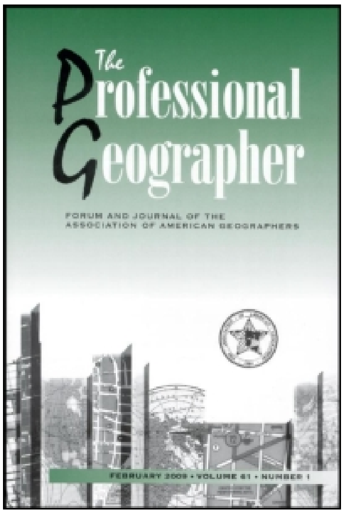

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Estimating the Time Since Final Stabilization of a Perched Dune Field Along Lake Superior

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ARTICLES

Estimating the Time Since Final Stabilization of a Perched Dune Field Along Lake Superior

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The Nodaway dune field is perched along Lake Superior in Upper Michigan. This study uses absolute and relative-age dating methods to test the hypothesis that the dune field finally stabilized after the Nipissing high stand, about 4,000 years ago. Surface soils on snouts of all dunes are moderately developed Spodosols, indicating that dunes stabilized within a few hundred years of each other. One thermoluminescence date provided an age of 8 ka from soil parent material, but is probably overestimated due to residual thermoluminescence. Subsequent optical stimulated luminescence and accelerator mass spectrometry age estimates indicate that the most recent accumulation of sand occurred between ~3.7 and 3.0 ka. This interval suggests one of three possibilities: 1) that the dune field was reactivated during the Algoma high stand and then stabilized; 2) that the dune field stabilized gradually, probably as sand supply diminished after the Nipissing high stand; and 3) that a combination of these two processes occurred. **Key Words: Nodaway Point, perched dunes, Lake Superior, Spodosols, absolute dating.**

Introduction

Coastal dunes are common along the Great Lakes (Olson 1958; Farrand and Bell 1982). Although most coastal dunes are near lake level, some are perched high on bluffs composed of late-Wisconsin glacial drift (Olson 1958). The best known perched systems, Sleeping Bear dunes in northwestern Lower Michigan (Snyder 1985) and Grand Sable dunes in Upper Michigan (Marsh and Marsh 1987; Anderton and Loope 1995; see Fig. 1), are partially active. Research has recently been conducted in both dune fields to determine whether their periodic mobilization is related to lake levels.

According to Snyder (1985), the bulk of the Sleeping Bear dune field formed less than 3,000 radiocarbon years before present (yrs B.P.), with most activity occurring during the past 1,000 yrs B.P. Anderton and Loope (1995) demonstrated that the Grand Sable dunes initially formed during the Nipissing transgression, which peaked about 5,500 yrs B.P. The Grand Sable dune field was episodically activated in the past 5,000 years. Mobilization occurred during high lake stages because sand supply from the upper bluff face increased through undercutting of the bluff. The dune field apparently stabilized during low lake stages due to diminished sand supply because the lakeside bluff became isolated from coastal erosion.

Another perched dune field is the Nodaway dune field, located on a high (~60 m) bluff overlooking Nodaway Point along the southern shore of Lake Superior about 150 km east of the Grand Sable Dunes (Figs. 1 and 2A). This dune field is not active and is entirely forested. Since no detailed study of the Nodaway dunes has been conducted, the length of time since the field last stabilized is unknown. There has been previous research in the area (Saarnisto 1974a, b; Futyma 1981; Farrand and Drexler 1985), but it focused on the glacial lake shorelines that occur near or within the bluffs (Figs. 2A and 2B).

In the context of the dune field's age, the northern bluff is particularly significant because it faces onshore winds and is the most likely eolian-sand source. According to Farrand and Drexler (1985), this bluff was probably cut by waves during high lake stands that occurred after the Algonquin and post-main Algonquin stages (~11,000 to 10,200 yrs B.P.), first during ancestral Lake Minong (~10,000

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Figure 1: *The location of Nodaway Point and related sites in the central Great Lakes region. The Grand Sable and Sleeping Bear dunes are also perched dune fields, and Wilderness State Park and sites nearby provide relative age control by Barrett and Schaetzl (1992, 1993).*

yrs B.P.) and last during glacial Lake Nipissing (~6,000–4,000 yrs B.P.; Larsen 1987). If Anderton and Loope's (1995) model for the Grand Sable dunes (i.e., perched dunes activate during high lake level and stabilize when lake level drops) can be applied to other perched dune fields, and the chronology of lake level and bluff erosion at Nodaway Point is accurate, then the Nodaway dune field probably stabilized shortly after the Nipissing high stand. This study tests this age hypothesis.

Study Area

The Nodaway dune field is perched near a steep bluff approximately one km inland from Lake Superior within the Hiawatha National Forest. Early investigators (Saarnisto 1974a; Futyma 1981) thought the nearby upland was part of the Newberry moraine, which theoretically formed around 11,000 yrs B.P. (Saarnisto 1974a). Drexler et al. (1983) questioned this interpretation, arguing that the upland consists of bedrock mantled thinly by glacial drift. Between the lake and the bluff lies a lower (60 m below the bluff crest) surface, probably a wave-cut terrace, that is 3 to 4 m above the lake (Fig. 2a).

Previous studies indicate that the oldest paleo-shorelines are those of the post-main Algonquin group, formed between 11,000 and 10,200 yrs B.P., which lie topographically above the modern bluffs (Saarnisto 1974b; Farrand and Drexler 1985). The Payette shoreline was probably truncated when the eastern bluff was cut, possibly from catastrophic discharge of Lake Agassiz (Farrand and Drexler 1985). According to Farrand and Drexler (1985) the bluffs were the active shoreline during the later Minong and Nipissing lake stages, ~10,000 and ~5000 yrs B.P., respectively.

