is anchored by a mature, mixed stand of red pine and yellow birch. Dune limbs are orientated to the northwest and the deposit is about 9 m in relief.

The surface soil on the crest of Site 9 is a well developed Spodosol that contains A-E-Bhs-Bs1-Bs2-BC-C horizonation (Figure 5:13). All horizons are texturally classified as sand, with fine sand being the dominant component. The C horizon begins at a depth of 1.03 m, is yellowish brown (10 YR 5/6) and has a pH of 6.5 (Table 5:3). Overlying the C horizon is a 20-cm thick (1.03-0.83 m) BC horizon, which is strong brown (7.5 YR 5/6) and has a pH of 6.0. The Bs2 horizon extends from 0.83-0.55 m. This horizon is strong brown (7.5 YR 5/6) and has a pH value of 5.6. Capping the Bs2 is the Bs1, which extends from 0.55-0.32 m. The Bs1 is also strong brown (7.5 YR 5/6) but is more acidic (pH = 5.3) than the underlying Bs2. The uppermost B horizon is a 5-cm thick Bhs (0.32-0.27 m), which is dark reddish brown (5 YR 2.5/2) and has a pH of 5.5. Each of the B horizons is well cemented with distinct ortstein concretions. Weighted Fe extractions of 3.51 (Fe_d) and 1.76 (Fe_o) are derived from the horizons. The E horizon is 23-cm thick (0.27-0.04 m), has a light brownish gray color (10 YR 6/2), and a pH value of 4.8. Capping the soil is a slightly eroded, 4-cm thick A horizon. This horizon has a dark gray color (10 YR 4/1) and also has a pH of 4.8. As a whole, this soil is well developed. The

POD index value for the soil is 18, suggesting a Typic Haplorthod classification.

Transect Site 10

Site 10 is located on a large dune immediately west of County Road 500 and approximately 2.0 km north of the Little Two Hearted River (Figure 5:21). The dune is situated on an upland area and is stabilized by a mixed stand of mature red pine and yellow birch. Dune limbs are orientated to the northwest and the deposit is about 12 m in relief.

The surface soil on the crest of Site 10 is a moderately developed Spodosol that contains A-E-Bhs-Bs1-BC-C horizonation (Figure 5:13). All horizons are texturally classified as sand, with fine sand being the dominant component. The C horizon begins at a depth of 1.17 m, is yellowish brown (10 YR 5/4) and has a pH of 5.4 (Table 5:3). Overlying the C horizon is an 18-cm thick (1.17-0.99 m) BC horizon, which is strong brown in color (7.5 YR 5/6) and has a pH of 5.4. The Bs1 horizon extends from 0.99-0.66 m. The Bs1 is also strong brown (7.5 YR 4/6) and has a pH of 5.0. The uppermost B horizon is a 7-cm thick Bhs (0.66-0.59 m), which is dark reddish brown (5 YR 3/3) and has a pH of 4.4. Moderate cementation of the B horizons is present and weighted Fe extractions of 4.62 (Fe_d) and 2.14 (Fe_o) are derived from these horizons. The E horizon is 31-cm thick (0.59-0.28 m), has a grayish brown color (10 YR 5/2), and a pH value of 4.3. Capping the soil is a 28-cm thick, black (10 YR 2/1) A horizon, which also has a pH of 4.3. Overall, this soil is moderately well developed. In that context, the calculated POD value for the soil is 10, which places the soil as a Typic Haplorthod.

