A CHRONOSEQUENCE OF SOILS ON LAKE TERRACES IN NORTHERN MICHIGAN

By

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ABSTRACT

A CHRONOSEQUENCE OF SOILS ON LAKE TERRACES IN NORTHERN MICHIGAN

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A chronosequence of well drained soils on four geomorphic surfaces in northwestern lower Michigan was investigated to assess rates of podzolization and changes in spatial variability of soils with age. The geomorphic surfaces were sandy terraces of proglacial and Holocene lakes in the Lake Michigan basin, with ages of 3000, 4000, 10,000, and 11,000 BP. On each surface between 10 and 24 pedons were examined to assess the spatial variability of morphological properties; a single modal pedon from each surface was sampled for laboratory analysis.

Soils on the various surfaces were classified as follows: 3000 BP, Typic Udipsamments; 4000 BP, Spodic Udipsamments; 10,000 BP, Entic Haplorthods; and 11,000 BP, Typic Haplorthods. Thus, more than 4000 but less than 10,000 years are required for the development of spodic morphology in Michigan.

Spatial variability of the soils generally increased with surface age. With increasing surface age, amounts of OC and extractable Fe and Al in the B horizon, as well as silt content of surface horizons, increased. Chronofunctions were developed using linear regression techniques.

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INTRODUCTION

Jenny's (1941) functional-factorial model of soil development provides a theoretical framework within which to examine separately the effects of each of five factors (climate, organisms, relief, parent material, and time) on the soil environment. The factors are considered to be independent, in that sites can be found where they appear to vary independently (Birkeland 1984b), although the strict independence of the factors has been questioned (Crocker 1952). If four of the factors are held constant, or if their variation is negligibly small, the functional relationship between the remaining factor and soil properties can be examined (Yaalon 1975).

In the northern part of Michigan, many soils occurring in well drained sandy parent materials commonly are Spodosols, provided the surface is old enough for spodic features to develop. The effects of time on the development of soils can be isolated where geomorphic processes have left a sequence of surfaces of different age, in similar parent materials (Yaalon 1975; Vreeken 1975).

During and following the final deglaciation of Michigan's southern peninsula, the opening of successively lower outlets and further downcutting of existing ones resulted in declining lake levels in the Great Lakes basins (Larsen 1987; Hansel and Mickelsen 1988), and the formation of a series of lake terraces (Leverett and Taylor 1915). This sequence of surfaces offers a unique opportunity to examine the development of Spodosols in Michigan. It also presents the possibility

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of examining the effect of time on the variability of properties within a surface; soils do not develop uniformly across an entire surface, but instead exhibit variability of properties within a surface (Switzer *et al.* 1988). The relationship between soil variability and duration of soil development is poorly understood (Harrison *et al.* 1990).

The purpose of this study is to examine the relationship between soil development and surface age on sandy parent materials in the northwestern part of Michigan's southern peninsula by 1) examining field-acquired morphological characteristics of soils on a sequence of four lake terraces and describing the range and variability of these properties both within and between surfaces; 2) assessing how surface age relates to variability of selected soil properties; 3) determining physical and chemical properties of typical pedons on each surface; and 4) assessing rates of soil development in the chronosequence by developing chronofunctions using linear regression techniques. Numerical indices of soil development are used to evaluate the degree of development of the pedons.

In accordance with features common to Spodosols, I expect to find that, when considered from the youngest surface to the oldest: 1) E horizon color will become lighter; 2) B horizon color will become redder and darker; 3) the thickness of the eluvial horizons and total solum thickness will increase; 4) organic C (OC) and extractable Fe and Al content of the B horizon will become greater as eluvial horizon content of these elements becomes less; and 5) variability of soil properties within a given surface will increase.

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REVIEW OF LITERATURE

Late Pleistocene and Holocene history of Emmet County

Glaciation and surficial deposits

The landforms of Emmet County have been largely shaped by late Wisconsinan deglaciation and Holocene lake level fluctuations (Leverett and Taylor 1915; Farrand and Eschman 1974). Few detailed studies of the geomorphology of Emmet County itself have been undertaken since that of Leverett and Taylor (1915), besides those of Melhorn (1954) and Spurr and Zumberge (1956). A general history of deglaciation events in the county can nevertheless be inferred from data gathered elsewhere in the Great Lakes area, as many events were contemporaneous throughout the region.

The Wisconsinan glacier reached its maximum extent by about 18,000 BP, and by 16,000 BP a general stage of retreat had begun (Farrand and Eschman 1974). This general retreat was punctuated by periodic readvances or stillstands of the ice margin that helped form the end moraines that are prominant in Michigan's topography (Wright 1971; Fullerton 1980). The entire state of Michigan was free of ice by around 10,000 BP (Farrand and Eschman 1974).

Although various researchers have subdivided the late Wisconsinan stage differently, I will, in general, follow the nomenclature conventions of Evenson *et al.* (1976), who designated two substages of general glacial advance:

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advance: the Woodfordian (ending around 12,300 BP), and the Greatlakean (11,800 to 10,000 BP), which were separated by a single substage of glacial retreat (the Twocreekan; 12,300 to 11,800 BP). As regards chronology, however, I will generally follow the interpretations provided by Fullerton (1980). This discussion will be limited to events of the Lake Michigan lobe after about 16,000 BP.

The first major episode of this period was the formation of the Kalamazoo-Mississinewa moraines about 14,800 BP. These moraines apparently mark the limit of a major readvance (Burgis and Eschman 1981). After this event, the ice margin retreated regularly, with only minor oscillations, until it stood at the position of the Lake Border moraines around 14,300 BP (Farrand and Eschman 1974; Fullerton 1980; Burgis and Eschman 1981).

Following the formation of the Lake Border moraines, the ice margin retreated rapidly during the time known as the Mackinaw Interstadial or Cary-Port Huron interstade (13,500-13,000 BP), such that the majority of the southern peninsula was free of ice. The Mackinaw Interstadial is considered to an interlude of glacial retreat within the larger Woodfordian substage of general glacial advance. Evenson *et al.* (1976) believed that the Straits of Mackinac were ice-free by around 13,300 BP, based on evidence provided by a thin bed of mosses (bryophytes) under a till layer in Cheboygan County, about 15 km south of the straits. They obtained several widely divergent radiocarbon dates for this bryophyte layer (Fullerton 1980); those accepted by Farrand *et al.* (1969) placed the age of the mosses between 13,300-12,500 BP, within the Mackinaw Interstadial. Other radiocarbon dates, however, suggest that this bryophyte bed is actually much younger, from around

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10,750 BP (Fullerton 1980). If the latter dates are accepted, the age of the bryophyte bed is similar to that of the Two Creeks forest in Wisconsin (Fullerton 1980), a period when the Straits of Mackinac had previously been assumed to have been ice-free. Thus, although the ice margin retreated significantly during the Mackinaw Interstadial, if the younger dates for the bryophyte bed are accepted, the straits were not ice-free at this time and the northernmost portions of the southern peninsula remained under ice.

A period of glacial advance followed the retreat of the Mackinaw Interstadial. By 12,950 BP, the ice margin had readvanced to the position of the Outer Port Huron moraine; the Inner Port Huron moraine was formed around 12,800 BP, and the Manistee moraine also dates from around the same time (Fullerton 1980). These moraines, located to the south of the study area, are the last such features formed during the Woodfordian substage.

The Twocreekan substage which follows (12,300 to 11,800 BP) was characterized by general and continued glacial retreat. Fullerton (1980) believed that the Straits of Mackinac were deglaciated during this time, although others (Farrand and Eschman 1974; Evenson *et al.* 1976) concluded, based on the single till layer overlying the Cheboygan Bryophyte Bed, that the ice margin remained farther south during this time, and that drainage of water from the Lake Michigan basin occurred through the Indian River Lowlands, about 40 km to the south of Mackinaw City. If the younger dates for the bryophyte bed are accepted, however, it is likely that the Straits were ice-free during the Twocreekan substage.

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After the Twocreekan retreat, the ice front readvanced far enough to override the Two Creeks forest bed in eastern Wisconsin about 11,800 BP (Burgis and Eschman 1981). The extent of this Greatlakean readvance in northeastern Michigan has recently been revised; it is now thought that the limit of this advance was still well inside the position of the Port Huron moraine (Burgis and Eschman 1981). The precise extent of this readvance in the northwestern part of the peninsula has not been determined, since the data gathered by Melhorn (1954) have been reinterpreted in the light of later research (Evenson *et al.* 1974, Burgis and Eschman 1981). It is likely, however, that most of western and northwestern Emmet County was affected by this advance.

The final deglaciation of the northwestern Lower Peninsula took place when the Greatlakean ice margin retreated from the area. The Greatlakean ice was apparently thin, and seems to have stagnated *in situ*, such that terminal and recessional moraines were not developed (Spurr and Zumberge 1956), and preexisting landform features were modified to a relatively minor extent. Instead, ice-melt and ice-contact features are prominent in higher elevations above the level of the high lake stages in the Lake Michigan basin. Buried and melting ice blocks also played a role in the formation of many of the area's inland lakes. including Douglas Lake, Burt Lake, and Black Lake in Cheboygan County (Spurr and Zumberge 1956). Lower elevation features were further modified by the high lake stages in the Lake Michigan basin that followed, as discussed in the following section.

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Glacial lake level fluctuations

Final deglaciation of the northwestern Lower Peninsula began soon after the peak of the Greatlakean ice advance around 11,500 BP (Futyma 1981). The Straits of Mackinac were ice free by about 11,200 BP (Larsen 1987), allowing the water in the Lake Michigan and Huron basins to become confluent and forming Glacial Lake Algonquin (Leverett and Taylor 1915). This lake had several substages, the first of which has been named the Main Algonquin stage (Hansel *et al.* 1985).

