SOIL DEVELOPMENT AND SPATIAL VARIABILITY ON GEOMORPHIC SURFACES OF DIFFERENT AGE

Linda R. Barrett and *Randall J. Schaetzl*
Department of Geography
Michigan State University
East Lansing, Michigan 48824-1115

Abstract: Twenty-four soil pedons on each of four sandy lake terraces in northwestern lower Michigan that ranged in age from 3000 to 11,000 years BP were studied to assess trends in soil morphological variability with time. After verifying the general uniformity of parent materials within and between the four surfaces, we examined temporal trends in the spatial variability of soil color, cementation, horizon thickness and development. E horizons attained high color values (lightness) by 3000 years and changed little after that time, whereas B horizons continued to get darker with time. Cementation within B horizons increased in strength and amount with time, as did B horizon thickness. Soils ≥4000 years old had deeper eluvial zones but much greater variabilities in the thickness of that zone than did younger soils. Soil development increased with time, but spatial variability in degree of development also increased with time. These patterns are best explained by invoking spatially random soil mixing upon a surface that is otherwise undergoing podzolization. [Key words: podzolization, chronofunction, Michigan, soil variability, soil genesis.]

INTRODUCTION

Soil spatial variability occurs at many scales, and on all landscapes (Beckett and Webster, 1971; Campbell, 1979). Numerous studies of soil variability have documented the highly complex, spatial interrelationships among soil characteristics and landform, drainage, land use, and other environmental factors (e.g., Protz et al., 1968; Asady and Whiteside, 1982; Beatty, 1987; Grigal et al., 1991). Wilding and Drees (1983) noted that spatial variability of soils is due to both systematic and random processes. Systematic variability is defined as gradual or marked changes in soil properties that can be explained by landform, geomorphic elements, soil-forming factors, or human activities. Random patterns of soil variability cannot be related to a known cause; although usually viewed as being "random," the lack of an observable, systematic pattern is more likely due to a lack of detailed soil information (Wilding and Drees, 1983). Causes of spatial variability include changes in the following variables: (1) parent material, (2) geomorphic and pedogenic processes, (3) flora and faunal activity, (4) microclimate, (5) hydrology, and (6) land use (Låg, 1951; Huggett, 1976; Alexander, 1986; Johnson et al., 1987; Schaetzl, 1990). Recent advances have enabled improved assessment of the degree of variability in the field (e.g., Truman et al., 1988), as well as in the analysis of field-

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derived data on soil variability (Norris, 1971; Crosson and Protz, 1974; Courtney and Nortcliff, 1977; Trudgill, 1983).

Although the spatial variability of soils has often been examined as a function of landscape position and landform type (Hall and Olson, 1991), surprisingly little work has been attempted on the temporal changes in soil spatial variability on a given landscape or geomorphic surface. Franzmeier and Whiteside (1963) studied soils on similar landforms only 60 km to the east of our study area and assessed soil spatial variability in a qualitative manner. They found little spatial variability on young (< 3000 BP) surfaces and considerable variability on older (> 8000 BP) surfaces, suggesting that variability increases with time. Busacca (1987) found that coefficients of variation (CV = standard deviation/mean) for soils on terraces in California decreased with time. Only two to four samples, however, were taken from each surface. Sondheim and Standish (1983) used multivariate techniques to describe soil development on recent moraines in British Columbia; variability of these soils was primarily attributed to the parent material, which had a masking effect on depth trends in chronofunctions. Harrison et al. (1990) found considerable variation in soil development within apparently isochronous surfaces, much of which they considered random because no pedogenic processes could be isolated to explain it at the scale of their study. They suggested that the existence of this variability may invalidate the use of chronofunctions as a means of describing time-dependent pedogenic changes.

We examined soils on four geomorphic surfaces to (1) assess how soil development and spatial variability change through time and (2) estimate the processes or vectors that may be most responsible for affecting such changes. In this sense, we have quantitatively examined temporal changes in development and spatial variability by substituting space for time in an ergodic hypothesis conceptual framework (e.g., Welch, 1970; James, 1988).

STUDY AREA

The study area is located in Emmet County, in the northwestern portion of Michigan's lower peninsula, near the shore of Lake Michigan (Fig. 1). The Algoma terrace study site is within Wilderness State Park; the Nipissing, Battlefield, and Main Algonquin terrace sites are clustered near Cross Village, 10 km to the south.