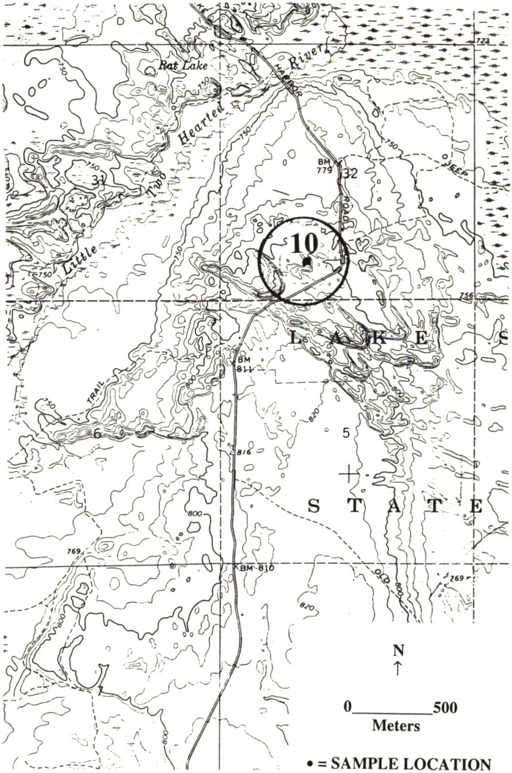


Figure 5:21 - Location of Transect Site 10 (Modified from U.S.G.S. Betsy Lake South, Mich., 1:24,000, 1968).

Relative Age of Transect Sites

Soil development and radiocarbon inconsistencies at the Seney Sand Pit and Tahquamenon Falls sites suggest irregular periods of dune stabilization. As such, relative dating techniques were applied to 10 dune pedons along the transect detailed above to better indicate spatial and temporal patterns of eolian activity. Resultant data further suggest that dune soil development within the study area was not uniform. Although soil-forming factors (excluding time) are held reasonably constant between sites, morphologic and chemical soils data indicate notable variability.

Morphologic Data

Parent material is largely uniform between sites. Fine and medium sands comprise roughly 90% of the sediment at nine of the ten sites (Figure 5:22A). Average silt content per pedon ranges from 1.7-3.8%. No measurable clay is present in any sample. Textural data of C horizons further suggest parent material uniformity, with all textures classifying as "sand" (>97% sand, <2% silt, 0% clay; Table 5:3).

Despite parent material consistency, morphologic indicators of soil development are highly variable. In particular, B horizon thickness and POD indices fluctuate greatly between the sites. B horizon thickness ranges from 36 cm (Site 7) to 64 cm (Site 3) with varying intermittent values (Figure 5:22B). The number of B sub-horizons is also inconsistent. Six sites (1, 3, 4, 6, 8 and 9) possess Bhs-Bs1-Bs2 horizonation, while Bs1-Bs2 (Sites 2 and 5) and Bhs-Bs1 (Site 10) sequences are also present (Figure 5:13). Site 7, the most weakly developed pedon, has only one B sub-horizon (Bs1).