The Main Algonquin shoreline can be traced throughout the southern peninsula of Michigan (Leverett and Taylor 1915). Due to subsequent differential uplift of the land surface, the shoreline is tilted upward toward the north (Leverett and Taylor 1915). During Algonquin times, the peninsula was truncated south of Little Traverse Bay; the area that is now Emmet County formed an archipelago in the lake (Spurr and Zumberge 1956). The many rills and minor water courses characterizing the Main Algonquin shoreline suggest that the water level was stable for a period of time (Spurr and Zumberge 1956). The Main Algonquin stage ended about 11,000 BP (Futyma 1981), when the still retreating ice margin to the north exposed a lower outlet and allowed the water level to drop.

Below the Main Algonquin shoreline a series of uplifted shorelines can be intermittently traced around the entire northern Lake Michigan basin (Hansel *et al.* 1985). Each shoreline is believed to relate to successively lower northern outlet thresholds resulting from the retreating ice margin (Larsen 1987). No specific chronology is available for the lake stages represented by these shorelines, but it is thought that each stage was quite brief (Eschman and Karrow 1985). Naming and

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correlation of the shorelines is not well defined due to the brevity of each stage, but Leverett and Taylor (1915) identified Lower Algonquin, Battlefield, and Fort Brady beaches below the Main Algonquin beach in the northwestern portion of the peninsula. The beaches range in age from 11,200 to 10,000 BP (Larsen 1987).

Deglaciation of the North Bay outlet of the Lake Huron basin at approximately 10,000 BP marked the beginning of a new lake stage in the Great Lakes basins and the end of ice margin control of lake levels (Hansel *et al.* 1985; Larsen 1987). The North Bay outlet was much lower than any previous one, causing water level in the basins 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). The much-reduced water bodies in the Lake Michigan and Lake Huron basins were no longer confluent, and a river flowed through the Straits of Mackinac (Larsen 1987).

After this low water period, water levels in the two basins again began to rise, such that, by 8150 BP, the water in two basins was again confluent (Hansel *et al.* 1985; Larsen 1987). Isostatic rebound of the North Bay outlet provided the major control on water levels during this period. This means that the North Bay outlet, which had been severely depressed due to the weight of the glacier, began slowly to gain elevation after the glacier had left the area. As the land surface rebounded, and with no alternative outlet, the water level in the lakes rose. 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).

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The ensuing high water stage, termed the Nipissing Great Lakes in both the Lake Michigan and Lake Huron basins (Leverett and Taylor 1915), had its outlet incised into glacial drift at Port Huron (Hansel et al. 1985). Some researchers have divided the Nipissing into two substages: Nipissing I (5500 - 4500 BP) and Nipissing II (4500 - 4000 BP) (Larsen 1985a). Water levels were probably slightly higher during Nipissing I than during Nipissing II, and the two substages were separated by a short time of lower water levels (Hansel et al. 1985). The Nipissing shoreline is prominent throughout the southern peninsula of Michigan (Leverett and Taylor 1915; Spurr and Zumberge 1956). Noticeable uplifting of the Nipissing beach due to isostatic rebound occurs north of Traverse Bay, Michigan (Larsen 1985b).

There is evidence for a high lake stage (Algoma) just prior to 3200 BP (Hansel *et al.* 1985); it is thought to have ended about 3000 BP (Larsen 1985a). The term "Algoma" (Leverett and Taylor 1915) is generally applied to any outstanding shoreline feature below the Nipissing in both the Lake Michigan and Lake Huron basins (Larsen 1985b), thought to correlate to this high water stand at 3200-3000 BP. In contrast to the Nipissing shoreline, the Algoma shoreline is not extensive (Spurr and Zumberge 1956) and is poorly correlated. It has been identified only at widely scattered locations (Eschman and Karrow 1985).

The modern Great Lakes followed the close of the Algoma stage, but the period should not be envisioned as a time of static water levels (Larsen 1985a). Fluctuations as much as 2 m above current levels may have occurred at 1500, 1000, and 450 BP (Hansel *et al.* 1985). Evidence for periodic lake level fluctuations suggests that, at least following

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Nipissing times, water levels in the Great Lakes basins have not been controlled solely by outlet thresholds, but also by major increases and decreases in the volume of water entering the lakes (Hansel *et al.* 1985; Larsen 1985a; Hansel and Mickelson 1988). Therefore, changes in the amounts of meltwater and streamflow within the Great Lakes watershed have probably affected lake levels. The implication of non-outlet controls on water levels in the Great Lakes is that the Nipissing and Algoma maxima, as well as the periodic fluctuations that have followed, were probably short-lived and separated from each other by lower levels (Larsen 1985a). Thus, water levels in the Lake Michigan basin probably did not decline gradually following the Nipissing high stage. Rather, water levels in the basin have both risen and fallen, and the Nipissing and Algoma shorelines were the outcome of particularly high stages that were separated from each other by periods of lower water levels.

Climate and vegetation

Climate and vegetation patterns in Michigan have not remained static throughout the Holocene (Davis 1969; Webb and Bryson 1972; Webb 1986). The soils in this study have developed during a time period that spans the entire Holocene. Because both climate and vegetation affect soil development, the Holocene climate and vegetation changes may have affected the soil patterns present on these surfaces. Whereas the duration of soil development is the focus of the present study, the effects of changes in climate and vegetation while the soils were developing cannot be ignored; such changes are a complicating factor in

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the interpretation of a chronosequence study. Therefore, I will summarize the history of Holocene climate and vegetation in this section.

Although it is not possible to directly observe past climate and vegetation patterns, proxy studies have been made on the basis of several types of indirect measurements. Information about Holocene climate and vegetation has been derived from data on tree rings, lake levels in small lakes (Winkler 1986), soils (Veatch 1937; Winkler 1986), archaeological site locations (Roberts 1979), and plant macrofossils (Webb 1981). For studies of Holocene climates in Michigan, however, fossil pollen deposits in peat formations have proved to be the most useful (Davis 1969, Webb 1981).

Palynological (pollen) reconstructions of paleoclimate and paleovegetation require several assumptions. The first, and most important, assumption is that the pattern of pollen deposition reflects, or is related to, regional vegetation patterns (Webb 1986), and that the modern analogues for the pollen plant communities can be used to generalize about the climate in which those communities existed (Davis 1967; Webb 1981). Assumptions must also be made about whether the plant community was in "equilibrium" with the climate (Webb 1986), or whether the speed of dispersal had been slow enough that the vegetation patterns did not entirely reflect climate conditions (Davis *et al.* 1986). *Local* soil and vegetation patterns will also affect the pollen distributions, and must be taken into consideration as well (Brubaker 1976). Non-climatic local disturbances, such as diseases or human activities, may affect the pollen record and must be distinguished from the overall climate-induced changes (Davis 1967).

Palynological studies have been conducted at many different scales throughout eastern North America and the Midwest. In this review, I will focus on the climate and vegetation of the northern Lower Peninsula only.

As the ice margin withdrew from the region, the cold late glacial climate gave way to the more moderated climate of the early Holocene (Webb and Bryson 1972). The latter is thought to have been cooler and moister than that of the present (Webb *et al.* 1983). Immediately following deglaciation (11,000 BP), vegetation in the Great Lakes area was dominated by white spruce (*Picea glauca* [Moench] A. Voss), probably as an open spruce woodland (Webb 1981). By 10,000 BP, however, red pine (*Pinus resinosa* Aiton) and jack pine (*Pinus banksiana* Lambert) had migrated into the area, and the dominant upland vegetation assemblage was a spruce-pine forest (Miller and Futyma 1987). Webb and Bryson (1972) estimated for the Great Lakes region that by 9500 BP the mean July temperature was about 3.3 °C lower than at present, and that moisture stress was less during this period than at any other time during the Holocene.

During the early Holocene, the climate moderated, getting gradually warmer and drier. By 8000 BP, white pine (*Pinus strobus* L.) appeared in the area. Around the same time spruce faded from the pollen record in the Great Lakes area (Webb 1981). Forests of spruce were gradually migrating northward, probably due to the warm, dry conditions (Wright 1968). Although not in full agreement, most authors assume that the mid Holocene in the Great Lakes area was both warmer and drier than at present. Quantitative estimates suggest that mean temperatures were about 1-2 °C warmer (McAndrews 1979) than today.

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with possibly a 5 cm decrease in precipitation (Webb and Bryson 1972). This mid-Holocene period is known variously as the Hypsithermal (McAndrews 1979), the xerothermic interval (Webb and Bryson 1972), or the climatic optimum. During this period, the ecotone between the southern Michigan deciduous forest community and the pine-dominated forests of northern Michigan was advancing northward.

After 6000 BP, the climate trends changed. The climate of the Great Lakes area became cooler and moister once again (Webb *et al.* 1986; Miller and Futyma 1987). After 6000 BP, white pine began to gradually decline in northern Michigan, while hemlock (*Tsuga canadensis* [L.] Carriere), maples (*Acer spp.*), and beech (*Fagus grandifolia* Ehrhart) increased in abundance, indicating the development of cooler and moister conditions (Webb *et al.* 1986). Water tables may also have begun to rise following the mid-Holocene, as evidenced by the initiation of organic deposition in the northern Michigan area (Futyma and Miller 1986). A widespread sharp drop in hemlock pollen occurred around 4800 BP, but has been attributed to factors other than climate, particularly an epidemic (Davis 1983). A climate somewhat cooler than today's appears to have prevailed from around 700 to 200 years ago (Miller and Futyma 1987).

To summarize, in the study area vegetation likely was dominated by white spruce forests immediately following deglaciation (c. 11,000 BP), which then gave way to red pine and jack pine by 10,000 BP. White pine became abundant in the area about 8000 BP and remained so throughout the mid-Holocene. Hemlock, maples, and beech gradually began to increase in abundance beginning about 6000 BP and continuing to the present.

Chronosequences

Soils and chronofunctions

Studies of soils on a series of related surfaces that differ from one another primarily with respect to duration of soil development are known as chronosequence studies (Jenny 1941, 1961; Birkeland 1984b). Chronofunctions are quantifications of soil properties plotted against time. Although all the terms in Jenny's (1941) factorial equation [s = f(cl, o, r, p, t,...)] probably cannot be solved simultaneously (Crocker 1952; Jenny 1961; Yaalon 1975), recently linear regression techniques have been used to statistically correlate soil properties with time (Bockheim 1980). Chronofunctions are of special interest because they reveal subtleties in rates of change that are not apparent from simple inspection of the data.