The dunes apparently buried a portion of the highest (Wyebridge) shoreline, but did not cover the later (e.g., Payette) post-main Algonquin, Minong, and Nipissing shorelines (Fig.

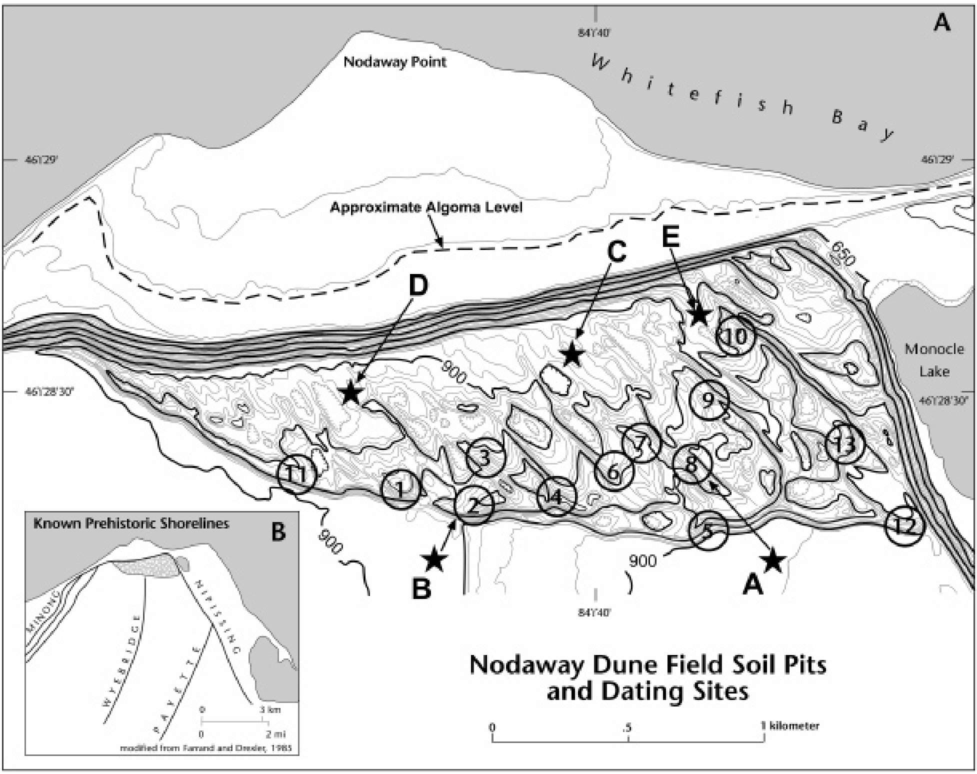

Figure 2: *A) Topographic map of the Nodaway dune field, relative to Lake Superior (Whitefish Bay), showing the position of soil pits (1–13) and sites where "absolute" ages were obtained (A–E). The contour interval that is represented by solid (dark) lines is 50 feet (~15 m), with the base of the bluff at 650 feet (~200 m) above sea level. The highest elevation within the dune field occurs near site 4 and is about 1,045 feet (~318 m) above sea level (modified from the USGS Dollar Settlement Quadrangle, 1951); B) Inset map showing the position of the dune field relative to prehistoric shorelines in the immediate area (modified from Farrand and Drexler 1985).*

2B). Although the dune field is small (<5 km^2), most individual dunes are immense (>50 m tall). Dunes do not occur immediately adjacent to the bluff, but begin about 100 m inland from the plateau edge. The vast majority of the dunes are very well defined, consisting of parabolic features (oriented ~315°; Fig. 2A) that extend southeast across the entire field as they climb to the crest. In addition to these largest dunes, some smaller hummocks (<5 m high) of eolian sand occur in the northern part of the dune field.

The climate of the area is lake-modified continental. The nearest National Weather Service station is Sault Ste. Marie, about 20 km to the east (Fig. 1). Here, the average January temperature is about −10°C and the mean July temperature is approximately 18°C. Mean annual precipitation is around 80 cm, with about half accumulating as snowfall (NOAA 1995). The dune field is entirely stabilized by a mature forest of mixed species, including sugar and red maple (*Acer saccharum* and *Acer rubrum*), American beech (*Fagus grandifolia*), and northern red oak (*Quercus rubra*; Voss 1985).

Methods

Field Methods

The original study goal was to reconstruct the entire geomorphic history of the dune field, consistent with the investigations conducted by Anderton and Loope (1995) in a largely nonforested locality where outcrops of buried soils are common in various stratigraphic settings (e.g., slopes, blowouts). After an initial reconnaissance, however, it was determined that the massive size of the Nodaway dunes, coupled with the dense forest and complete lack of natural exposures, precluded a full reconstruction unless highly sophisticated equipment (e.g., ground penetrating radar, large drill rig) was used at great effort (e.g., cutting wide paths through the forest) and expense. Resources were not available for such an undertaking; thus, the project was scaled back to estimate the time since final stabilization of the dune field and to correlate that time with known lake level fluctuations in the area (e.g., Saarnisto 1974b; Farrand and Drexler 1985).