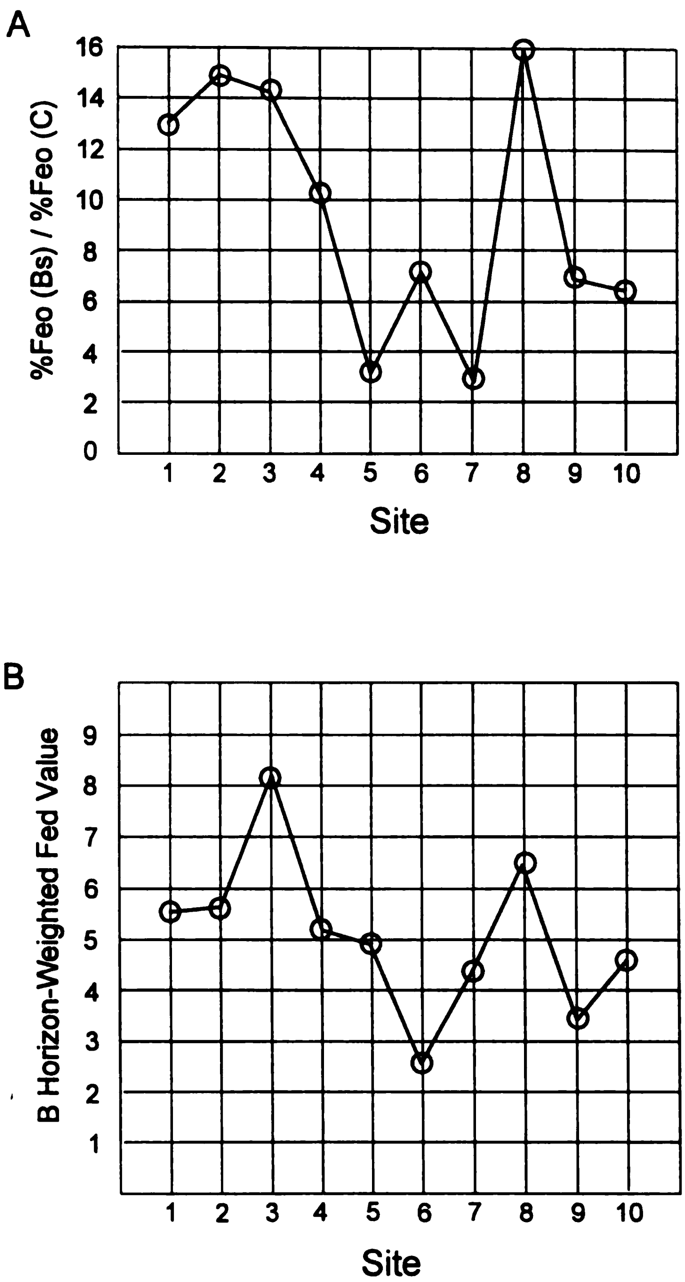


Figure 5:24 - (A) Variation in %Feo(Bs) / %Feo(C) and (B) Horizon-weighted Fed content by site.

Resultant POD index values also reflect the differences in soil development (Figure 5:23). For example, Site 1 has a POD index value of 22, indicating an extremely well developed Spodosol (Schaetzl and Mokma, 1988). In comparison, Sites 5 and 7 express very weak (Entic) development and have POD index values of 2, respectively. Using POD indices as a criterion for soil taxonomic unit (2-6 = Entic Haplorthod; 7+ = Typic Haplorthod), six sites probably classify as Typic Haplorthods, while four could be Entic Haplorthods.

Chemical Data

Differences in B horizon thickness and POD indices suggest a great deal of variability between sites. In an effort to better establish these inconsistencies, chemical measures were applied to soil samples. Twenty measures of Fe and Al were calculated for each site (Table 5:4). While differences exist in C horizon Fe and Al content (% of extract in C), variation within Bs horizons is much more pronounced (% of extract in Bs). Indeed, the degree of variation is even more evident when Bs horizon values are controlled for parent material Fe and Al. For example, $\%Fe_o(Bs)/\%Fe_o(C)$ values range from 3-16 (Figure 5:24A). These results suggest that fluctuating Fe and Al within the Bs horizons are due to differing degrees of soil development, rather than parent material discrepancies. Horizon-weighted extraction data (Wtd. extract) were also calculated for all Bs sub-horizons (Table 5:4). This method has been especially effective in demonstrating the relative age of Podzols in Michigan (e.g. Barrett and Schaetzl, 1992, 1993; Arbogast *et al.*, 1997; Jameson, 1997). For example, plotting B horizon-weighted