It is generally accepted that most soils reach a steady state at some point in their development (Birkeland 1984b), when inputs of energy and materials are balanced by outputs (Yaalon 1983). This idea is not, however, without controversy (Bockheim 1980). Theoretically, in the absence of environmental changes, a "mature" soil in the zonal sense should eventually reach a steady state (Bryan and Albritton 1943). Some authors have identified steady states for certain soil properties, such as total organic carbon (OC) and nitrogen (N) content, in soils less than 200 years old (Sondheim and Standish 1983; James 1988); others have found no evidence of steady state in over 1.6 million years of soil development (Busacca 1987; Pastor and Bockheim 1980).

Yaalon (1971) used the steady state concept to divide soil processes and properties into three sets: 1) those that change rapidly

and quickly reach a dynamic equilibrium with the environment, 2) those that change much more slowly and that appear to be in steady state only because their rate of change is too slow to be detected, and 3) those that are irreversible and self-terminating so that the balance of inputs and outputs is not maintained. These last processes do not, by definition, reach a steady state.

Properties in Yaalon's (1971) "rapidly adjusting" category, including OC and N content, have been those most frequently associated with steady states by researchers. For example, N and OC totals of soils on mudflows near Mt. Shasta, California, apparently had stabilized in less than 566 years (Dickson and Crocker 1953). In northern Indiana, N content of soils developed in dune sand approached a steady state with the establishment of "climax" vegetation (Olson 1958). Similarly, Sondheim and Standish (1983) showed that N reached a possible dynamic equilibrium in less than 200 years on moraines in British Columbia, although they believed this to be merely temporary, the culmination of a "phase" of ecosystem development.

Stevens and Walker (1970) theorized that, ultimately, soil "degradation" would follow a steady state, due to the complete weathering of stones and gravels. Few researchers, however, have reported evidence of a such a stage (Stevens and Walker 1970). Adams *et al.* (1975) did identify a degradation stage, which they termed a "terminal steady state," in a chronosequence on weathered granite in New Zealand. The oldest soil had degraded, rather than developed, with respect to OC and N content. These results are questionable, however, because their study sites were selected on the basis of profile morphology rather than independent dating of the surfaces.

Commonly, researchers have been unable to demonstrate the existence of a steady state in soils. Birkeland (1984a) expected that OC content of soils at high elevations in New Zealand had reached a steady state, but was unable to confirm this hypothesis. Even for time periods up to 1.6 million years, other researchers have failed to find evidence of a dynamic equilibrium for such properties as Fe-oxyhydroxides (McFadden and Hendricks 1985), clay accumulation (Levine and Ciolkosz 1983), clay films, total profile texture, and rubification (Busacca 1987). Further research may help to clarify the steady state concept, how it might be exhibited, and its mathematical "identification."

Utility of chronosequence study

Quaternary geological events are often difficult to date because of a lack of material suitable for age determinations (Birkeland 1984b). A technique for assessing development of soils on a particular surface could be used to establish relative dates for similar surfaces based solely on soil development (Machette 1985); chronosequence studies could provide such a technique (Bockheim 1980).

Towards this end, an aim of some chronosequence studies has been to determine the most reliable indicators of age across several climates and standardize those data using numerical indices (Harden 1982). Such indices are most useful for providing relative dates for geomorphic features (McFadden and Weldon 1987). Statistical methods have been developed, however, that employ chronosequence data from a surface of known age to derive the absolute age of a similar surface (Switzer *et al.* 1988).

Chronosequence studies can also be used to predict recovery rates of disturbed soils (Bockheim 1980). Specific examples of this are not frequent in the literature; some studies have been done on soils formed in mine spoil (e.g., Anderson 1977; Schafer *et al.* 1980). Research on soil development in very young surfaces (e.g., Sondheim and Standish 1983; James 1988) could perhaps also provide insight into development rates and processes on artificially disturbed sites as well.

Assumptions in chronosequence investigations

An investigation into the effect of time on soil development is done within the framework of Jenny's (1941) state factor model, which defines the state of a soil system as a function of five factors. Although many shortcomings of this model have been recognized (Stephens 1947; Crocker 1952; Chesworth 1973; Yaalon 1975), and Jenny himself has revised it (Jenny 1961, 1980), it remains a practical and understandable approach to studying soil development (Birkeland 1984b). In keeping with this model, chronosequence investigations rest on three major assumptions: 1) that the soil-forming factors vary independently of each other; 2) that the soil-forming factors climate, relief, organisms, and parent material have remained relatively constant since the inception of pedogenesis for the oldest soil; and 3) that space distance is a reasonable surrogate for time distance. Three further assumptions are fundamental to the sampling design of a chronosequence study: 1) that soil development is essentially uniform across an entire surface; 2) that soil development proceeds smoothly over time; and 3) that the duration of soil development on a surface is known. The validity of

these assumptions and problems with their implementation in chronosequence design are discussed below.

Factor independence. A rigorous, quantitative solution for the state factor equation would require the essential independence of the five soil-forming factors (Jenny 1941; Crocker 1952; Birkeland 1984b). Jenny's (1941) definitions of the factors were designed to ensure their independence, although their independence in reality has been called into question (Crocker 1952; Vreeken 1975; Phillips 1989). Time is considered by many to be the only unequivocally independent factor (Stephens 1947; Stevens and Walker 1970; Chesworth 1973) because, while temporal changes may occur in vegetation, climate, and topography, none of those influence time itself (Phillips 1989). Even parent material is not truly independent of time because additions of matter to the soil are time-related (Vreeken 1975).

Constancy of the soil-forming factors. A second major assumption of chronosequence design is that four factors have remained constant over the duration of the soils' development (while time is allowed to vary). Although theory requires that such an assumption be made, inadequate control of factors other than time is a problem common to almost all chronosequence investigations (Stevens and Walker 1970), especially if the time offset involved is relatively long (Yaalon 1975). Climatic change during soil development is often a concern in chronosequence design (Crocker 1952), particularly since the climate of the Quaternary has been especially variable (Davis 1969). Studies involving long time periods must deal with the effects of climatic change (Bockheim 1980).

Theoretically, constancy of a factor can be claimed both when its range of variation is narrow and also when effect on the soil property is negligible, regardless of the magnitude of variation (Jenny 1961; Birkeland 1984b). In practice, however, it is difficult to assess the relative effectiveness of variation of a factor (Birkeland 1984b); it is likely that any variation of a state factor will have some effect on soil genesis (Crocker 1952). In most chronosequence studies it is assumed that the range of climate variation is narrow or ineffective, either by ignoring the question completely (e.g., Levine and Ciolkosz 1983), by discussing past climate changes and failing to discuss the implications (e.g., Muhs 1982; Protz *et al.* 1984; Busacca 1987), or by claiming explicitly that the magnitude of climate variations was small relative to the duration of soil development (e.g., Franzmeier and Whiteside 1963a: Birkeland 1984a).

The problem of climatic change can be alleviated simply by limiting chronosequence investigation to those situations where the time scale involved is relatively short (Yaalon 1975; Birkeland 1984b). Although some studies do involve short time scales (e.g., Hallberg *et al.* 1978; Schafer *et al.* 1980; Sondheim *et al.* 1981), the properties that can be investigated within that time frame are restricted. Limiting all chronosequence studies to short time scales in order to eliminate climate change difficulties would severely limit the scope of chronosequence studies.

Vreeken (1975) felt that a change in chronosequence design could mitigate the problems discussed above. In a "pre-incisive" chronosequence, development begins at all sites simultaneously, but is arrested at different times. Such a sequence would provide a "timelapse" series of pictures of the soil at various points along its development, which would aid in discerning the effects of environmental change. Cessation of development in a pre-incisive situation usually involves burial, however, which in itself can cause modifications of soil properties (Mausbach *et al.* 1982; Schaetzl and Sorenson 1987). Post-burial changes could introduce unacceptable levels of error (Yaalon 1975). Pre-incisive sequences are also rare and have been little studied (Vreeken 1975).

Although most chronosequence studies fail to hold all four factors constant (Stevens and Walker 1970), the accumulated evidence of many studies, taken as a whole, may permit the formation of a consistent picture against which the overall results may be judged (Crocker 1952; Yaalon 1975). It may also be possible to develop methods to quantitatively assess the effects of climatic change in chronosequence studies (Bockheim 1980). Richardson and Edmonds (1987) have suggested a linear regression and correlation method for estimating the polygenetic influence in state factor equations.

Comparative geography technique. Most chronosequence studies involve the selection of soils on surfaces of different ages and in different places, in order to study the effects of time on the soils. This "comparative geography technique" requires the assumption that all the soils have passed through the same stages as the younger ones in the sequence, and have, in fact, been identical to them at some time (Savigear 1952; Stevens and Walker 1970; Welch 1970; Brunsden and Kesel 1973). Vreeken (1975) questioned the validity of this assumption; it implies that soil history repeats itself quite literally throughout time, which is unlikely. Even the formation of a new surface may in itself have permanently changed the local environmental conditions on older surfaces.

Lateral homogeneity. Within a single surface of uniform age, a given soil property will exhibit spatial variability (Beckett and Webster 1971). This variability derives both from measurement error and from heterogeneity of samples unrelated to their age (Switzer *et al.* 1988). Chronosequence studies must account for spatial variability (Sondheim and Standish 1983), but sampling design has rarely permitted this. Many researchers have simply assumed lateral soil homogeneity without verifying that assumption (Vreeken 1975), or may have introduced bias by employing sampling strategies involving a single pedon on each surface (e.g., McFadden and Hendricks 1985; Little 1986a, b; McFadden and Weldon 1987; James 1988).

In some chronosequence investigations the existence of spatial variability has been acknowledged, but sampling design has not been statistically sound. For example, Dickson and Crocker (1953) attempted to minimize within surface variability by sampling only under a particular vegetation type, yet there was no replication of samples. Similarly, others have endeavored to minimize variability by sampling only the most developed pedon (Franzmeier and Whiteside 1963a) or a modal pedon (Sondheim *et al.* 1981) from each surface with no attempt at replication.