In order to test the study hypothesis, a field methodology was designed to locate and investigate sites where eolian sand most recently accumulated in the dune field. These sites were selected on the basis of known dune processes (e.g., McKee 1979; David et al. 1998) and were probed with a shovel and bucket auger. The field investigation thus centered on dune crests, mostly on the best defined dunes that dominate the dune field, because they are the primary depositional zone when sand is deflated from blowouts (i.e., depressions between limbs of parabolic dunes; McKee 1979; David et al. 1998). Although truncated older deposits of sand hypothetically exist in the paleo-blowouts of the dune field, these sites were not probed because of the previously noted study limitations and adjusted research goals. In addition to the focus on dune crests, the contact between the southern margin of the dune field and the underlying glacial surface was investigated because it theoretically records the most recent downwind migration of the dune field. Lastly, the hummocks in the northern part of the dune field were probed because they may represent the youngest (albeit relatively minor) pulse of eolian deposition in the system.

The first goal of the fieldwork was to empirically determine whether all of the largest dunes stabilized concurrently by analyzing surface soils on their crests. Given the apparent consistency in soil texture, biota, and climate, it was assumed that time is the discriminating soil-forming variable (e.g., Franzmeier and Whiteside 1963; Barrett and Schaetzl 1992). Field reconnaissance indicated that the soils trended toward Spodosols, which are particularly good for relative dating (e.g., Arbogast et al. 1997; Arbogast and Jameson 1998) within dune fields because they rapidly form distinctive horizons (albic, spodic) that vary markedly with time (Franzmeier and Whiteside 1963; Barrett and Schaetzl 1992). Spodosols form by podzolization, a suite of pedogenic processes that translocate organic and inorganic (Fe and Al) compounds (Rourke et al. 1988). Podzolization is notably vigorous in northern Michigan because the litter layer is acidic, precipitation is abundant, and water rapidly infiltrates the sandy sediments (Schaetzl and Isard 1991). The soils in the Nodaway dune field were accessed in 13 soil pits, (Fig. 2a) evenly scattered over the vast majority of individual dunes, which were excavated to about 2.0 m depth. Soils were described using standard methods (Soil Survey Division Staff 1993), and horizon-based samples of approximately 400 grams were collected for laboratory analyses.

In an effort to quantify Spodosol development based on morphologic data (i.e., horizonation, color), podzolization (POD) Indices (Schaetzl and Mokma 1988) were calculated for each site. The POD Index assumes that podzolization produces E and B horizons with colors that increasingly contrast with development. Thus, E and B horizon colors, as well as the number of B subhorizons, are required to calculate the Index. Assuming that time is the dependent variable, Schaetzl and Mokma (1988) theorized that POD values should increase with age, and demonstrated that Entic Haplorthods (A/E/Bs/BC/C horizonation) have a POD Index of about 2, whereas Typic Haplorthods (A/E/Bhs/BC/C horizonation) have a POD Index ≥ 6.

Once the soils were described, samples were collected for "absolute" age determination (Table 1). Two samples were collected for radiocarbon dating, one from the dune/drift contact in the toe of the dune field and the other from the same contact in the hummocky northern part of the field. Additional ages were obtained through thermoluminescence (TL) and optical stimulated luminescence (OSL) dating of individual sand (quartz) grains. These methods measure the amount of radiation emitted by electron traps upon the application of heat or laser luminescence, respectively. This value, when correlated with a known dose rate for the

Table 1 *Radiocarbon and Luminescence Ages Obtained in This Study*

Site	Method	Stratigraphic Position	Lab #	Age[1]
Soil pit #8	TL	C horizon (~1.75 m)	W2342	8.4 ± 0.8ka
Soil pit #8	OSL	C horizon (~1.75 m)	24/MHD1	3.38 ± 0.16ka
Age #2	Radiocarbon (AMS)	Drift/eolian sand contact	NSRL-10498	3480–3160
Age #3	Radiocarbon (AMS)	Drift/eolian sand contact	NSRL-10497	7030–6730
Age #4	OSL	C horizon (~1.75 m)	24MHD2	3.72 ± 0.14ka
Age #5	OSL	C horizon (~1.75 m)	24MHD3	3.01 ± 011ka

[1] *Radiocarbon ages are calibrated from conventional δ^{13}C-corrected radiocarbon age to calendar years using a tree-ring curve. All calibrations are reported here at 2σ and were based upon the 20-year atmospheric curve (e.g., Linick et al. 1985; Stuiver et al. 1986). The calibration program used is discussed in Stuiver and Reimer (1993).*

surrounding sand, provides an estimate for the last time the grains were exposed to sunlight (i.e., during eolian transport).

Laboratory Methods

A variety of laboratory procedures were employed to characterize the soils and provide the geochronological estimates. Both radiocarbon samples were assayed by Accelerator Mass Spectrometry at the Institute for Arctic and Alpine Research (INSTAAR) in Boulder, CO and were calibrated to calendar years (Stuiver and Reimer 1993). The TL sample was analyzed at the University of Wollongong (Australia), where it was examined by the combined additive and regenerative methods using the 90–125 μm fraction of quartz grains (Shepherd and Price 1990). Each OSL sample (~150–210 μm fraction) was treated by the single aliquot procedure (Wintle and Murray 1999) at the Luminescence Laboratory in Aberystwyth, Wales.