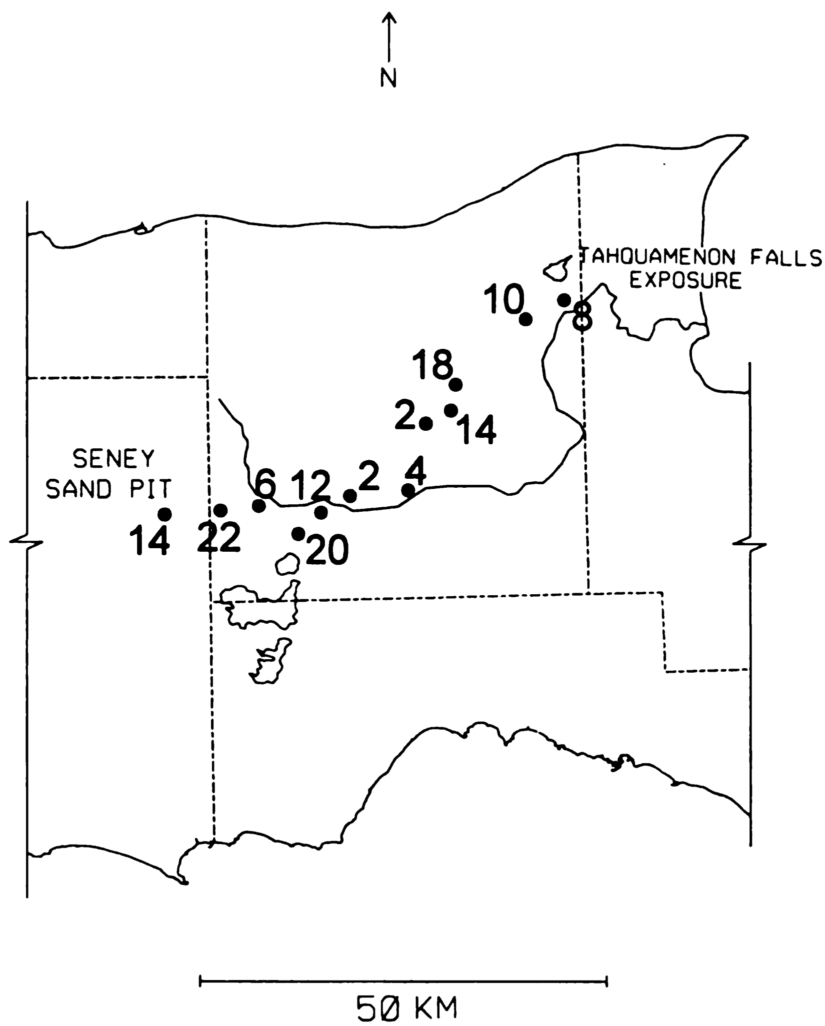


Figure 5:23 - POD index values by site (POD values shown in red; Modified from MIRIS base map, 1998).

Measure	Site 1	Site 2	Site 3	Site 4	Site 5	Site 6	Site 7	Site 8	Site 9	Site 10
%Feo (Bs)	0.175	0.152	0.147	0.126	0.085	0.089	0.051	0.128	0.106	0.168
%Feo (C)	0.013	0.01	0.01	0.012	0.024	0.012	0.017	0.008	0.015	0.026
%Feo(Bs)-%Feo(C)	0.162	0.142	0.137	0.114	0.061	0.077	0.034	0.12	0.091	0.142
%Feo(Bs)/%Feo(C)	13.46	15.2	14.7	10.5	3.54	7.42	3	16	7.07	6.46
Wtd. Feo	2.488	4.28	3.302	2.178	1.941	1.208	1.836	2.298	1.758	2.138
%Fed (Bs)	0.394	0.211	0.352	0.301	0.233	0.197	0.118	0.376	0.221	0.385
%Fed (C)	0.018	0.008	0.017	0.021	0.02	0.016	0.023	0.016	0.028	0.043
%Fed(Bs)-%Fed(C)	0.376	0.203	0.335	0.28	0.213	0.181	0.095	0.36	0.193	0.342
%Fed(Bs)/%Fed(C)	21.89	26.38	20.71	14.33	11.65	12.31	5.13	23.5	7.89	8.98
Wtd. Fed	5.686	5.848	8.192	5.115	4.942	2.72	4.248	6.438	3.51	4.619
%Alo (Bs)	0.495	0.26	0.537	0.375	0.143	0.23	0.073	0.604	0.303	0.449
%Alo (C)	0.038	0.011	0.035	0.037	0.03	0.025	0.02	0.032	0.041	0.033
%Alo(Bs)-%Alo(C)	0.457	0.249	0.502	0.338	0.113	0.205	0.053	0.572	0.262	0.416
%Alo(Bs)/%Alo(C)	13.03	23.64	15.34	10.14	4.77	9.2	3.65	18.88	7.39	13.61
Wtd. Alo	8.14	4.929	11.337	6.865	3.34	2.47	2.808	11.784	4.855	7.953
%Ald (Bs)	0.802	0.727	0.78	0.549	0.392	0.366	0.256	0.902	0.798	0.777
%Ald (C)	0.064	0.028	0.061	0.057	0.061	0.035	0.054	0.051	0.09	0.079
%Ald(Bs)-%Ald(C)	0.738	0.699	0.719	0.492	0.331	0.331	0.202	0.851	0.708	0.698
%Ald(Bs)/%Ald(C)	12.53	25.96	12.79	9.63	6.43	10.46	4.74	17.69	8.86	9.84
Wtd. Ald	13.272	13.299	16.905	9.963	9.03	6.096	9.216	17.312	15.194	14.071

Table 5:4 - Chemical extract data by site.