Surface Ages

The soils in the study area are developed in sandy sediments that were deposited in nearshore environments. Four lake terraces are prominent within the study area and can be intermittently traced along much of the Lake Michigan shoreline (Fig. 1), especially in the northern part of the southern peninsula (Leverett and Taylor, 1915). The terraces are a result of lake level changes in the Michigan-Huron basin during and following final deglaciation of the region, which began soon after the peak of the Greatlakean ice advance around 11,500 BP (Futyma, 1981). Glacial Lake Algonquin was the first lake to form. The highest shorelines of the Algonquin group, known as Main Algonquin shorelines, date from about 11,000 BP (Futyma,

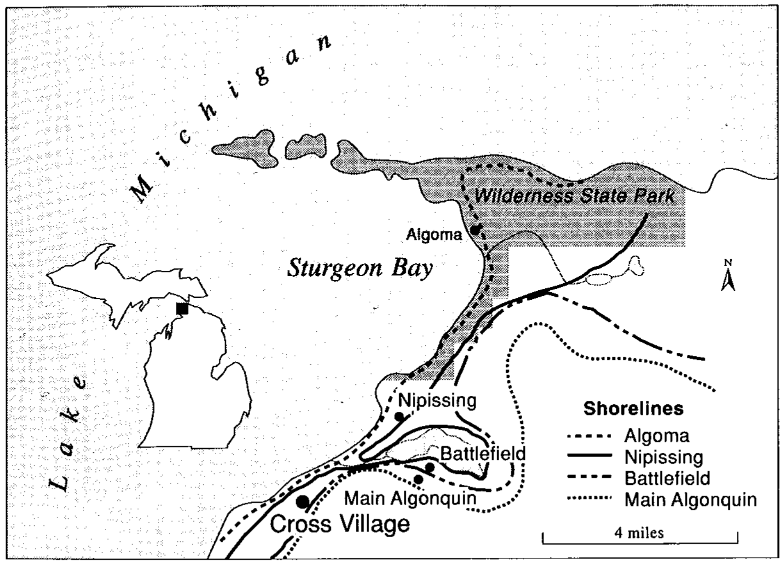

Fig. 1. Study area and sample site locations. Blocks of sample points were established at these locations (Emmet County, Michigan): Algoma: SW 1/4, section 29, T39N, R5W (3 blocks); Nipissing: SE 1/4, section 24 (2 blocks) and NE 1/4, section 26 (1 block), T38N, R6W; Battlefield: NE 1/4 (1 block) and NW 1/4 (1 block), section 36, T38N, R6W, and NW 1/4, section 31 (1 block), T38N, R5W; Main Algonquin: SW 1/4, section 31, T38N, R5W (3 blocks).

1981). The retreating ice margin to the north subsequently exposed a lower outlet and allowed the water level to drop, resulting in a series of shorelines, one of which is known as the Battlefield shoreline (Larsen, 1987). Around 10,000 BP deglaciation of the North Bay outlet of the Lake Huron basin, which was lower than previous outlets, caused water levels in the basin to drop drastically; the level of the resultant Lake Chippewa (Lake Michigan basin) was at least 61 m, and possibly as much as 107 m, below current levels (Hansel et al., 1985). After this low water period, water levels again began to rise due to isostatic rebound of the North Bay outlet. Continuing uplift at North Bay brought about the opening of the Port Huron outlet of Lake Huron at about 6000 BP, and marked the end of the low water phases (Larsen, 1987). The ensuing high water stage is known as the Nipissing Great Lakes in both the Lake Michigan and Lake Huron basins (Leverett and Taylor, 1915). Finally, the name "Algoma" is generally applied to any outstanding shoreline feature below the Nipissing in both the Lake Michigan and Lake Huron basins, and is thought to correlate to a high water stand at 3200–3000 BP (Larsen, 1987).

While recognizing that a small range of age estimates has been reported for each of the four geomorphic surfaces, we have chosen to use the following ages in this study, most of which are near the center of the ranges found in the literature: Algoma: 3000 years BP, Nipissing: 4000, Battlefield: 10,000, Main Algonquin: 11,000. We have assumed that pedogenesis began on each surface immediately following the close of the associated lake stage and the subaerial exposure of the sediments.

Topography, Parent Material, and Soils

Slopes within the study area are level to gentle (<6%), except where occupied by sand dunes. In some places, especially at the Battlefield study site, there are many small pits and mounds up to 2 m in diameter, indicative of tree uprooting (Schaetzl et al., 1989). The soils are well or excessively drained, although non-sampled areas on the Algoma surface are somewhat poorly or poorly drained.

The parent materials in the study area are primarily lacustrine sands. Although they are not common, the abundance of coarse fragments (> 2.0 mm) varies considerably both between and within pedons. The Algoma site is underlain at >150 cm by very gravelly calcareous sand.