Several methods were applied to the soil samples. Moist Munsell colors were calculated by two people working independently under a fluorescent light. Although some of the subtle variability in horizon color may be lost, this method is a standard procedure for determining and comparing basic horizon colors (c.f., Arbogast et al. 1997; Schaetzl 1998; Arbogast and Jameson 1998). The samples were then oven-dried at 40° C, sieved, and repeatedly halved to a final weight of approximately 30 g for all subsequent analyses. Fe and Al were extracted from all B horizon samples using sodium citrate-dithionite and acid ammonium oxalate (Soil Survey Laboratory Staff 1996). Fe and Al contents of the extracts (Fe_d, Fe_o, and Al_o, respectively) were measured on a DCP spectrometer. Sodium citrate-dithionite extracts Fe and Al in "free form"—that is, not bound within phyllosilicate lattices (Mehra and Jackson 1960). As a result, it reflects the Fe and Al released by weathering, but not translocated out of the profile. Since oxalate does not extract crystalline oxides efficiently, oxalate-extractable Fe and Al is assumed to be organically-bound and/or inorganic and amorphous (McKeague and Day 1966). In general, Fe_o and Al_o are thought to measure Fe and Al in amorphous forms (i.e., either forms bound to organic chelates or forms such as allophane).

Sand contents were determined by dispersing samples in a $Na_2CO_3[NaPO_3]_6$ solution, shaking overnight, and wet sieving through a 53 μm sieve. Subsequently, the relative percentages of silt and clay were determined by hydrometer analysis. Sands were dry sieved to segregate the very coarse (vcs), coarse (cs), medium (ms), fine (fs), and very fine sand (vfs) fractions (Carter 1993). Reaction was measured in 2:1 water-soil mixtures using an Orion 720A combination pH/ISE meter.

In an effort to detect patterns in data that may reflect relative soil development at the study sites, soils data were statistically analyzed using two techniques. Principal components analysis (PCA) established the sources of data variability. Numerous raw data (Table 2) were input into this analysis, including the weighted (content × horizon thickness) Fe_d, Fe_o, Al_o in the B horizons, because they are good measures of relative age (Barrett and Schaetzl 1992). The PCA was rotated using a varimax rotation (Johnson 1978) and components were limited to those with eigenvalues ≥1.00 and which explained at least 10% of the variance.

Once the variation was identified, cluster analysis objectively grouped test sites by generating classifications based on one or more variables (Everitt 1980). Soils data were first grouped by Ward's minimum variance algorithm (Ward 1963), a hierarchical procedure that groups

Table 2 *Statistical (Mean) Data for Soils at the Nodaway Dune Field and Results of Principal Components Analysis.*

Variables	Mean (σ)	Factor 1	Factor 2	Factor 3	Factor 4
Eigenvalue		5.78	2.48	2.47	2.18
Rotated loadings of variables					
POD index	4.5 (.6)	0.15	−0.68	0.46	−0.14
B thickness (cm)	61.5 (11.7)	0.71	0.58	−0.15	0.00
Solum thickness (cm)	79.1 (14.2)	0.63	0.60	0.00	0.37
A horizon pH	4.5 (.4)	0.18	0.00	0.27	−0.67
Bs horizon pH	5.3 (.3)	−0.14	0.91	0.00	0.00
BC horizon pH	5.8 (.4)	0.00	0.71	0.49	0.00
C horizon pH	6.0 (.3)	0.00	0.64	0.17	−0.24
C horizon VCS (%)	.1 (3.3)	0.00	0.00	−0.12	0.91
C horizon CS (%)	5.6 (3.4)	0.70	−0.27	−0.19	−0.27
C horizon MS (%)	39.3 (13.0)	0.95	0.00	0.00	−0.14
C horizon FS (%)	49.2 (12.8)	−0.93	0.00	0.00	0.00
C horizon VFS (%)	5.1 (3.3)	−0.87	0.16	0.00	0.30
C horizon sand (%)	99.3 (.4)	0.21	0.00	0.92	0.20
C horizon silt (%)	.7 (.4)	−0.21	0.00	0.92	−0.20
C horizon clay (%)	0 (0)	−0.21	0.00	−0.92	0.20
C horizon Fe_d (%)	.04 (0)				
C horizon Fe_o (%)	.02 (0)				
Weighted B horizon total Fe_d (% * thickness in cm)	6.13 (1.06)	0.62	0.00	−0.39	−0.40
Weighted B horizon total Fe_o (% * thickness in cm)	5.14 (1.12)	0.96	0.00	−0.19	0.00
Weighted B horizon total Al_o (% * thickness in cm)	8.3 (2.7)	0.73	0.00	0.00	0.11
Total variance explained (%)		34	15	15	13

within-group sum of squares for each soil pit or cluster. Euclidean distance subsequently measured the association between the soil pits. Although no satisfactory method exists for choosing the optimum number of clusters, the clusters in this study were based on inspection of the linkage tree and the rate of change of the multiple squared correlation coefficient (R^2) with fusion level (e.g., Winkler 1992).

Results

Soils: Morphology

Soils on crests of the best developed dunes were morphologically similar throughout the dune field, consisting of moderately developed Spodosols with A-E-Bhs (weak)-Bs-BC-C horizonation (Fig. 3). The most common colors for each horizon were: A (10YR 2/1 and 2/2), E (7.5YR 4/2), Bhs (5YR 3/2 and 3/3), Bs (5YR 3/3 and 7.5 YR4/4), BC (10YR 4/4 and 4/6), and C (10YR 5/4). Accordingly, POD Indices ranged from 4 to 5; one site (# 7; Figs. 2, 3) had a POD value of 6 (Table 2). B horizon thickness varied from 41 to 80 cm (x = 61.5 cm). Overall solum thickness ranged from 53 to 103 cm (x = 79.1 cm). All pedons had sand or fine sand textures (Soil Survey Division Staff 1993), consisting of approximately 99% sand, about 1% silt, and no measurable clay (Table 2).