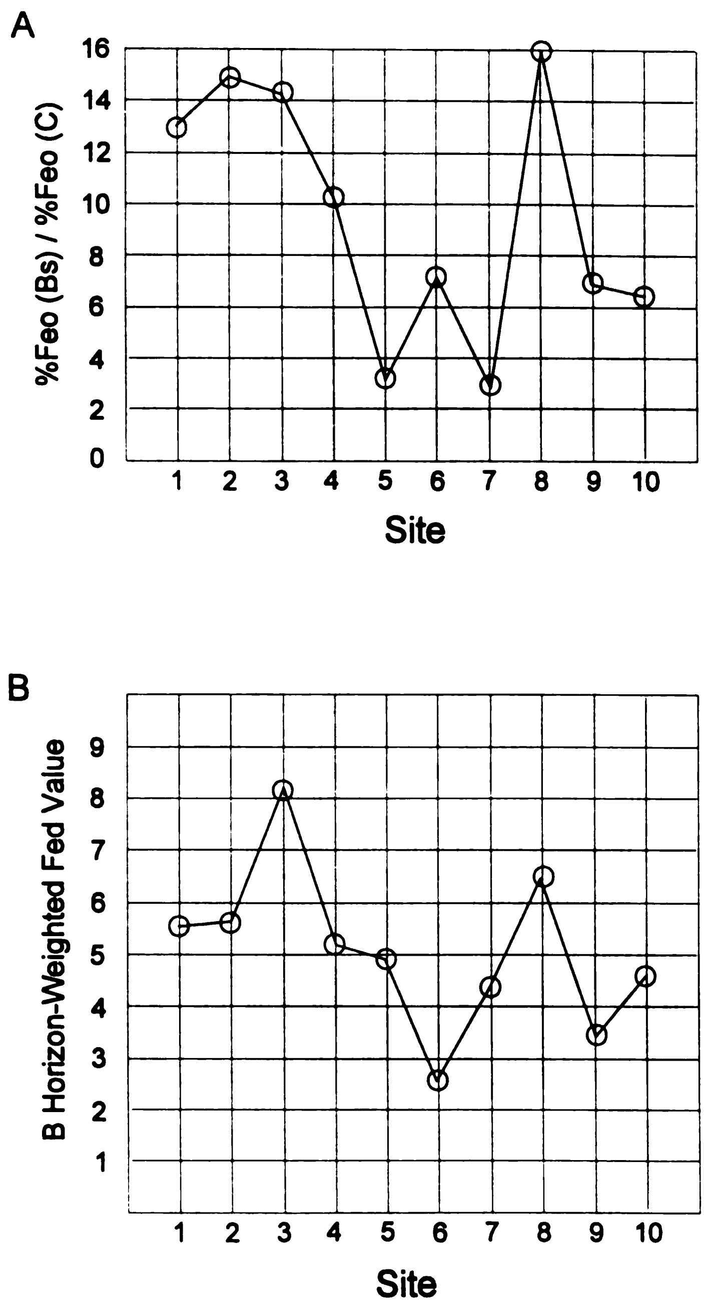


Figure 5:24 - (A) Variation in %Feo(Bs) / %Feo(C) and (B) Horizon-weighted Fed content by site.

Fe_d by site suggests notable variation in Fe_d within each soil (Figure 5:24B). This pattern of variability is further expressed by the remaining chemical extract measures.

Soil pH values and ranges are moderately variable between sites (Table 5:3). The range of pH values within the solum are fairly consistent, however, with extreme values ranging from 4.2-6.1 (Figure 5:25). The vast majority of soil horizons increased in pH with depth and all A horizons are strongly acidic with pH values of 5.0 or lower. Due to the variability in parent material pH (Figure 5:25), I can not assume a correlation between solum pH and relative age.

In an effort to estimate the age of the Seney dune field, chemical data from this study are compared with soils of known age. Specifically, Barrettt and Schaetzl (1992, 1993) reported similar B horizon-weighted Fe_d and Fe_o data for Podzols on lake terraces of known age in northwest lower Michigan (Figure 2:5), and were able to suggest corresponding amounts of soil Fe_d and Fe_o with these ages. The authors also concluded that Fe continues to increase within the B horizon with time. B horizon-weighted values from this study are plotted against the Barrett and Schaetzl data to determine whether any temporal patterns are present (Figure 5:26).