Podzolization is a dominant soil-forming process in this region (Schaetzl and Isard, 1991). On upland surfaces underlain by coarse-textured glacial tills and outwash, Spodosols (Haplorthods) are common. On finer parent materials, podzolization and lessivage processes have led to the development of Alfisols (Eutroboralfs) and Inceptisols (Eutrochrepts). Histosols are common in closed depressions.

Vegetation and Climate

During the past 11,000 years, temperatures have at times been cooler, and at times warmer, than those of the late twentieth century, probably affecting vegetation patterns and soil development slightly. The initial post-glacial vegetation in the area was probably dominated by spruce (*Picea* spp.) and pine (*Pinus* spp.), and was in turn succeeded by a pine-dominated assemblage (Futyma and Miller, 1986). Subsequent assemblages with lower pine percentages and more sugar maple (*Acer saccharum*), birch (*Betula papyrifera*), beech (*Fagus grandifolia*) and hemlock (*Tsuga canadensis*) are more characteristic of the later warmer climates (Miller and Futyma, 1987). Regional temperatures peaked during the climatic optimum, ca. 6000 BP (McAndrews, 1979). Despite these variations in vegetation patterns, palynological studies in the northern Lower Peninsula reveal a significant presence of pine and other evergreen species throughout the Holocene (Futyma and Miller, 1986; Miller and Futyma, 1987).

The present climate in Emmet County reflects its proximity to the waters of Lake Michigan. Mean annual temperature (1951–1980) at Pellston is 5.2°C, with July and January temperatures averaging 18.6° and -9.1°, respectively. Mean annual precipitation is 83 cm, with the highest monthly average occurring in September (10.0 cm); February is the driest month (3.9 cm). Mean annual snowfall is 298 cm. Climate at the study sites is modified because of their greater proximity to the lake; somewhat lower precipitation amounts and more moderate temperatures are common at nearshore locations. Data for Cross Village, located near the study sites, show a mean annual precipitation of 73 cm and snowfall of 200 cm. Temperature data are not available for Cross Village.

Current vegetation within the study area consists primarily of second-growth mixed deciduous and coniferous forest. Common tree species at the study sites include, in approximate order of abundance, aspen (*Populus* spp.), paper birch, white pine (*Pinus strobus*), red maple (*Acer rubrum*), red pine (*Pinus resinosa*), black oak (*Quercus velutina*), red oak (*Quercus borealis*), beech, and sugar maple. On the wetter Algoma surface balsam fir (*Abies balsamea*), paper birch, aspen, northern white cedar (*Thuja occidentalis*), striped maple (*Acer pensylvanicum*), red maple, and white spruce (*Picea glauca*) assume greater importance.

METHODS

Upper elevation limits for each geomorphic surface at locations in Emmet, Charlevoix, Cheboygan, and Presque Isle counties were taken from published reports (Futyma, 1981; Leverett and Taylor, 1915; Larsen, 1987). The study area was chosen such that sites were located on successive surfaces in sufficiently close proximity so as to minimize differences in parent material, vegetation, and climate. Only sites with slopes of $\leq$ 5%, containing well-drained, uncultivated soils were chosen for study.

Although some parts of each geomorphic surface were deemed unsuitable because of wetness or parent material textures that deviated significantly from those in other areas, at least two sampling sites were chosen for study on each surface except the Algoma, which had a very small area suitable for sampling. At each site, a systematic sampling scheme (Schellentrager and Doolittle, 1991) was utilized to define a 3 x 3 grid of sampling points, with grid points spaced 50 m apart. Where the location of roads or the boundaries of the surface did not permit the establishment of a full 3 x 3 grid, the shape of the grid was modified, but a spacing of at least 25 m between points was maintained. An additional sampling site on each surface was examined two years later using similar methods, except that a 25 m spacing was used between all points and a 2 x 3 grid was used, to bring the total number of points sampled from each surface to 24. In all cases, if a sample point fell too close to a tree or was located on or near a treethrow pit or mound, it was avoided in favor of the nearest representative point. The elevation of each point was estimated from topographic maps. At each point a bucket auger was used to sample the soil to the C horizon or to a depth of about 150 cm, whichever was shallower. Color (moist), texture, and thickness of horizons were recorded, as well as the presence of coarse fragments or ortstein (soil cemented by aluminum, iron, and organic matter) in each horizon. The 25–50 m spacing between points was used to provide coverage of relatively large areas while at the same time minimizing covariation between points. The inclusion of three separate blocks (grids) of points per surface was intended to maximize potential variability.