Soils: Statistical Analyses

Principal components analysis revealed four primary soil factors, which explain 77% of the total variance (Table 2). Factors 1 and 2 are clearly the most significant (together 49% of total variance) and are specifically related to soil development. Factor 1 represents 34% of the variance and appears to reflect the podzolization process, with B horizon thickness, solum thickness, weighted Fe_d, Fe_o, and Al_o, coarse sand, and medium sand (i.e., coarse texture of parent material reflecting water infiltration) having strong (>.5) positive loadings, and fine sand and very fine sand (i.e., fine texture) loading strongly (<−.5) negative. In factor 2 (15% of the variance), the pH of the BC, Bs, and C horizons loaded strongly positive, as did solum and B horizon thickness. The POD variable loaded strongly negative (<−.5). The remaining factors (3 and 4) accounted for 15% and 13%, respectively, of the total variance. In factor 3 the only variable that loaded strongly positive was percent sand, whereas percent silt

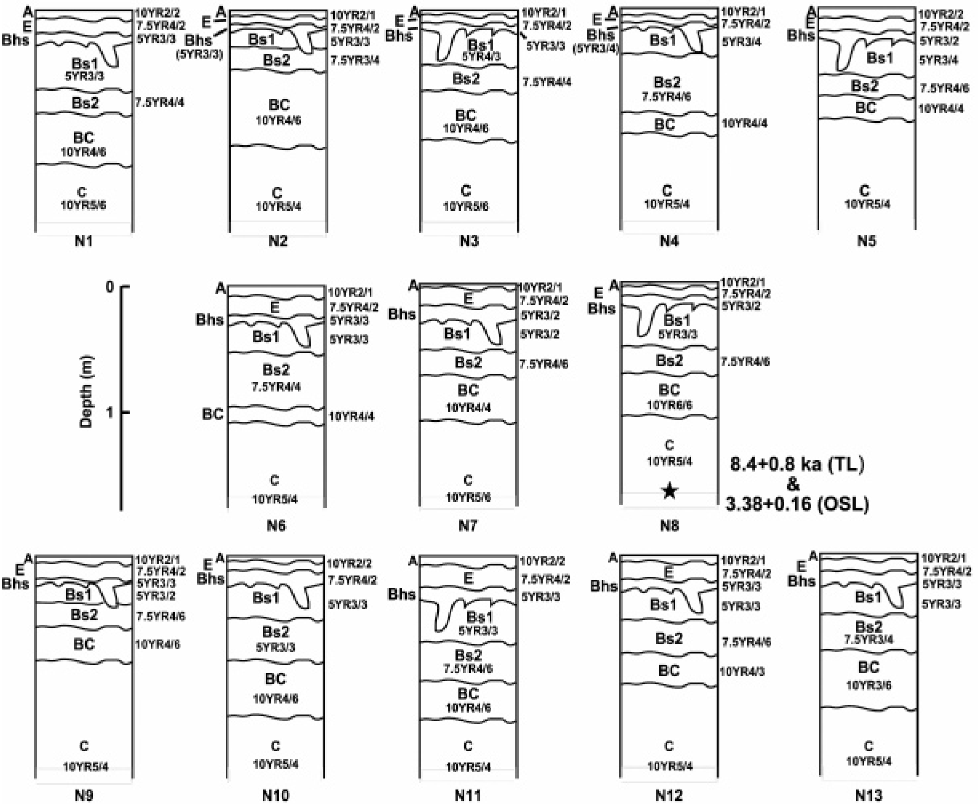

Figure 3: *Soil morphology (drawn from field notes) from the sites N1–N13. Wavy line of horizon boundaries is intended to show the irregular and transitional nature of horizon margins. Note the TL and OSL ages (star) at Site 8.*

loaded strongly negative. This factor may be somehow related to pedogenesis (i.e., water infiltration), but is probably insignificant because silt comprises only ~1% of soil texture. Factor 4 is also an insignificant factor, with very coarse sand loading highly positive and A horizon pH strongly negative.

Inspection of the within-group sum of squares and the linkage tree diagram (not shown) suggested that the soil pits should be grouped into three clusters. This clustering is discriminated by solum thickness, which admittedly may be biased because the base of soils is difficult to precisely determine in the field. Cluster 1 includes soil pits 1, 4, 6, and 11, containing the pedons with the thickest (≥89 cm) sola. The second cluster is the largest, containing the seven (soil pits 3, 5, 7, 8, 10, 12, 13) sites with sola closest to the mean solum thickness, ranging from 69 (soil pit 7) to 84 cm (soil pit 10) thick. Cluster 3 is the smallest and includes the sites (soil pits 2, 9) where sola are relatively thin (53 cm at pit 2; 60 cm at soil pit 9).

Similarity of Soils

Although soils across the Nodaway dune field are morphologically similar, statistical analyses suggest that some subtle developmental differences exist. Although this diversity could be a function of time (i.e., the thickest soils formed in dunes which have been stable the longest), this is not believed to be the case. Spodosols are highly sensitive to microgeographic influences, including leaf-litter distribution (Alexander 1986), microclimate (Macyk et al. 1978; Hunkler and Schaetzl 1997), and the proximity to trees (Crampton 1982). Schaetzl et al. (1990) demonstrated that microtopographic changes caused by treethrow results in significant spatial variability in Spodosols. Moreover, Phillips et al. (1996) argued that Spodosol variability can result from minuscule variations in initial conditions (i.e., deterministic uncertainty). Since the overall differences in soil development are very small, the majority of pits cluster together, and variability can be logically explained, the conclusion is that soils on crests are essentially the same age on all of the large dunes in the dune field. This finding requires that the largest dunes, which dominate the dune field, stabilized more or less concurrently (i.e, within a few hundred years).