As expected, soil transect data are variable. The majority of points, however, group into one general time bracket within each graph. Ignoring outliers, B horizon-weighted Fe_o data suggests soil formation began roughly 2000-4000 yrs B.P. for most dunes, while most points fall within a 3500-5500 time bracket for the Fe_d data. While soil-forming factors are not identical between both studies, the current study area is probably as, if not more, favorable for podzolization than the Barrett and Schaetzl studies due to thicker

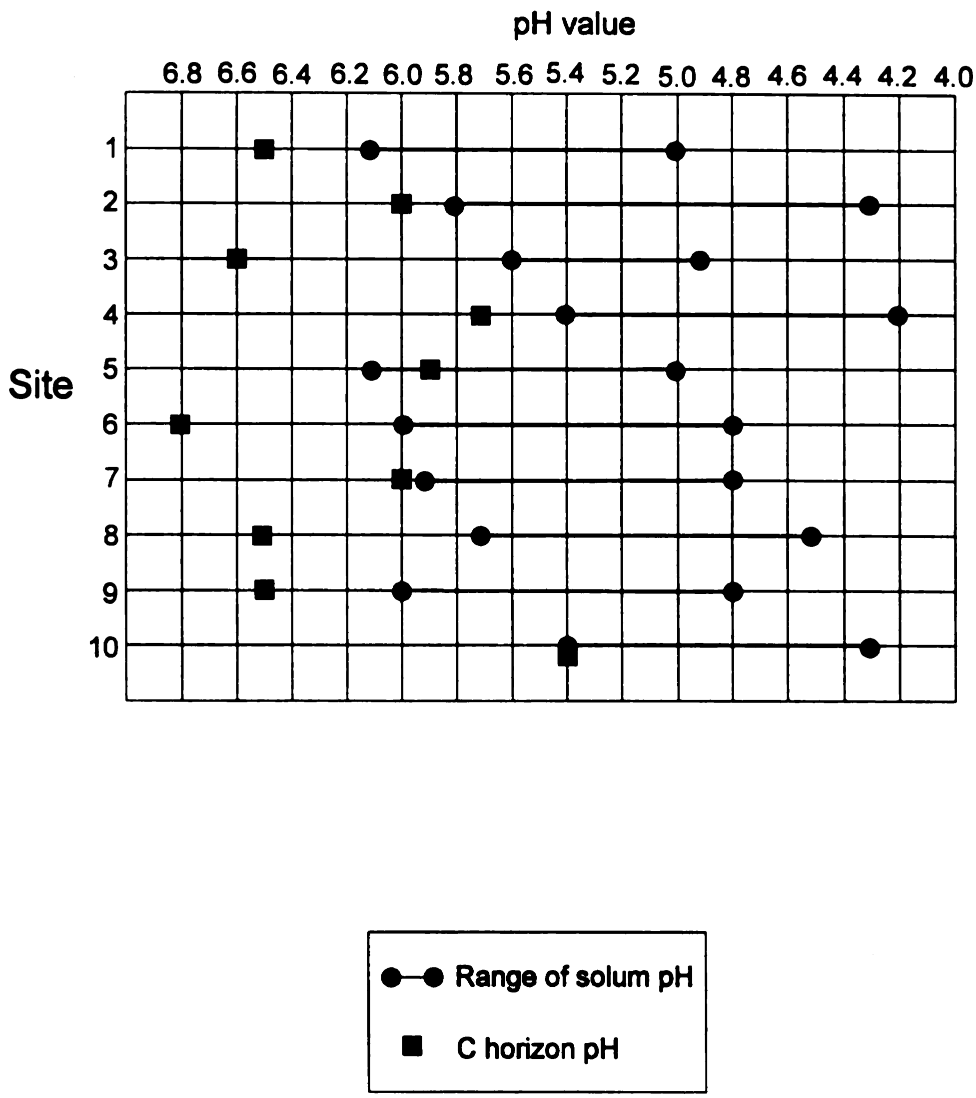
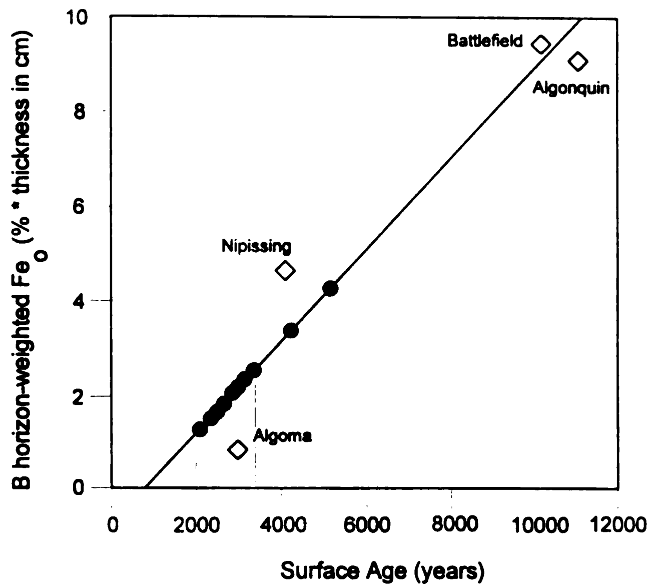