The POD Index is a numerical index of soil development developed for podzol soils. It was calculated for each sampled point following the method of Schaetzl and Mokma (1987). The POD Index is based on the assumption that podzolization processes tend to produce soils that have E horizons that thicken and increase in color value with time, while B horizons thicken and become redder and darker. Therefore, the Index depends on E and B horizon colors and the number of B subhorizons. Linear regression analyses (chronofunctions) were developed using surface ages reported above. Analysis of variation (ANOVA) testing was performed using Fisher's Least Significant Difference (LSD) test.

RESULTS AND DISCUSSION

Parent Material Uniformity

We used three indicators to examine the uniformity of the parent materials among sampling points, both within and between surfaces: (1) C horizon texture as determined in the field, (2) content of coarse fragments in the solum, and (3) C horizon moist color. If these indicators were not dissimilar within and between surfaces, we assumed that current soil variability was due to pedogenesis rather than parent material effects.

Parent material textures can generally be considered uniform both within and between surfaces, as the C horizon of each pedon examined had a sand texture. Medium sands dominated (Barrett and Schaetzl, 1992); occasional variations in the proportions of finer to coarser sands were noted but not quantified.

The abundance of coarse fragments varied considerably both within and between surfaces (Table 1). Coarse fragments were noted in at least one horizon of at least one pedon on all surfaces, although usually in fewer than half the horizons of any given pedon. On the Algoma surface, most coarse fragments were noted in connection with a gravelly 2C horizon that underlies the relatively gravel-free surficial deposits. Soils on the Nipissing surface contained more coarse fragments than soils on the other surfaces, and, in a few cases, were so abundant as to obstruct augering. In general, coarse fragments on the Nipissing surface were encountered below the A and E horizons, but could not be attributed to a lithologic discontinuity as on the Algoma surface.

The effect that differences in coarse fragment distributions might have on soil development is unknown, though probably minor. Although surface rocks have been shown to affect heat and water gradients in soils (Jury and Bellantuoni, 1976), the coarse fragments in these soils were small and not generally found at the surface. Abundant coarse fragments may affect the movement of water through the solum and greatly reduce the particle surface area available for chemical processes (Schaetzl, 1991), but the extent to which the small amount of coarse fragments found in these soils could influence pedogenesis is unclear.

C horizon colors were quite similar, both within and between surfaces (Table 1). The Munsell hue for the C horizon was, in all cases, 10YR; Munsell values of 5 or 6 were most common. On the Algoma surface, the mean chroma of the C horizon colors was somewhat lower than that of the other surfaces, but the predominant hues and values were similar to those of the other surfaces. C horizon color values averaged slightly lower on the Nipissing surface than on the other surfaces, due to the presence of a number of pedons with a color of 10YR 4/4. Variability in C horizon color appears to be no greater between surfaces than within surfaces. Analysis of variance testing of color data was not performed due to the nonlinearity of the Munsell color system.

Change in Soil Properties with Surface Age

Within a surface of assumed uniform age, soil variability is related to factors other than age. Several workers (Franzmeier and Whiteside, 1963; Soil Survey Staff, 1975;

Table 1. Selected Morphological Characteristics of Soils

Characteristic	Algoma	Nipissing	Battlefield	Main Algonquin
Coarse fragments				
Frequency in any horizon[a]	29%	92%	50%	63%
Frequency in majority of horizons[b]	17%	42%	8%	8%
Color				
Mean C horizon value[c]	5.7	5.2	5.6	6.0
Mean C horizon chroma[c]	3.6	5.2	5.6	5.3
Mean "highest value" E horizon value[c]	4.6	4.8	4.6	4.7
Mean "highest value" E horizon chroma[c]	1.9	2.0	0.9	2.0
Mean lowest-value B horizon value	4.7	4.0	3.5	3.3
Mean lowest-value B horizon chroma	3.6	4.9	0.2	3.6

[a]Percentage of pedons in which coarse fragments were present in any horizon.
[b]Percentage of pedons in which coarse fragments were present in $>50\%$ of horizons.
[c]Hue is 10YR for all observations.

Evans and Cameron, 1985; Schaetzl and Mokma, 1988) have found that the morphological data that vary most on a surface containing sandy, podzolized soils include (1) horizon color, (2) frequency and degree of cementation of ortstein (cemented spodic horizon materials; Bsm or Bhsm horizon), (3) depth to the top of the B horizon (thickness of the eluvial zone), (4) solum thickness, and (5) POD Index.

The moist colors of the E horizon were very similar for all pedons, both within and between surfaces (Table 1); 10YR 5/2 and 10YR 4/2 were most common. The hue, in all cases, was 10YR (in the case of a pedon with more than one E subhorizon, the subhorizon with the highest value was used in calculations for this section, as *per* Schaetzl and Mokma, 1988). The similarity of E horizon colors probably reflects the color of the sand grains, which are predominantly composed of quartz. Assuming that the differences between E and C horizon colors are primarily a result of the loss of Fe-based coatings from the individual grains of the E horizon, once the coating is lost, the color of the E horizon is unlikely to change further, except perhaps in response to small additions of organic matter from the A horizon. E horizons on the Algoma surface are characteristically somewhat thinner than E horizons on the older surfaces.