Estimated Final Stabilization of the Dune Field

Once the character of soils was determined, the study focus shifted to estimating the time since final stabilization of the largest dunes. The initial age sample was collected from the upper part (1.75 m deep) of the C horizon at soil pit #8 (also age site A; Fig. 2), which has the most average pedon in the dune field. This strategy was logically chosen on the premise that an accurate date could be extrapolated to the other large dunes because of the general consistency in soils. This first sample was analyzed by TL and dated to 8.4 ± 0.8 ka (W2342; Fig. 3). According to Price (1997), this age is highly suspect because the temperature comparison had a double plateau (275–325°C and 375–500°C), which likely occurred because the sand was insufficiently exposed (i.e., heated) during its most recent activation to completely minimize the older TL signal. In addition, the age is likely overestimated because a modern sample was not available for surface correction and the TL starting point was artificially calculated. Given these qualifying variables, the TL age should be viewed simply as a maximum-limiting date for final stabilization of the dune field.

Following the questionable results of the TL sample, additional funds became available for further dating. Five estimates were obtained with these resources, with two acquired by AMS and three obtained by OSL dating. OSL dating is a relatively new method that provides more accurate age estimates than TL because the electron trap is rapidly bleached by sunlight (Wintle 1993). According to Wintle (1993), a quartz sand grain had its OSL essentially (99%) removed upon exposure to sunlight for only ten seconds, whereas no change was observed in the TL over the same interval of time.

Given that five ages could be acquired, an effort was made to sample representative sites that were widely spaced. In that vein, one of the OSL ages was used to redate the uppermost part of the parent material at soil pit #8 (also age site A). This sample provided an estimate of 3.38 ± 0.16 ka (24/MHD1), substantially younger than the TL estimate. The second sample was an AMS estimate of charcoal found at the eolian sand/drift contact at the southern toe of the dune field. This sample was obtained downslope (age site B; Fig. 2) from soil pit #2 and provided an age of 3480–3160 cal. yrs B.P. (NSRL-10498). The consistency of these two ages is encouraging, suggesting that sand accumulation ceased on the crest of the largest dunes at about the same time (~3.2 ka) that the dunes stopped migrating downwind.

In addition to the dates from the largest dunes, three estimates were obtained from the hummocky northern portion of the dune field. One of these dates was an AMS estimate from the eolian sand/drift contact in the northcentral part of the dune field (age site C; Fig. 2) that provided an age of 7030–6730 cal yr B.P. (NSRL-10497). The other pair of estimates were derived by OSL from the upper part (~1.75 m deep) of the C horizons at two sites, one towards the western end of the dune field (age site D) and the other near the eastern margin (age site E; Fig. 2). At each of these two sites the surface soils were qualitatively similar to those formed in the crests of the largest dunes. The OSL estimates were generally consistent as well, with ages of 3.72 ± 0.14 ka (24MHD2) and 3.01 ± 0.11 obtained from western and eastern sites, respectively. These results suggest that eolian sand stopped accumulating in the hummocky northern part at a time more or less concurrent with the largest dunes.

Correlation with Lake Chronology

The deglacial and Holocene history of Lake Superior is complex, depending on the position of the ice front, the lowest available outlet, the effects of isostastic rebound, and climatic fluctuations. The Algonquin and post-main Algonquin stages were the highest levels, occurring between approximately 11,000 and 10,200 yrs B.P. when the ice front was across the Superior basin. As deglaciation progressed and the isostatically depressed North Bay outlet opened, lake level dropped significantly, resulting in the Houghton low phase around 8,000 yrs B.P. As the North Bay outlet rebounded, lake elevation rose dramatically during the Nipissing transgression and peaked approximately 5,500 yrs B.P. at the Nipissing I stage. The lake likely remained relatively high, including the Nipissing II stage, until about 4,000 yrs B.P. (Larsen 1987). Subsequently the lake began to drop as the Port Huron outlet downcut (Farrand and Drexler 1985). During this overall decrease in lake level, a distinct high stand, the Algoma, occurred approximately 3,200 yrs B.P. (Larsen 1987). In all probability, the lake has fluctuated slightly in the late Holocene due to climatic variability (Larsen 1994; Anderton and Loope 1995).

Data from this study shed light on the history of the Nodaway dune field and its relationship to lake level (Fig. 4). A maximum-limiting date of about 7,000 cal yrs B.P. was obtained at the eolian sand/drift contact near the bluff edge. This age was derived from charcoal, suggesting that the plateau was forested (and perhaps burned) during the early Holocene (c.f., Anderton and Loope 1995). Deposition of eolian sand probably began as the rising waters of post-glacial lake Nipissing reached the base of the bluff (Farrand and Drexler 1985) and the upper bluff was destabilized by wave undercutting. According to the chronology at the Grand Sable dune field (Anderton and Loope 1995), approximately 150 km to the