Others have sampled two or three replicate pedons from each surface in order to minimize error. Busacca (1987) described opposite ends of a very long (6 m) backhoe trench in each surface. Harden

(1982) sampled at least one pedon from each surface, with replicates on most. Protz *et al.* (1984) dug up to four pits on each surface, taking three replicate sets of samples from the most representative pit. In these studies, however, the rationale whereby the sampled pedons were chosen, whether random or systematic, was not discussed.

Quantitative methods designed to account for soil spatial variability are rare in chronosequence investigations. Sondheim and Standish (1983) incorporated a modified stratified random sampling design in their study of six recent moraines in Canada. This procedure involved digging 10 randomly placed pits on each moraine, and sampling at specified depths in each pit. Their sampling design allowed for thorough statistical analysis of the data.

Uniform soil development over time. An underlying assumption in many chronosequence investigations, especially in the derivation of chronofunctions, is that soil development proceeds uniformly when examined over long time scales. In reality, however, soil development may be episodic or complexly related to climate and feedback mechanisms (Switzer *et al.* 1988). Intrinsic thresholds may change pathways of soil development (Muhs 1984); extrinsic thresholds of climate and time have also been reported (Birkeland 1984a; Busacca 1987). Chronosequence design may not accurately depict the "steps" in soil development resulting from thresholds, especially when results are represented by linear regression equations (Switzer *et al.* 1988).

Independent age determinations. Proper design of a chronosequence study requires accurate knowledge of the duration of soil development on each surface. This, in turn, demands that the date of surface formation can be established independently of soils data and that inception of soil development coincide with surface formation. Inadequate age control in chronosequences may result when absolute surface dates are not known, or when ages of the soils are inferred relative to only one absolute date. The latter type of chronosequence has limited value in the study of rates of pedogenesis (Bockheim 1980). Occasionally, circular dating, when the development of soils on a surface is taken as evidence for the age of the surface, is used in chronosequence design. For example, Adams *et al.* (1975) relied on soil profile morphology and degree of weathering as criteria in choosing pedons to sample for a New Zealand chronosequence. In another case, Levine and Ciolkosz (1983) dated older deposits by extrapolating from a chronofunction developed on younger soils, then utilized those dates in constructing a second chronosequence.

Although an established surface age may be taken as a maximum limit on duration of soil development, the minimum duration of soil development is not easily determined. Generally the inception of pedogenesis has been assumed to coincide with surface formation, but unrecognized periods of erosion or deposition on that surface may invalidate that assumption.

Types of chronosequence investigations

Time has two aspects that relate to soil development: (1) time-length or duration of development, and (2) time-context, the historical time period during which development has occurred (Vreeken 1975). Both aspects are important to the structure of chronosequence investigations.

Vreeken (1975) described three designs for chronosequence studies: 1) pre-incisive, 2) post-incisive, and 3) time-transgressive. In a pre-incisive arrangement, development begins at all sites simultaneously, but is arrested at different times, usually by deep burial. If development begins at different times on each surface and the soils have never been buried, a post-incisive arrangement results. By far the majority of published studies have been post-incisive.

Pre-incisive chronosequence studies. Most pre-incisive studies have involved just two soils, one buried by human activity and a similar soil still developing on the unburied surface nearby. A "sequence" of only two soils does not provide a sufficient base from which to derive chronofunctions or to trace development over time.

In Russia, comparison of soils buried under Bronze Age mounds called "kurgans" with surrounding surface soils provided a pre-incisive two-soil sequence (Madanov and Tyumenko 1968; Madanov *et al.* 1968); a similar study in Iowa compared soil buried under railroad spoil for 100 years with nearby undisturbed soils (Hallberg *et al.* 1978). In the Russian study, the buried soil exhibited lower humus and salt content, and higher levels of carbonates and bases in comparison to the unburied soils. Similarly, in Iowa the buried soils had lower surface OC and an increase in clay and fine silt in the A horizon. Many of these differences could be best explained through post-burial changes in the buried soil, rather than to continued development of the modern (surface) soil. As such, these studies have more bearing on soil degradation (due to burial) than to soil development.

<u>Post-incisive chronosequence studies.</u> Post-incisive studies involve a wide variety of climates and soil types, with time spans ranging from less than 200 years (e.g., James 1988) to as much as 3.0 million years (Pastor and Bockheim 1980). Limitations of the post-incisive design (discussed earlier) may be minimized in studies covering very short time ranges by carefully holding all other factors nearly constant (e.g., Sondheim *et al.* 1981). More often, design flaws (e.g., climatic change) have been simply acknowledged (e.g., McFadden and Weldon 1987).

<u>Disturbed soils.</u> Native American burial mounds and mine spoil sites have proven to be fruitful locations for chronosequence study if the date of disturbance has been historically recorded or established through archeological study. Poor mixing of the material used to construct the mounds could be a disadvantage in some studies of this type (Parsons *et al.* 1962). If the intent is to compare the spoil-developed soils with the surrounding natural soils (Parsons *et al.* 1962; Hallberg *et al.* 1978; and Schafer *et al.* 1980), the parent material in the mound soils would probably not have been identical to the parent material of the surrounding soils due to the weathering of the material prior to mounding. If only soils on mounds of different ages are considered in the study, this effect can be mitigated (Bettis 1988).

Studies of disturbed soils have documented strikingly rapid OC accumulations in new deposits. Schafer *et al.* (1980) noted that thin A horizons formed even on raw mine spoils (i.e., no attempt was made to hasten "reclamation" of the spoils) in a few years under good vegetative cover. In railroad spoil materials in Iowa, surface horizons meeting most of the criteria for mollic epipedons (Soil Survey Staff 1975) had developed within 100 years, although the OC content of these horizons was not as high as that of the surrounding soils (Hallberg *et al.* 1978). Development of subsurface horizons, however, apparently occurs more slowly. No argillic horizon development was found in 1600 year old Iowa burial mounds (Bettis 1988), and a 50 year old Minnesota minespoil exhibited a nearly uniform distribution of carbonates with depth (Schafer *et al.* 1980).

Soils on moraines. Sequences of natural soils spanning less than 200 years, such as moraines of active glaciers in Alaska and British Columbia, are valuable for studying changes in those soil properties that are more ephemeral or reach "steady state" levels in hundreds of years (Yaalon 1971). Crocker and Dickson (1957) observed rapid initial increases in total N and OC contents in Alaskan soils, and rapid decreases in pH levels. Respective rates of increase or decrease of these properties had slowed within 200 years. James (1988) noted negligible OC accumulation in the first 15 years of pedogenesis on Alaskan moraines, followed by a period of accelerating accumulation. In British Columbia, Sondheim and Standish (1983) observed an initial rapid increase in total N, followed by a leveling off at about 200 years after deglaciation. They speculated that this plateau might represent the beginning of a steady state for N, although maybe only temporarily. All the above authors associated changes in N and OC accumulation rates to patterns of vegetational succession.

In contrast, a time span of 3.0 million years and a cold and arid climate contributed to the development of strikingly different soils on moraines in Antarctica (Pastor and Bockheim 1980). Maxima in free Fe, clay, and silt contents, as well as morphological properties, were reached within 1.6-2.1 million years. No increase in soluble salts with age was observed, which might be explained by climatic change or spatial variations in precipitation.

Soils on alluvial and marine terraces. Alluvial and marine terraces provide a sequence of surfaces appropriate for studying soil development. Busacca's (1987) study of alluvial terrace soils in California, ranging in age from 600 to 1.6 million years, revealed that individual soil properties develop at different rates. Properties of the A horizon reached steady state levels within about 1000 to 40,000 years, and soil structure by 250,000 years, but clay films, total texture, and rubification still appeared to be changing even after 1.6 million years.

In alluvial fan deposits and fluvial terraces in California, Fe-oxyhydroxide content changed systematically with time (McFadden and Hendricks 1985). Steady state for the Fe-oxyhydroxides was not reached in these soils even after 500,000 years. At another California location, however, profile totals of Fe-oxides appeared to reach a maximum and then decline within one million years (Muhs 1982). These soils, developed on marine terraces, showed development proceeding linearly with respect to time for solum thickness, total profile clay and salts, and profile smectite/mica ratios.

In fluvial terrace soils of California up to 500,000 years old, a time-related decrease in A horizon thickness was attributed to upward thickening of the B horizon and increases in organic matter oxidation rates at the expense of the A horizon (McFadden and Weldon 1987).

Eolian dust caused an increase in silt content in Pleistocene-age soils, which impacted soil permeability and water balance. A sudden increase in clay content after 8,300 years was attributed to an extrinsic pedologic threshold (McFadden and Weldon 1987).

Ahmad *et al.* (1977) analyzed a series of three alluvial deposit soils, ranging in age from less than 100 years to more than 10,000 years, in the Punjab river plains of Pakistan. They observed a positive correlation between solum thickness, clay content, and increased age of the soil. Organic matter content of all three soils was low and showed no definite trend with age, perhaps due to the torric soil moisture regime (Soil Survey Staff 1975).

In Finland north of the Arctic Circle, podzolization was studied on a sequence of terraces of fluvial and glaciofluvial origin ranging in age from 300 - 9500 BP (Koutaniemi *et al.* 1988). Parent materials were acid sands; in the surface horizons, median particle size was medium sand. Translocation of Fe and Al resulted in an accumulation of sesquioxides in the B horizon. There was, however, an unanticipated decrease in Fe content with age in the B horizon. Al content increased with age in both the B horizon and the C horizon.

Soils on sand dunes. Soil development has been studied where eolian sand deposits have formed successions of dunes. In northwestern Indiana, a sequence of dunes ranging in age from less than 10 to more than 12,000 years revealed rapid early leaching of carbonates (Olson 1958). The leached zone had reached two meters in less than 1000 years. Steady-state levels of nitrogen were approached with the establishment of mature forest after 8000 years. Increasing silt content in upper layers of older soils had apparently resulted from additions of loess from newly-deposited nearby moraines. Miles and Franzmeier (1981), studying soils developed on dunes in the same area, felt that both lithology and time had influenced the development of these soils, since parent material mineralogy differed as well as duration of development. Older soils had continuous argillic horizons, younger soils had textural bands, and incipient soils lacked evidence of illuviation.