On the Algoma surface all B horizons had hues of 10YR (Fig. 2a). The percentage of pedons with the lowest-value (darkest) B subhorizon redder than 10YR increases with surface age (Fig. 2b, c, d). On the three oldest surfaces, the mean color value of the darkest B subhorizon decreases with surface age (Table 1). Often

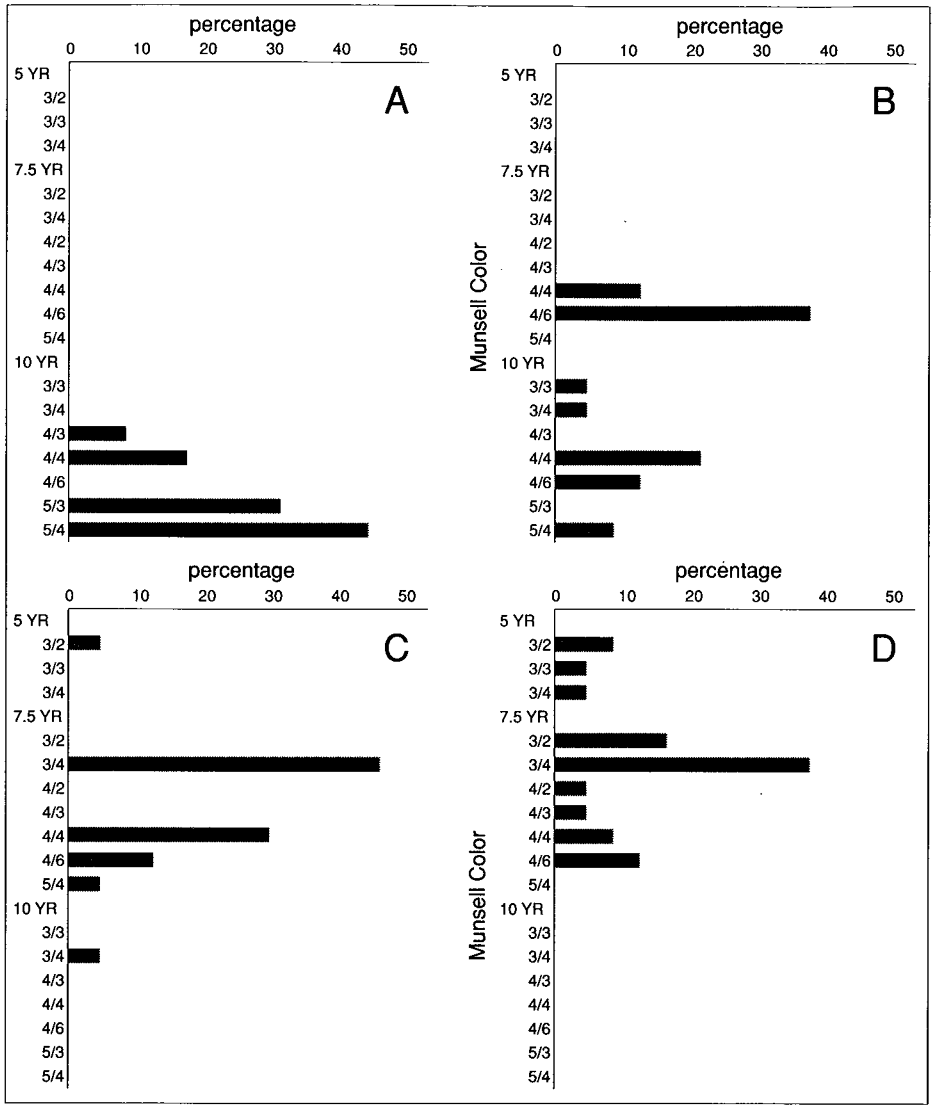

Fig. 2. Munsell colors of lowest-value B subhorizons in pedons of successive terrace deposits: a) Algoma; b) Nipissing; c) Battlefield; d) Main Algonquin.

the lowest-value B subhorizon also had the reddest hue. The redder hues and lower values of the B horizons on the older surface can be attributed to the accumulation of Fe in the B horizon through illuviation (Evans and Cameron, 1985; Barrett and Schaetzl, 1992). The B horizons of the older surfaces also have generally lower chromas (Table 1), which can be attributed mainly to the accumulation of organic C in the B horizon (Evans and Cameron, 1985; Barrett and Schaetzl, 1992).