Figure 5:25 - Range of solum pH values by site (C horizon values are also included).



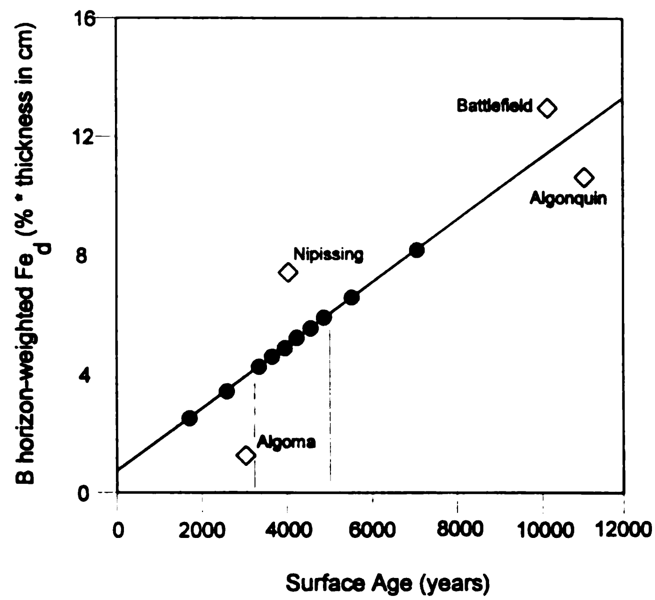


Figure 5:26 - Comparison of data (Barrett and Schaetzl, 1992, 1993) for chemical extractants from dune soils in the study area with similar data for soils of known age (Algoma = 3000 yrs B.P.; Nipissing = 4000 yrs B.P.; Battlefield = 10,000 yrs B.P.; Algonquin = 11,000 yrs B.P.) in northwest lower Michigan.

insulating snowpacks within the region. As such, this comparison suggests that dune stabilization occurred no earlier than 5000 yrs B.P.

Summary and Conclusions

Due to their sensitivity to climatic change, sand dunes provide useful information about paleoenvironmental conditions. Forested dunes are of particular interest because their mobilization presumably requires large-scale environmental change. Research conducted in forested dune fields of northeastern North America indicates that the cause and timing of regional eolian activity was not uniform across the region, with dunes mobilizing during deglacial conditions (e.g., Thorson and Schile, 1995), mid-Holocene climate shifts (e.g., Keen and Shane, 1990) and episodes of increased wildfire (e.g., Filion *et al.*, 1991). As such, no singular mode of dune mobilization within the region can be assumed. Central to the previous studies is Michigan, where forested dunes are common, but poorly understood. This pilot study examined forested dunes in the eastern upper peninsula of Michigan. Although this investigation is limited by the number of dunes (12) and radiocarbon determinations (2) examined, the data derived from this study provide a preliminary record of eolian activity within the region.

The primary goal of this study was to determine the probable age of these dunes. Toward that end, three multiple working hypotheses were tested: 1) dune formation shortly after deglaciation, 2) activation and stabilization of dunes during mid-Holocene climate shifts and 3) episodic remobilization of dunes in the late Holocene. Dune formation in a deglacial setting would have transpired immediately following the drainage of post-glacial Lake Algonquin from the area about 10,000 yrs B.P., when barren and unconsolidated sands could have been reworked by strong winds. Although dunes may have developed shortly after Lake Algonguin receded, results from this study suggest that the present eolian landscape formed well after deglaciation. One piece of evidence that disfavors a deglacial age for present dunes is their orientation. Where other deglacial eolian activity is reported (e.g., Thorson and Schile, 1995; David, 1981), a northeasterly dune orientation reflects anticyclonic winds, even as late as 8000 yrs B.P. Conversely, the dune orientation rose calculated for this study (Figure 3:5) indicates that most eolian mobilization occurred when northwesterly winds prevailed. Thus, the dunes probably formed after anticyclonic circulation broke down within the region.