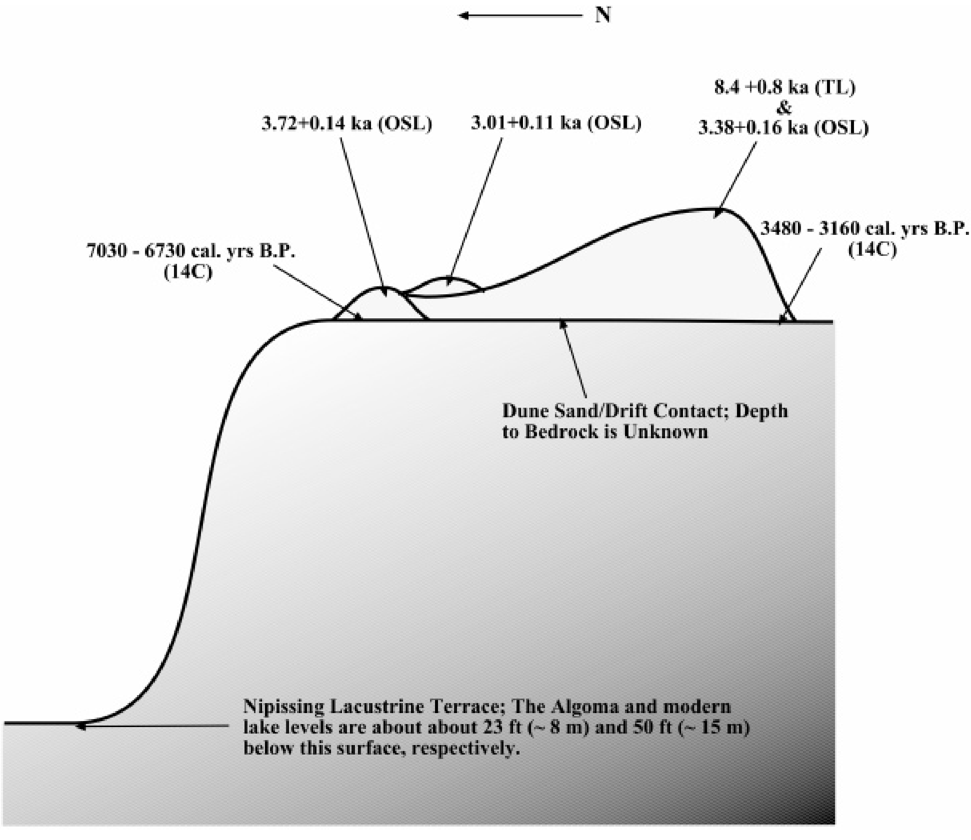

Figure 4: *Schematic cross-section (not to scale) of dune field with a view to the east, including the Nipissing lacustrine surface, bluff, perched dunes, and "absolute" ages.*

west, intensive dune building may have begun at Nodaway Point around 5,500 yrs B.P. The basal date from age site C permits this possibility because it is a maximum-limiting estimate.

Although the overall geomorphic history of the dunes could not be reconstructed given the limitations of this study, the interval when the dune field finally stabilized has been determined. The initial TL date (~8 ka) from the sand parent material is unfortunately nothing more than a maximum-limiting age because of residual TL (e.g., Wintle 1993) and the lack of an ideal modern correction (Price, 1997). Subsequent OSL and radiocarbon data correlate relatively well, however, suggesting that sand last accumulated in the dune field between ~3.7–3.0 ka.

The interval of most recent sand accumulation corresponds best to the transition that occurred between 4,000 and 3,000 yrs B.P. from the Nipissing high stand (including Nipissing I and II) to lower lake levels (Larsen 1987). There are two possible scenarios that explain this chronology. One is that the dune field stabilized shortly after the Nipissing stage and sand was supplied again during the Algoma high phase around 3,200 yrs B.P. (Larsen 1987). The reconstructed history certainly permits this possibility and makes intuitive sense. For this scenario to fit in the context of the perched dune model (c.f., Anderton and Loope 1995), however, the Algoma level must have reached the base of the bluff (so the bluff could be undercut), which is now at an altitude of about 650 ft (~198 m; Fig. 2) due to isostatic rebound. The closest locality where the elevation of the Algoma terrace is known is at Algoma Mills, Ontario (Fig. 1). This site is virtually the same distance (570 miles/917 km) as Nodaway Point from Larsen's (1987) regional isostatic datum (the southern end of Lake Michigan), and is within sufficient geographical proximity (~120 km) to serve as a proxy Algoma level on the lake plain topographically below the dune field. At Algoma Mills the raised elevation of the Algoma terrace is only about 623 ft (~190 m; Larsen 1987), almost 30 ft (~8 m) below the base of the bluff at Nodaway Point (Fig. 2). This dichotomy suggests that during the Algoma level undercutting of the Nodaway bluff occurred only rarely, probably during the strongest storms (allowing extensive wave

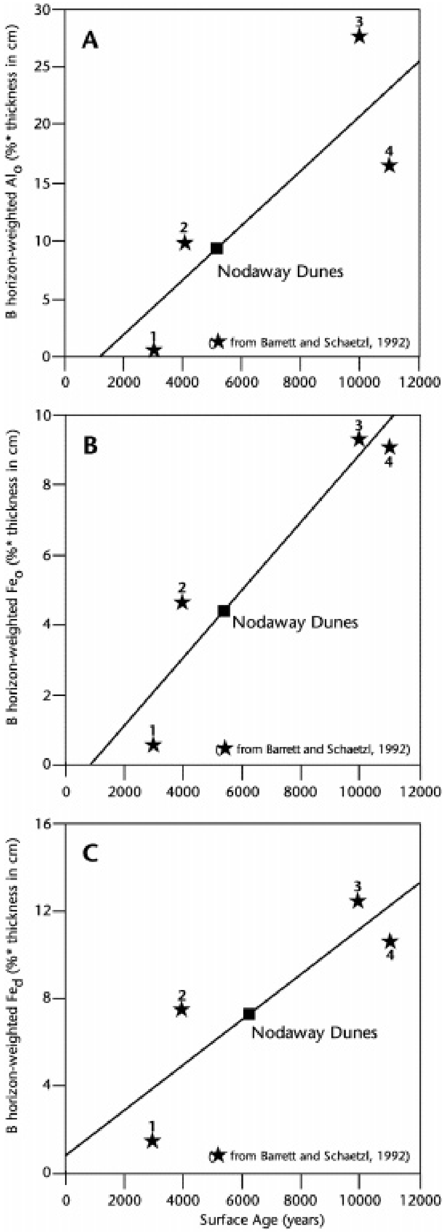

Figure 5: *Comparison of data for chemical extractants from soils in the Nodaway dune field with similar data for soils on surfaces of known age (1 = Algoma; 2 = Nipissing; 3 = Battlefield; 4 = Algonquin) in northwest lower Michigan (modified from Barrett and Schaetzl 1992); (A) B horizon weighted Al_o; B) B horizon weighted Fe_o; C) B horizon weighted Fe_d.*