Ortstein (cemented Bsm or Bhsm horizon material) is a characteristic of spodic horizons that often indicates strong development (Wang et al., 1978). In this study, the number of pedons in which ortstein was encountered in at least one horizon increased with increasing surface age. Ortstein was not observed on the Algoma surface, while 8% of the pedons on the Nipissing surface showed weakly developed ortstein in at least one B subhorizon. Ortstein was noted in 42% and 88% of the pedons on the Battlefield and Main Algonquin surfaces, respectively. In general, the strength of ortstein cementation also tended to increase with surface age.

On the Algoma surface, the B horizon was encountered at much shallower depths than on the older surfaces (Fig. 3a). Two pedons on the Algoma surface lacked an identifiable E horizon, such that the B horizon was encountered directly below the A horizon; one pedon lacked an identifiable B horizon. Although the mean and median depths to the top of the B horizon appeared to increase slightly with surface age for the soils on the three oldest surfaces, an analysis of variance test revealed significant differences only between the means of the Nipissing and Main Algonquin surfaces, and between the mean of the Algoma surface and that of each of the older three ($p < 0.05$) (Fig. 3a). On the three older surfaces, the range of depths to the B horizon was quite large within each surface but varied little among the surfaces. It is likely that the large range of depths to the B horizon on these surfaces was partly due to boundary irregularity, tongueing, and waviness (e.g., Låg, 1951; Schaetzl, 1990), which cannot be detected by augering. The increase in depth to the B horizon was statistically significant when regressed against surface age ($p < 0.001$), but due to the wide scatter on each surface, the association was weak (adjusted $r^2 = 0.21$).

Previous chronosequence studies of podzolic soils in sandy materials have reported conflicting results regarding the trend in depth to the top of the B horizon. Franzmeier and Whiteside (1963) showed that, following an initial rapid increase in depth for the first 3000 years of development, this trend either leveled off or reversed, such that the E–B horizon boundary became gradually shallower with time. Similarly, Nørnberg (1977) reported that the upper boundary of the B horizon became nearer the surface with increasing age in podzols in Denmark. In Finland, however, a positive correlation existed between soil age and thickness of the E horizon (Koutaniemi et al., 1988).

Measures of B-horizon thickness (mean, median, minimum, maximum) all increase with increasing surface age ($p < 0.0001$; adj. $r^2 = 0.36$; Fig. 3b). An analysis of variance test showed statistically significant ($p < 0.05$) differences in mean thicknesses between all pairs of surfaces except the Algoma and Nipissing surfaces. This finding is not unexpected, however, because the Algoma and Nipissing surfaces are separated by only a small age difference. Increases in B horizon

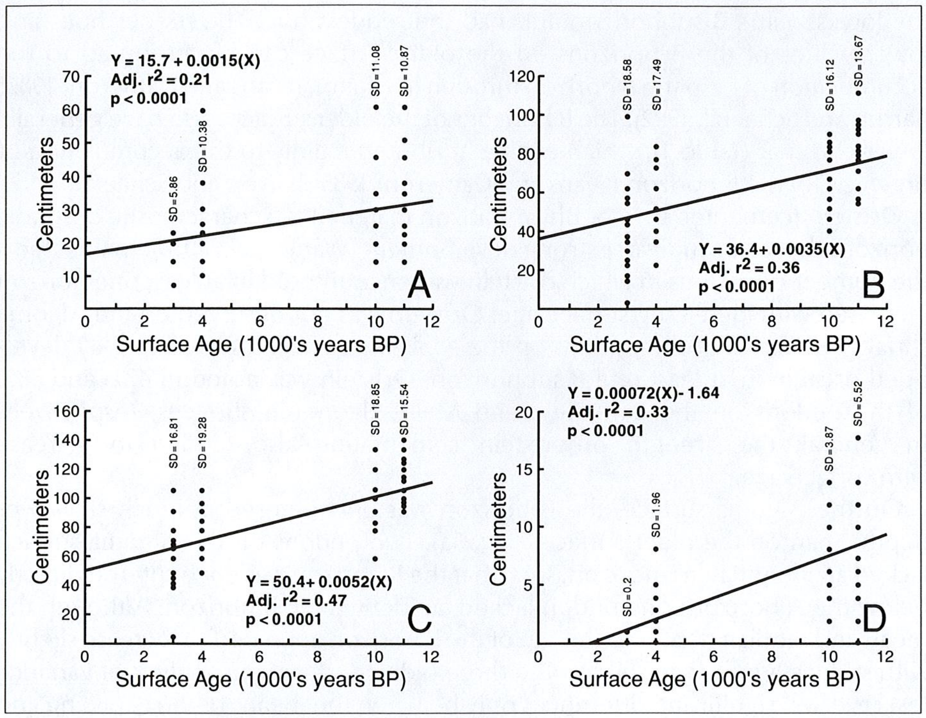

Fig. 3. a) Depths to top of the B horizon as a function of surface age; b) B horizon thickness as a function of surface age; c) solum thickness as a function of surface age; d) POD Index as a function of surface age.

thickness with increasing age have also been reported for podzols by Koutaniemi et al. (1988).