In addition to dune orientation, other, more quantitative evidence conflicts with a deglacial scenario. A radiocarbon age of about 5000 yrs B.P., derived from organic-rich sediment, was obtained from the base of of a dune at the Seney Sand Pit. This date indicates that mobilization occurred well after the subaerial exposure of lacustrine sands. Likewise, a comparison of chemical soils data with landforms of known age (Barrett and Schaetzl, 1992, 1993) suggests that soil development began no earleir than 5,500 yrs B.P. While these results are not consistent with deglacial dune formation, they do correspond with a mid-Holocene warming scenario. These findings are also consistent with research in Minnesota (Grigal *et al.*, 1976; Keen and Shane, 1990), where northwesterly orientated dunes formed between 8000-5000 yrs B.P. during warmer, drier conditions.

While a mid-Holocene mobilization scenario is supported, results also indicate that localized remobilization of dunes may have occurred. A radiocarbon date of about 700 yrs B.P. was obtained from sub-dune strata at the Tahquamenon Falls site, although this date may be unreliable. In addition, surficial soil development varies dramatically across the study area. For example, notable differences are present regarding B horizon thickness (ranging from 36-64 cm) and POD index values (ranging from 2-22). This variability could be due to differing periods of stabilization. Minor climatic fluctuations and isolated wildfire occurrence, perhaps working in tandem, are suggested as possible causes of late-Holocene remobilization. While evidence of wildfire (i.e., charcoal) was not noted at the study sites, and the degree of climate change needed to destabilize forested dunes is not fully understood, these processes seem to be probable catalysts of dune reactivation.

In summary, results suggest that the current eolian landscape is not immediately deglacial in age. While dunes may have initially formed as Lake Algonquin receded, this scenario is impossible to determine from the results of this study. Thus, dunes could be reworked deglacial deposits or may have been first formed during the mid-Holocene, probably as a result of warmer, drier conditions. A subsequent reactivation of dunes may have occurred episodically as a result of small-scale shifts in climate and/or destabilization due to localized wildfires.

Contributions Of This Study

This study has established a preliminary record of late Quaternary eolian activity in the eastern upper peninsula of Michigan. Within a regional context, results correspond reasonably well with studies indicating mid-Holocene mobilization in Minnesota (Figure 1:1; Grigal *et al.*, 1976 ; Keen and Shane, 1990) and late Holocene destabilization in Quebec (e.g., Filion *et al.*, 1991). Conversely, this study conflicts with deglacial scenarios reported to Connecticut and northern Saskatchewan (Thorson and Schile, 1995; David, 1981). As a result, a gap within the regional geomorphic record has been further clarified.

This study also illustrates the sensitivity of forested eolian landscapes to change. Although forested dunes are presently stabilized in wetland environs, results suggest that they may remobilize much more readily than previously thought. As such, these data could potentially contribute to greenhouse warming studies regarding landscape response. This relationship is especially critical in the context of human-induced deforestation in the region. Further, this study suggests multiple periods of dune activity within the study area, making the probability of multiple episodes of dune mobilization more likely in the future.

Further Research

A more intensive examination of the study area is required before a comprehensive eolian record can be established. This thesis provides a generalized record based largely upon qualitative comparisons. An increase in sample sites would allow for the use of statistical measures to aid in quantifying spatial and temporal trends within the study area. Further, more demonstrative age control is needed to provide absolute dates of eolian activity. Increased radiometric dating of sub-dune deposits would provide more definitive maximum-limiting ages of dunes, while the application of thermoluminescence dating would better establish periods of sediment deposition. Additionally, the relationship between soil development characteristics and absolute age should be clarified to test hypotheses relating soil development and landform age. Lastly, an intensive investigation of soils in a dune is recommended. This examination would provide information regarding the natural variability in soil development and other age determining characteristics within a dune.

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