runup). Such periodic undercutting may have been enough to supply limited volumes of sand to the dune field, and perhaps those dated in this study. Conversely, it is difficult to imagine that the Nodaway bluff was a significant source of eolian sand during the Algoma period because the lake was apparently not at the base of the bluff for extended periods of time.

Given the problems associated with bluff erosion during the Algoma level, the only other logical scenario is that the dune field gradually stabilized following termination of the Nipissing II stage around 4,000 yrs B.P. The elevation of the Nipissing II terrace at Sault Ste. Marie (20 km east) is 646 ft (~197 m), just below the 650-ft (~198 m) elevation of the bluff base at Nodaway Point (Larsen 1987). Thus some significant bluff destabilization probably occurred at Nodaway Point as late as 4,000 yrs B.P. As lake level subsequently fell from the Nipissing II stage, however, the dune field hypothetically changed from being fully or mostly active to being increasingly vegetated (i.e., stable) because sand supply from the bluff diminished (c.f. Anderton and Loope 1995). As noted previously, it is conceivable that some sand was supplied to the dune field during this stabilization interval as a direct result of Algoma bluff undercutting. Final stabilization of the dune field may have also been delayed in space and time because strong winds across Lake Superior could have locally reworked deposits of sand that were thinly vegetated within the dune field. Such variation exists in coastal dunes along the eastern shore of Lake Michigan, where dune fields that are essentially stable still contain some active blowouts (Arbogast unpubl. data).

In summary, although some limited amount of sand may have been supplied to the dune field during the Algoma lake stage, there are problems with this model because the Algoma was insufficiently high to consistently undercut the bluff and cause extensive enlargement of existing dunes. This process may have worked in conjunction with a gradual stabilization of the dune field following the Nipissing II regression, with local *reworking* of existing dunes also occurring until vegetation became dense enough to stabilize the entire field around 3 ka. Regardless of the process(es) involved, this study demonstrates that the Nodaway dune field stabilized between 3.7 and 3.0 ka. In addition to the absolute ages that were obtained, the soils evidence (i.e., POD Indices, B horizon chemical composition) permits this chronology by indicating that dunes crests stabilized within an interval of several hundred years sometime after about 5 ka (Figs. 5 and 6). This study thus supports the basic premise of the perched dune model (Anderton and Loope 1995) that dune fields build when lakes are high, but implies that local variations may occur in dune fields after bluff source areas stabilize and sand supply diminishes.

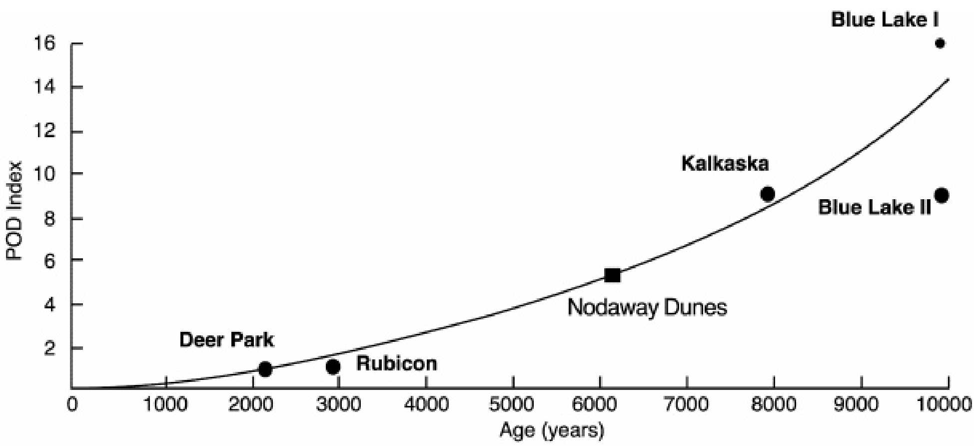

Figure 6: *Comparison of POD values derived from Spodosols of known age with soils in the Nodaway dune field (modified from Schaetzl and Mokma 1988).*

Conclusions

This study establishes the length of time that (perched) dunes at Nodaway Point have been stable by using absolute and relative dating methods. Dune building began sometime after ~7 ka, probably because the bluff adjacent to the dunes was undercut and sand supply to the plateau increased (e.g., Snyder 1985; Anderton and Loope 1995). Deposition of eolian sand dunes may have begun during the Nipissing transgression, but was probably most intense when the bluff was actively undercut during the Nipissing high stand (Saarnisto 1974b; Farrand and Drexler 1985) around 5,500 yrs B.P. (e.g., Anderton and Loope 1995). OSL and radiocarbon dating indicates that the most recent accumulation of eolian sand occurred between ~3.7 and 3.0 ka. Correlations between soils data from this study and soils of known age permit the age conclusion. Moreover, the chronology derived here suggests that the dune field may have stabilized gradually in association with some potential sand influx during the Algoma period. ■

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