Solum thickness also increased with time ($p < 0.001$; adjusted $r^2 = 0.47$; Fig. 3c). Analysis of variance results demonstrate that the mean solum thickness on each surface was significantly different ($p < 0.05$). Koutaniemi et al. (1988) found a similar trend; analogous findings have been reported for soils in calcareous sands in southern Michigan (Olson, 1958) and for beach ridge soils near Hudson Bay, Ontario (Protz, 1984).

The POD Index is a field-based index of soil development that is designed to estimate the strength of spodic morphology expression in soils (Schaetzl and Mokma, 1988). The mean, median, and maximum values of the POD Index increase with surface age (Fig. 3d). The range and standard deviation of POD Index values also increase, in part due to the large number (23 out of 24) of pedons with a POD Index = 0 on the younger surfaces. Nonetheless, a POD Index of zero can denote a wide range of development. POD Index values of 0–2 merely suggest that the soils are non-Spodosols (in this case, Psamments). A regression of POD Index against surface age produced an equation significant at $p < 0.0001$, but due to the wide variability in values on each surface, the adjusted r^2 is low (0.33; Fig. 3D). Schaetzl and Mokma (1988) provide guidelines for assigning pedons to probable Soil Taxonomy classifications based on POD Index data: 0–2 (non-Spodosols), 2–6

Table 2. Probable Subgroup Classifications of Pedons Based on POD Index Values

Soil Taxonomy[a] POD Index value:		Udipsamments 0–2		Entic Haplorthods 2–6		Typic Haplorthods ≥6	
	N	Freq	%[b]	Freq.	%	Freq.	%
Algoma	24	24	100	0	0	0	0
Nipissing	24	17	70	12	50	1	4
Battlefield	24	10	42	14	58	8	33
Main Algonquin	24	4	17	11	46	13	54

[a]Taxonomy follows Schaetzl and Mokma (1988) and Soil Survey Staff (1992).

[b]Due to overlap in POD Index values between categories, totals may sum to > 100%.

(Entic intergrades of Spodosols), 6+ (central concept Spodosols). Thus, based on the POD Index value, the pedons on these four surfaces have been assigned to probable subgroup classifications (Soil Survey Staff, 1975) (Table 2). Using this method, the percentage of pedons on each surface that classify as Spodosols increases with surface age.

Soil Variability on Each Surface

An examination of the variability statistics for depth to the top of the B horizon revealed conflicting trends (Table 3). The coefficient of variation (CV) of the depths to the top of the B horizon is greatest on the Nipissing surface, whereas the standard deviations (SD) of the same data showed a marked minimum on the Algoma surface due to the low mean value (Table 3). The 25th percentile–75th percentile range showed slight increases in variability with age. The minimum-maximum range, like the SD, was noticeably lower on the Algoma surface than on the three older surfaces. Most of this variability comes from a few outlier pedons on these surfaces. Based on these data, we conclude that variability in depth to the top of the B horizon does not appear to be time dependent, with the possible exception of the Algoma surface of 3000 years.

Variability in B horizon thickness decreased with time (Table 3). The wide range (min–max) in B horizon thickness on the Algoma surface is not observed when the 25th–75th percentile figures are used; a few pedons with exceptionally thin or thick B horizons are responsible for the high range of variability. Solum thickness also showed a general trend of decreasing variability with time (Table 3).

Although the CV for the POD Index decreased with surface age, the SD and range of the POD Index show an increasing trend due to increasing means (Table 3). The wide range and high SD of the POD Index on the Main Algonquin surface reflect the strong spatial disparity in soil development on the surface, from Entisol

Table 3. Variability Measures

Surface	*N*	CV	SD	Mean	Range	
					Min–Max	25th–75th pctl.
					.... cm	
Depth to top of the B horizon						
Algoma	24	0.34	5.86	17	18	8
Nipissing	24	0.42	10.38	25	50	10
Battlefield	24	0.39	11.08	28	53	8
Main Algonquin	24	0.33	10.87	33	46	13
B horizon thickness						
Algoma	22[a]	0.41	18.6	45	84	18
Nipissing	18	0.33	17.5	54	61	20
Battlefield	24	0.24	16.1	67	51	28
Main Algonquin	24	0.17	13.7	79	58	17
Solum thickness						
Algoma	22	0.26	16.8	64	69	20
Nipissing	18	0.24	19.3	79	56	30
Battlefield	24	0.20	18.8	96	76	25
Main Algonquin	24	0.14	15.5	112	53	22
POD Index						
Algoma	24	4.90	0.20	0.04	1	0
Nipissing	24	0.98	1.96	2.00	8	3
Battlefield	24	0.91	3.87	4.25	14	5
Main Algonquin	24	0.75	5.52	7.37	18	9