Podzol soils developing in dune sands have been studied in Australia by several researchers (Burges and Drover 1952; Thompson 1981; Little 1986a, b). Burges and Drover (1952), whose transect through dune deposits included surfaces ranging in age from newly formed to 4000 years, observed leaching of carbonates, a decrease in pH, and downward translocation of Fe. Organic matter in the surface 10 cm appeared to reach steady state levels within 1500 years. The depth to the B horizon increased with time, reaching 40 cm in 4000 years. Thompson (1981), whose sequence included sandy soils ranging in age from 3800 years to more than 45,000 years, also noted an increase in depth to the B horizon, together with an exaggerated thickening of the E horizon (up to 15 m on the oldest sites). He felt that the top of the Bhs horizon was being progressively remobilized and deposited deeper in the profile, because neither its thickness nor development increased proportionately to E horizon thickness.

Similarly, Little (1986a) found podzols with depths to the B horizon exceeding 8 to 10 m on older surfaces. Accumulation of Fe and Al in the B was not sufficient to account for A and E horizon losses of these minerals, possibly suggesting that Fe and Al had been remobilized and removed from the profile. These "giant podzols" probably were no longer in equilibrium with their surface horizons, and might have been related to past climates (Little 1986b). A strong relationship between depth to the B horizon and age existed in these soils (Little 1986b).

Soils on beach sands. Beach sand deposits, eroded into beach ridges and subaerially exposed as water level subsided, can provide a sequence of soils appropriate for a chronosequence investigation. Protz et al. (1984) studied six sequential storm beach ridges along a 70 km transect near Hudson Bay, Ontario. Ages of the surfaces ranged from less than 100 to 5400+ years. Although the climate had undoubtedly changed periodically during this time, the effects of these changes could not be quantified. In these soils, total organic matter content increased rapidly, doubling every 1000 years up to 3000 years, at which time it peaked. Depth of carbonate leaching exhibited a linear relationship with time.

Sondheim et al. (1981) and Singleton and Lavkulich (1987) studied soils developed along a prograding beach deposit on Vancouver Island, British Columbia. The soils, dated at 127 - 550 years, were sampled at intervals along a transect perpendicular to the shoreline. Spacing varied slightly so that pedons considered to be "modal" could be sampled. With increasing age, Fe, Al, and organic matter contents of the soil increased exponentially, and organic matter accumulations were found at lower depths within the profile. Ca, Mg, and K contents of the profiles showed no definite relationship with time. It appeared that in this environment a Humo-Ferric Podzol (Canada Soil Survey Committee 1978) could develop within 350 years (Singleton and Lavkulich 1987).

In northwest Finland, a sequence of 16 sandy beach ridges ranging in age from 130 - 3010 BP showed visible differentiation of podzol horizons within 400 - 500 years, and chemical differentiation of horizons within 200 - 300 years (Jauhiainen 1973). The distribution of Fe, Al, and silicic acid within the profiles showed a clear minimum in the E horizon and a maximum in the B or B/C horizon in all but the youngest profile. The amount of sesquioxide and silicic acid accumulation increased with age of the profile. The ratios of A to B horizon content of sesquioxides and silicic acid increased with profile age ($r = 0.56$; $p < 0.5$). The variation in these ratios was greater in old than in young podzols.

A chronosequence on four sandy marine terraces spanning 300 - 13,000 BP in northern Denmark showed increasing silt and clay content of the surface horizons with surface age, due either to weathering or eolian additions (Nørnberg 1977). Total C content of the soil surface appeared to reach an equilibrium by 4000 years. Although an accumulation of Fe and C in the B horizon was apparent in the older pedons, it appeared that some clay migration had also occurred in pedons on 10,000 and 13,000 year old surfaces.

Franzmeier and Whiteside (1963a, b) and Franzmeier *et al.* (1963) investigated soil development on a sequence of beach ridges ranging in age from 2250 to 10,000 years in Cheboygan and Emmet Counties, Michigan. They discovered no relationship between pH and age of the soils. OC maxima in the B horizon (Bhs horizon) had been established by 8000 years. Fe and Al showed subsurface maxima, even in the youngest soil, that increased in intensity through time. Al accumulation peaked slightly deeper in the profile than did Fe accumulation. A peak in total profile P was observed at around 3000 years, and in total profile Fe and Al at 8000 years; however, profile contents of organic matter, silt, and clay were not observed to peak. Therefore, Franzmeier *et al.* (1963) concluded that podzolization processes reached maxima at different times. Several problems in the experimental design, however, may limit the validity of the results. Although the youngest three surfaces were indeed beach ridges and were located within one km of each other, the fourth surface was a sandy moraine, located 45 km inland. Particle size distributions of the moraine parent materials were distinct from those of the beach ridges. Although a definite variability was observed among the soils on a surface, especially at the moraine sites (Franzmeier and Whiteside 1963a), only one pedon per surface was sampled, that with the strongest development. This method could introduce experimental bias. No attempt was made to formulate chronofunctions or to quantitatively analyze the data; results were presented graphically.

Chronofunctions and other statistical treatments

Basic chronofunction models

Jenny (1961) defined a chronofunction as the mathematical solution of the relationship between a soil property and time, with other factors held constant:

$$s = f(t_{cl,o,r,p,...})$$

Mathematical solution of the chronofunction equation for a large number of cases, along with similar treatment of climo-, bio-, topo-, and lithofunctions, is widely believed to be the best approach for solving the general state factor equation:

$$s = f(cl, o, r, p, t,...)$$

(Yaalon 1975; Bockheim 1980; Birkeland 1984b; Richardson and Edmonds 1987; Phillips 1990). One approach, and that most widely attempted, to mathematically solving the chronofunction equation, is the statistical technique of linear regression (Bockheim 1980, Richardson and Edmonds 1987; Phillips 1990). Another possible approach is the stochastic modeling method suggested by Yaalon (1975) and employed in a geological context by Lafon and Vacher (1975). This second approach is more mathematically complex, however, and seldom attempted (e.g., Sondheim et al. 1981).

Using linear regression techniques, Bockheim (1980) calculated chronofunctions for 32 published chronosequence studies. Three separate regression models were employed in the analysis:

$$\text{Linear:} \quad Y = a + bX$$

$$\text{Single logarithmic:} \quad Y = a + (b \log X)$$

$$\text{Logarithmic:} \quad \log Y = a + (b \log X)$$

where Y = soil property and X = time. Many of the resulting functions were statistically significant, although no indication of the r^2 values was provided. The single logarithmic model resulted in the largest number of statistically significant equations. Bockheim (1980) also tried exponential and polynomial regression models, but rejected them because they did not fit the data.

After the publication of Bockheim's (1980) paper, most chronosequence studies have included chronofunctions derived using linear regression techniques (e.g., Muhs 1982; Harden and Taylor 1983;

Levine and Ciolkosz 1983; Birkeland 1984a; McFadden and Hendricks 1985; Mellor 1986; Busacca 1987; James 1988). As Bockheim (1980) reported, linear and log-linear functions often adequately describe chronosequence data. For example, the linear model has been used to describe increasing leaching depth on storm beach ridges (Protz *et al.* (1984), solum thickness, clay content, and salt content on California marine terraces (Muhs 1982), and the micromorphological development of moraine soils in Norway (Mellor 1986). Other researchers have found the log of time or log-log models to be more appropriate. All eight profile properties investigated by Busacca (1987) from soils on alluvial terraces in California increased linearly with the log of soil age. The log-log model gave high r^2 values for Fe-oxyhydroxide increase California soils (McFadden and Hendricks 1985) and clay accumulation increase in till soils in Pennsylvania (Levine and Ciolkosz 1983).

Although Bockheim (1980) discarded the exponential model because of lack of fit, Birkeland (1984a) found that it best explained most index values for alpine soils in New Zealand. Other index values were better explained with the log of time or linear models.

A final model, not considered by Bockheim, but quite interesting, is the logistic model. The graphs of this model are sigmoidal in form, showing slow increases at first, followed by a period of rapid accumulation, gradually tapering off as the value approaches a final limit. Birkeland (1984b) and Yaalon (1975) postulated that such a shape would be appropriate for a number of soil properties, and it has been applied to beach soils on Vancouver Island, British Columbia (Sondheim *et al.* 1981) and recent moraine soils in British Columbia (Sondheim and Standish 1983). Yaalon (1983) felt that the equations derived by

Bockheim (1980), which were linear and therefore admitted no trend toward a steady state, probably represented the straight middle portion of the sigmoidal graph, the period of rapid accumulation. Mathematically, however, the logistic model is more complex; thus it has not often been employed. James (1988) suggested that such a model would be appropriate in describing the rate of OC accumulation in recent moraine soils in Alaska, but he did not actually attempt it.

Statistical treatments

Principal components analysis. Statistical analyses of chronosequence data, other than the derivation of chronofunctions using linear regression techniques, have seldom been attempted, although in three studies principal components analysis (PCA) has yielded interesting results. Chemical data from soils on a prograding beach on Vancouver Island, British Columbia, were subjected to a PCA (Sondheim *et al.* 1981). Two significant components with eigenvalues greater than one were derived. The loadings on the first component were related to "dominant pedogenic processes," while the loadings from the second related to the distribution of the basic cations, possibly influenced by sea spray sweeping onto the island, or by biocycling.

Little (1986b) found that beach and dune sand soils in Australia yielded two significant components in a PCA. These components were similar to those of Sondheim *et al.* (1981): the first, termed "katamorphic," was related to K, Al, and Fe, and was considered a measure of podzolization, but the second, "biotic," factor related most closely to Ca, Mg, and Na in the soils.

In a study of moraine soils in British Columbia, oblique factor analysis yielded two significant factors, with the correlation between them .20. Loadings from the first factor related to C, N and pH, and loadings from the second related to Fe, Al, and sand and clay particle sizes (Sondheim and Standish 1983). The authors did not provide a simple interpretation of the two factors.