[a]N values $<$ 24 indicate that the bottom of the B horizon was not reached for one or more of the 24 pedons sampled.

to strongly developed Spodosol morphology. The contrast in range between the younger and older surfaces is exacerbated because a POD Index value of zero, common on the younger surfaces, can denote a fairly broad range of generally weak soil development.

Spatial Variability: Causes and Effects

Our data have shown that solum and B horizon thickness increase with time of soil development, but that the spatial variability in these parameters decreases with time, as measured by the CV. Based on the POD Index, ortstein abundance, and B horizon color, soil development also appears to increase with time, as does the spatial variability of these parameters.

Soil development and horizonation in this region generally proceed along definite pedogenic pathways over time, from Udipsamments to Typic Haplorthods after 11,000 years (Barrett and Schaetzl, in press). Variation in the spatial intensity of pedogenic processes, however, which can be caused by microclimate (Macyk et al., 1978), microrelief (Låg, 1951; Schaetzl, 1990), leaf-litter distribution (Rowe, 1955; Alexander, 1986); even the trees themselves (Gersper and Holowaychuk, 1971; Crampton, 1982; Boettcher and Kalisz, 1990), can affect soil morphology greatly. These effects may not be manifested morphologically or chemically for some time, as evidenced for example by the development of deep E horizon tongues below treethrow pits dated at 2010 years but the lack of such tongues below pits that are $<$ 200 years old (Schaetzl, 1987). Without due attention to such small-scale processes, the spatial variability that they can produce could be incorrectly assigned to random variation (Wilding and Drees, 1983).

Furthermore, "interruptions" in the pathways of soil development constantly occur over space and time (Johnson and Watson-Stegner, 1987). On these forested, sandy geomorphic surfaces, such interruptions take the form of pedoturbations and/or periods of erosion. Pedoturbation acts to regress soils by two major suites of processes: (1) floralturbation, especially by tree-root growth/expansion and uprooting, which can alter the horizonation and classification of whole pedons, and (2) faunalturbation, usually by insects and mammals, which typically affects only small portions of "randomly" located pedons. Pedoturbation processes often affect soil development on a micro-scale (Johnson et al., 1987) and may contribute to the increased variability in degree of soil development with time by regressing pedons of more developed and horizonated soils back to a simpler state (Schaetzl, 1986; Johnson and Watson-Stegner, 1987) while leaving adjacent pedons in a relatively undisturbed, and hence more developed, state. On younger surfaces, where most soils are Psamments, such events will rarely change the classification of a pedon. On older surfaces, small-scale pedoturbations can lead to changes in horizonation patterns and classification (Veneman et al., 1984; Schaetzl, 1991), which will lead to increased variability in POD Index values but may not affect such developmental parameters as ortstein abundance or B horizon color.

The frequency of soil disturbance by tree uprooting is affected by numerous variables, including soil morphology and chemistry (Cremeans and Kalisz, 1988; Schaetzl et al., 1989). Our field observations have repeatedly shown that frequency of uprooting on sandy, podzolized surfaces increases with spodic horizon development. Although we lack quantitative data on uprooting rates and pit/mound densities vs spodic development, in part because of the difficulties encountered in delineating pits and mounds in the field, it is highly likely that rates of uprooting, when expressed as area of soil surface disturbed yr^{-1}, increase with increasing soil age in the study area, and lead in large part to the high spatial variabilities in numerous soil characteristics on the older surfaces. Thus, it is not unexpected that on older surfaces, weakly developed soils can be found adjacent to better developed soils, whereas only weakly developed soils are present on younger surfaces. Because uprooting of even the largest trees typically distorts only the A, E, and upper B horizons of Spodosols (Schaetzl, 1986), scattered events of this type will not directly affect solum thickness, thereby allowing solum thickness to continue to increase through time and causing minimal increases in its spatial variability. Thus, we conclude that the patterns of temporal and spatial variability observed on these surfaces are explainable by invoking microscale variability in spatial intensity of pedogenic processes and spatially random pedoturbations, especially tree uprooting, upon a surface that is otherwise undergoing broad-scale podzolization.

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