Soil variability change over time. Little work has been attempted that could help explain changes in soil variability with time. Busacca (1987), however, found that the soils on alluvial terraces in California were less variable with increasing age of the surface, judged by the coefficient of variation. Since a different number of samples was taken on each surface, and because sample size was so small (2-4 per surface), he discounted this finding somewhat. In contrast, Franzmeier and Whiteside (1963a), in their study of beach ridge soils in Michigan, observed little variability on younger surfaces and considerable variability on older surfaces. The observation, nonetheless, was not quantitatively tested.

Indices of soil development

Numerical indices of soil development permit quantification of morphological soil properties, so that the researcher can easily compare development of pedons by examining one number (or range of numbers) per pedon. Additionally, an index aids in interpretation of morphological data so that non-field scientists can more easily understand them (Bilzi and Ciolkosz 1977). Indices developed for use on soils of one climatic region are not necessarily applicable to soils of another region: testing is necessary to establish the validity of their application (Bilzi and Ciolkosz 1977; Harden 1982).

Many indices provide a method of evaluating soil development based on soil color. Buntley and Westin's (1965) index was developed for prairie soils as a method of relating Munsell hue and chroma measurements to the development of a pedon. Hurst's (1977) redness index relates the redness of the soil to its Fe-oxide content. McFadden and Hendricks (1985) found very good correlation between the Hurst index and the Fe-oxide content of alluvial terrace soils in southern California, especially in older Pleistocene soils. Another index based on color data (Bilzi and Ciolkosz 1977) was developed for use on till soils in Pennsylvania. This index assesses points for changes in Munsell hue, value, and chroma measurements, either between horizons of between pedons.

The Harden (1982) Index has been widely applied. A modification of the Bilzi-Ciolkosz index, it was developed for use on soils of California alluvial terraces. The index has been employed successfully in at least three published studies of soils developed on California alluvial terraces (Harden 1982; Busacca 1987; McFadden and Weldon 1987), as well as in New Zealand alpine soils (Birkeland 1984a). Although designed to apply to a wide range of soils (Harden and Taylor 1983), the index relies on field evaluations of properties such as clay films, texture, structure, and consistence. These properties are important for assessment of the development of many soils; other soils, however, have developed along other pedogenetic pathways such that these are not important diagnostic properties. Application of the Harden Index to such soils is difficult (Schaetzl and Mokma 1988).

Some indices have been targeted primarily for use with specific types of soils. Development of argillic horizons can be measured with indices devised by Rockwell et al. (1985) or Levine and Ciolkosz (1983). The podzolization index of Duchaufor and Souchier (1978) uses chemical data to measure the degree of podzolization in a soil. A morphology-based index of development intended only for use with Spodosol or Spodosol-like soils, the POD Index (Schaetzl and Mokma 1988), utilizes the color differences between the E and B subhorizons of the pedon and the number of B subhorizons to evaluate spodic development. An advantage of this index is that no chemical data are required; soil development is assessed using field descriptions only.

The present thesis attempts to assess the variability of soils on four surfaces in a chronosequence. Because the POD Index requires only field data, it is possible to examine a relatively large number of pedons per surface. Few previous studies have examined more than one or two pedons per surface, and none of these few have investigated soils developed in sandy parent materials in a Spodosol region. This thesis, therefore, builds on a previous chronosequence study of sandy soils in the northern Lower Peninsula (Franzmeier and Whiteside 1963a, b) by examining the spatial variability of the soils and by developing quantitative chronofunctions for a variety of soil properties and development indices.

STUDY AREA

The study area is located in Emmet County, in the northwestern portion of Michigan's southern peninsula, near the shore of Lake Michigan (Fig. 1). The Algoma study site is within Wilderness State Park; the Nipissing, Battlefield, and Main Algonquin sites are in Cross Village township, about 10 km to the south. The study area was chosen because it appeared to be the only location in the northern Lower Peninsula where extensive sites with well drained, sandy soils could be found on all four surfaces.

Surface ages

The soils in the study area are developed on sandy sediments that were deposited in near-shore lake environments. Four lake terraces are prominent features within the study area and can be intermittently traced along much of the Lake Michigan shoreline, especially in the northern part of the southern peninsula (Leverett and Taylor 1915). Estimates of lake elevations at Cross Village based on shoreline measurements and surface ages were adopted in this study as shown in Table 1 and Figure 2. I have assumed that pedogenesis began on each surface immediately following the close of the associated lake stage and the subaerial exposure of the sediments.

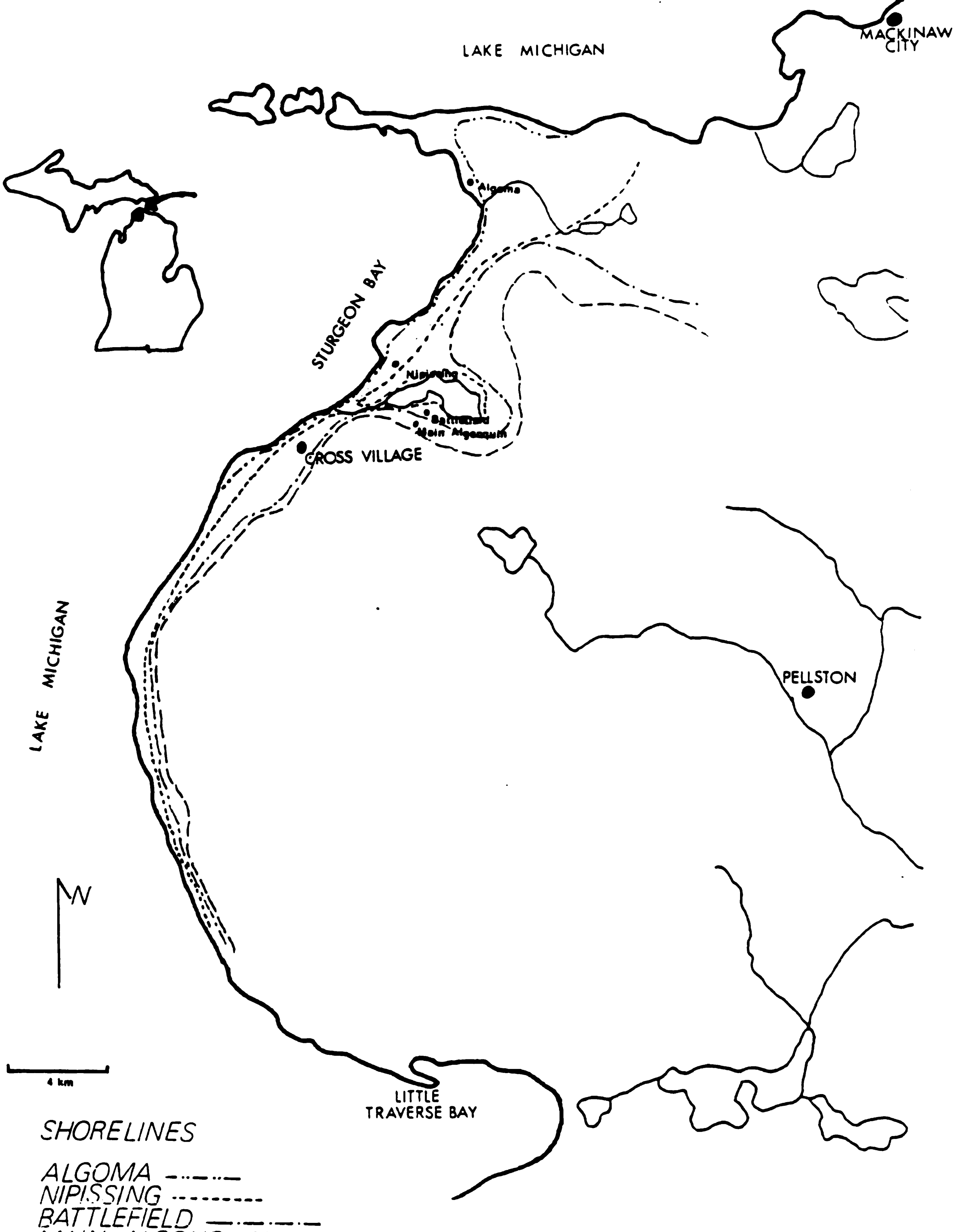

Figure 1. Location of study sites, Emmet County, Michigan, USA.

Table 1. Surface ages and shoreline elevations at Cross Village, Emmet County.

Surface	Years BP Range	Years BP Adopted in present study	Shoreline elevation	References[a]
Algoma	4000-2500	3000	183 m	2,3,4,5,6,7
Nipissing	6300-3800	4000	189 m	2,3,4,5,6,7
Battlefield	10,300-10,000	10,000	210 m	1,2,6,7
Main Algonquin	11,200-10,500	11,000	227 m	1,2,6,7

a References: (1) Futyma 1981; (2) Hansel *et al.* 1985; (3) Hansel and Mickelson 1988; (4) Larsen 1985a; (5) Larsen 1985b; (6) Larsen 1987; (7) Leverett and Taylor 1915

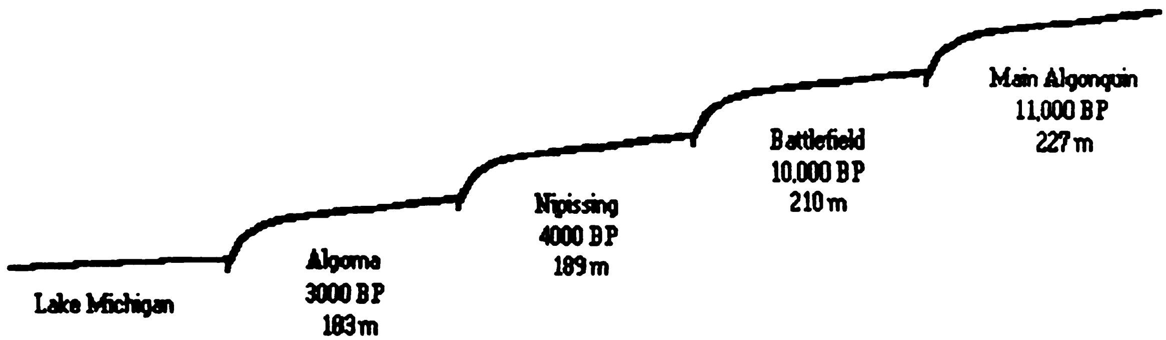

Figure 2. Schematic diagram showing relationships of terrace surfaces, assumed surface ages, and shoreline elevations (not drawn to scale).

Topography and parent material

Slopes within the study area are level to gently sloping (<6%), except where the surface has been covered 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 drained or excessively drained, although extensive areas on the Algoma surface are somewhat poorly and poorly drained.

The parent materials in the study area are primarily lacustrine sands. Although coarse fragments (> 2.0 mm) are not prominent components of most soils in the study area, the percentage of coarse fragments varies considerably both between and within pedons. The Algoma site is underlain by a deposit of extremely gravelly calcareous sands and boulders, which may occur within 150 cm of the surface.

<u>Vegetation and climate</u>

While the soils on the oldest surface have been developing, regional climate patterns have averaged both cooler and warmer than the climate of today, probably affecting vegetation patterns as well.[1] The first post-glacial vegetation in the area was probably dominated by spruce and pine, which was succeeded in turn by a pine-dominated assemblage (Futyma and Miller 1986). Subsequent assemblages with lower pine percentages and more sugar maple (*Acer saccharum* Marshall), birch (*Betula papyrifera* Marshall), beech (*Fagus grandifolia* Ehrhart) and hemlock (*Tsuga canadensis* [L.] Carriere) are characteristic of the region (Miller and Futyma 1987), and are associated with a warmer climate; temperatures peaked around 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 air temperature (1951-1980) at Pellston is 5.2 ^{o}C, with July temperatures averaging 18.6 ^{o}C and January temperatures averaging -9.1 ^{o}C. Mean annual precipitation is 83 cm. with the highest monthly average occurring in September (10 cm), and the lowest monthly average in February (3.9 cm). A mean of 298 cm of snow falls each year between October and May. Snowcover 2.5 cm (1 in) or more deep is on the ground for an average of 130 days per season, but this varies greatly from season to season. Because the study sites

[1] For a summary of climate and vegetation patterns during the period the soils were developing, see the discussion beginning on p. 10.

are closer to the lake than is Pellston, climate at the study sites is more modified by the presence of the lake, with somewhat lower precipitation amounts and more moderated temperatures. Precipitation data for Cross Village, located near the study sites, show an annual precipitation of 73 cm and an annual average snowfall of 200 cm, both lower than the averages for Pellston. Temperature data are not available for Cross Village.

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* L.), red maple (*Acer rubrum* L.), red pine (*Pinus resinosa* Aiton), black oak (*Quercus velutina* Lamarck), red oak (*Quercus rubra* L.), American beech, sugar maple, jack pine (*Pinus banksiana* Lambert), and serviceberry (*Amelanchier spp.*). Red and jack pine are most abundant on the Nipissing surface; American beech is found only on the Main Algonquin surface, which has the greatest spodic development (cf. Mokma and Vance 1989). On the wetter Algoma surface, the most abundant trees, in order, are balsam fir (*Abies balsamea* [L.] Miller), paper birch, aspen, northern white cedar (*Thuja occidentalis* L.), striped maple (*Acer pensylvanicum* L.), red maple, white spruce (*Picea glauca* [Moench] A. Voss), and ash (*Fraxinus spp.*). These trees tolerate a wide variety of drainage conditions (Barnes and Wagner 1981) and may be indicative of the extent of poor drainage in the area, although poorly drained sites were avoided in sampling.

Bracken fern (*Pteridium acquilinium*) is the most common groundcover plant on the three oldest surfaces, and is least common on the Nipissing surface. Other understory plants on those surfaces include wintergreen (*Gaultheria* spp.), blueberry (*Vaccinium* spp.), lady slipper (*Cypripedium* spp.), and bramble (*Rubus* spp.). Groundcover vegetation on the Algoma surface is sparse, consisting primarily of conifer seedlings, small ferns, perennial herbs, and grasses.

METHODS

Field Methods

Preliminary survey and site identification

Prior to field study, potential sites were identified on USGS 7.5 minute topographic quadrangles and modern SCS county soil survey maps. Site prerequisites included slopes that were less than 5%, and unclutivated soils that were well drained and developed on lacustrine sands. Upper elevation limits for each surface at locations in Emmet, Charlevoix, Cheboygan, and Presque Isle counties were taken from published reports (Futyma 1981, Leverett and Taylor 1915, Larsen 1987). Ground elevations were interpolated from contours on the USGS quadrangles.

The study sites were located on successive surfaces in sufficiently close proximity so as to form a sequence with the greatest possible uniformity of parent material, vegetation, and climate. The most satisfactory sequence was found in Emmet County, and several potential sites were identified for field evaluation.

Spatial variability analysis

At least two sampling areas were chosen for a study of soil spatial variability on each surface except the Algoma, where only one area was identified because of the small extent of well drained soils on that surface. Some areas were undesirable for the study because of poorer than expected natural drainage or parent material textures that deviated significantly from those in other areas. Such sites were not sampled.

In each area, sampling points were located in blocks composed of a 3 x 3 grid, with points spaced about 50 m apart. On the Nipissing and Algoma surfaces, the location of roads or the boundaries of the surface forced some deviation from the 3 x 3 grid pattern, but the 50 m spacing between sites was maintained (see Appendix 2, p. 146). If a site fell too close to a tree or was located on a treethrow pit or mound, it was avoided in favor of the nearest more representative site. Two blocks of 9 sites each were sampled on each surface, for a total of 18 sites per surface.[1] Each site was marked at the time of sampling for ease of later re-identification. The elevation of each site was estimated from topographic maps. Vegetative cover was recorded.

Little precedent exists in the literature for a sampling scheme adequate for estimating soil variability without the use of geostatistics. Therefore, the 50 m spacing between points of the present sampling design was intended to cover relatively large areas on a surface. The inclusion of two separate blocks of points per surface was intended to maximize potential variability. Finally, at least 10 - 20 points were desired per surface in order to have enough samples for meaningful simple descriptive statistical tests.

A bucket auger was used to sample the soil to the C horizon or to a depth of about 150 cm, whichever was shallower. Moist color, texture,

[1] A total of only 10 sites was sampled on the Algoma surface due to the problem of identifying well drained soils on the surface. On the Battlefield surface, an additional partial block of 6 sites was also sampled, for a total of 24 sites on that surface.

and thickness of horizons were recorded, as was the presence of pebbles or ortstein in each horizon.

Selection of modal pedons

POD Index values were calculated for each augered site according to the method of Schaetzl and Mokma (1987). The median POD Index value was then determined for each surface. The "modal" pedon for each surface was chosen at random from among the pedons that had the median index value and which did not contain atypically high amounts of coarse fragments.

After a modal pedon had been selected, the original auger hole was located and a pit was dug around the auger hole, or within 1 m if some obstruction was present (e.g., low tree branches). The pits were approximately 1.5 m x 0.7 m, and deep enough to expose the C horizon. Standard field descriptions were made (Soil Survey Staff 1981), and bulk samples (about 1 kg) were taken by genetic horizon for laboratory analysis.

On the Algoma surface, all sites had a POD Index value of zero, but since many sites on that surface showed some signs of poor drainage or atypical parent materials, the selected modal pedon was the one with the fewest constraining characteristics. When this pedon was described and sampled (Algoma-1), it was found to contain a buried A horizon, suggesting that this soil had actually developed in eolian, rather than lacustrine, sand. Therefore, another site approximately 150 m away was identified and sampled, after further examination of the surface. This latter pedon is referred to as Algoma-2, and also had a POD Index of 0.

Laboratory Methods

Horizon samples from the modal pedons were air dried and passed through a 2 mm sieve to remove coarse fragments. Sand content was determined by sieving. Silt and clay content was determined by the pipette method (Singer and Janitzky 1986). Reaction was measured in a 2:1 soil-water mixture. Organic C content was determined by a modified Walkley-Black method (Singer and Janitzky 1986).

Extractions for analysis of Fe and Al content were performed by shaking with Na citrate-bicarbonate-dithionite (CBD), Na pyrophosphate (McKeague 1978), and with oxalic acid-ammonium oxalate (Anonymous 1990). Fe and Al concentrations of the extracts were measured by ICP spectrometry. Optical density of the oxalate extract (ODOE) was measured at 430nm following the method of Daly (1982) and Anonymous (1990).

Indices of Development

POD Index values for auger sites and for sampled pits were calculated following the method of Schaetzl and Mokma (1987). Other indices calculated for sampled pit sites include the Index of Profile Anisotropy (IPA) (Walker and Green 1976), the modified Index of Profile Anisotropy (mIPA) (Birkeland 1984a), and the Podzolization Index (K_{Al}) of Duchaufour and Souchier (1978). For these indices, the methods of calculation described in the original articles were followed.

The Harden Soil Development Index (Harden 1982) was also calculated for the sampled pit sites. This index contains two parts: separate indices for each of eight soil properties, and a composite index for the profile based on the eight properties. For the present study, only five of the eight soil properties (consistence, melanization, pH, rubification, and structure) were applicable to the soils under study here, and thus only these five were included in the calculation of the index.

Chronofunctions

Chronofunctions were developed both for data derived from auger hole sites and for pit-sampled sites. Surface ages used for chronofunction calculations were as reported in Table 1. Chronofunctions were devised by performing standard linear regression analysis of the soil property in question against surface age. Three chronofunction models were followed for data derived from the pit-sampled sites: 1) linear (untransformed data); 2) single logarithmic (soil property data log-transformed); and 3) logarithmic (both surface age and soil property data log-transformed) (Bockheim 1980).

RESULTS AND DISCUSSION

Variability of soil development[1]

Parent material uniformity

Three indicators that can provide insight into the uniformity of the parent materials in which these soils developed, both within a single surface and between surfaces, have been utilized: (1) the texture of the C horizon, as determined in the field, (2) the abundance of coarse fragments, and (3) C horizon moist color.[2] Similarity of these measures between surfaces may indicate that differences in soil development are not due to parent material effects.

C horizon texture. The C horizon of each pedon examined was determined to have a sand texture. Medium sands dominated; occasional variations in the proportions of finer to coarser sands were noted, but could not be quantified. In general, parent material textures can be considered uniform both within and between surfaces. Thus, since no lithologic discontinuities were observed above the C (or C1) horizon, profile textures at time zero can also be assumed to have been uniform.

[1] Data reported in this section include only field-acquired auger hole data.

[2] All soils were well drained or better, so that the effect of mottling on soil colors can be negated.

Coarse fragments. The abundance of coarse fragments varied considerably both within and between surfaces (Table 2). Coarse fragments were present in at least one subhorizon of at least one pedon on all surfaces, although usually in fewer than half the subhorizons of any given pedon. On the Algoma surface, most coarse fragments were noted in connection with a 2C horizon containing large gravel- and cobble-sized fragments, which 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, coarse fragments were abundant enough to obstruct augering at depth (60 - 90 cm). In general, coarse fragments on the Nipissing surface were more often encountered below the A and E horizons, but could not be attributed to a lithologic discontinuity, as on the Algoma surface. Coarse fragment distribution on the Battlefield and Main Algonquin surfaces was irregular and seemingly without pattern.

Table 2. Frequency of pedons on four surfaces in which coarse fragments were encountered in any subhorizon or in more than half of subhorizons.[a]

Surface	Block 1		Block 2		Block 3		Total	
	>0%[b]	>50%[c]	>0%	>50%	>0%	>50%	>0%	>50%
	-----------------		%		-----------------		----%----	
Algoma	30	0	nd[d]	nd	nd	nd	30	0
Nipissing	100	67	100	44	nd	nd	100	56
Battlefield	0	0	78	11	83	17	50	8
Main Algonquin	100	11	56	22	nd	nd	78	16

a: N = 9 for all blocks except Algoma Block 1 (N = 10) and Battlefield Block 3 (N = 6).
b: % pedons in which coarse fragments present in any subhorizon
c: % pedons in which coarse fragments present in >50% of subhorizons
d: no data.

The effect that differences in coarse fragment distributions might have on soil development is unclear, although surface rocks have been shown to affect heat and water gradients in the soil (Jury and Bellantuoni 1976). The coarse fragments present in these soils, however, were small and not found at the surface, so this possibility is diminished. The presence of coarse fragments may affect the movement of water through the solum, as well as reduce the particle surface area available for chemical processes, but the extent to which the amount of coarse fragments found in these soils could influence pedogenesis is unclear.

C horizon color. The moist color of the C horizon may be used as an indicator of parent material uniformity in these well drained soils. It is assumed that the C horizon represents relatively unweathered parent material, and that C horizon color can indirectly provide a general indication of its mineralogical composition. Thus, on related surfaces in a small geographical area, C horizons with similar colors may represent comparable parent materials.

The color of the C horizon material for all pedons on the four surfaces is quite similar, both within and between surfaces (Table 3). The Munsell hue for the C horizon is, in all cases, 10YR, and Munsell values of 5 or 6 are most common (Figure 3). On the Algoma surface, the mean chroma of the C horizon colors is somewhat lower than that of the other surfaces, but the predominant hues and values are similar to those of the other surfaces. The C horizon values are lower on the Nipissing surface, on the average, than on the other surfaces, due to the presence of a number of pedons with a C horizon color of 10YR 4/4. Variability in C horizon color appears to be no greater between surfaces than within surfaces.[3]

[3] Analysis of variance testing of color data was not performed due to the non-linearity of the Munsell color system.

Table 3. Mean C horizon Munsell value and chroma of pedons on the four surfaces.[a]

Surface	Block 1		Block 2		Block 3		Total	
	Value	Chroma	Value	Chroma	Value	Chroma	Value	Chroma
Algoma	5.8	3.6	nd[b]	nd	nd	nd	5.8	3.6
Nipissing	5.3	5.7	4.8	5.0	nd	nd	5.0	5.3
Battlefield	5.0	5.0	6.1	5.8	5.8	6.0	5.6	5.6
Main Algonquin	6.1	5.8	5.8	5.7	nd	nd	5.9	5.7

a: N = 9 for all blocks except Algoma Block 1 (N = 10) and Battlefield Block 3 (N = 6). Hue is 10YR for all observations..
b: no data

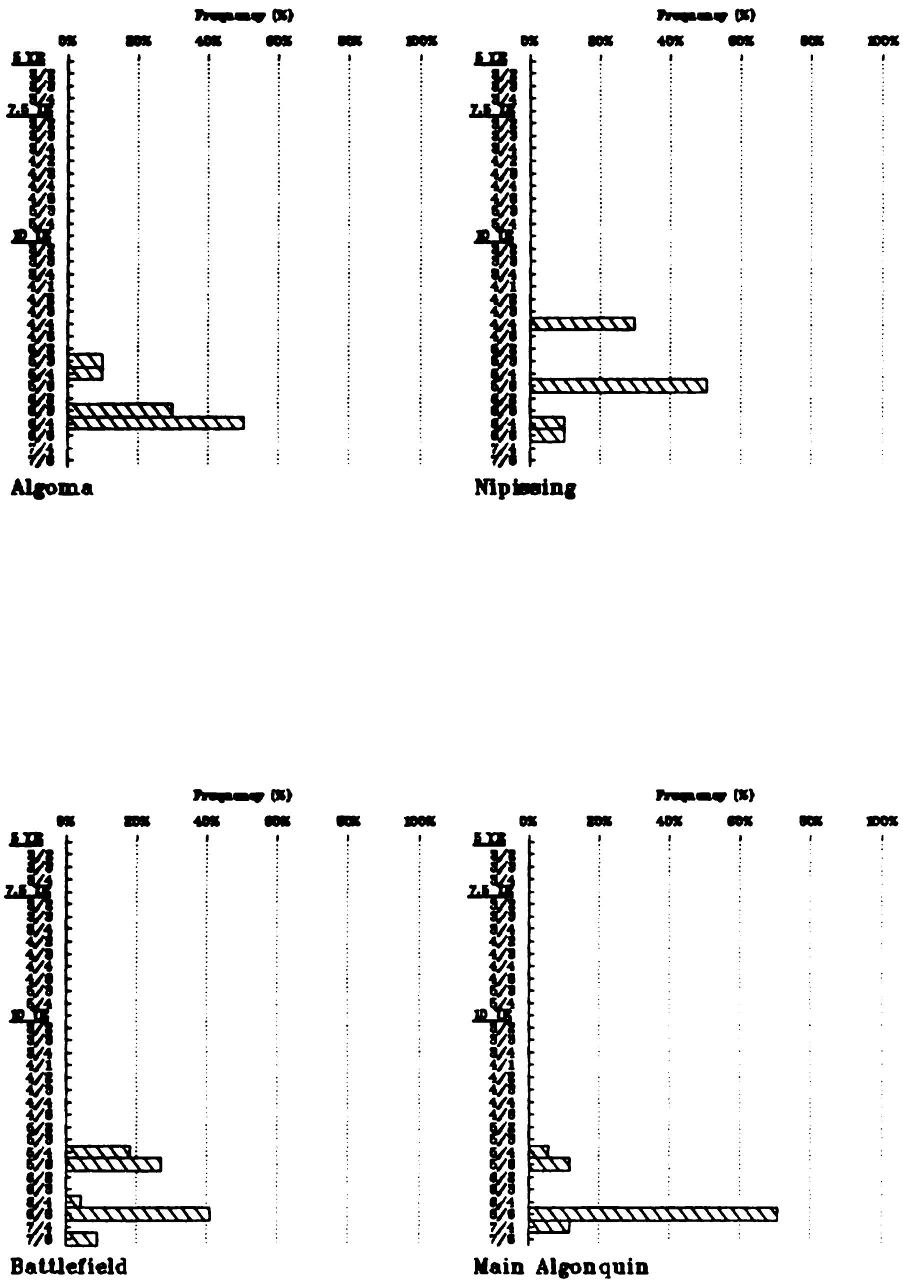

Figure 3. Distribution of C horizon colors by surface for four surfaces: Algoma, Nipissing, Battlefield, and Main Algonquin.

Change in soil properties with surface age

Other field-determined measurements examimed in this study are related to soil development rather than parent material properties. Differences in the data between surfaces may reflect changes in, or degree of, soil development, which may in turn be related to surface age. Within a single surface of assumed uniform age, variability must be related to factors other than age. Field measurements that may vary with soil development in sandy Michigan Spodosol-like soils include: 1) E and B horizon color; 2) the frequency with which ortstein is present in a given pedon and its degree of cementation; 3) the depth to the top of the B horizon, or the thickness of the eluvial horizon(s); 4) solum thickness; and 5) the POD Index (Franzmeier and Whiteside 1963a; Soil Survey Staff 1975; Evans and Cameron 1985; Schaetzl and Mokma 1988). Changes in these measurements can be observed between surfaces in the present study, as shown in the following sections.

E horizon color. The Munsell moist color of the E horizon[4] is very similar for all pedons, both within and between the surfaces (Table 4); 10YR 5/2 and 10YR 4/2 are most common (Figure 4). The hue, in all cases, was 10YR. The mean E horizon value was lower on the Algoma surface, as no 10YR 5/2 horizon colors were observed. The similarity of E horizon colors on the three older surfaces probably reflects the similarity of the color of the quartz parent material sand grains. If the differences between the E horizon and the C horizon colors reflect primarily the loss of Fe-based coatings on the individual grains of the E

[4] In the case of a pedon with more than one E subhorizon, the subhorizon with the highest value was used in all calculations in this section, as per Schaetzl and Mokma (1988).

horizon (Schaetzl and Mokma 1988), once the coating is lost, the color of the E horizon is unlikely to change further, except in response to additions of organic matter from the A horizon. That the E horizons on the Algoma surface are darker in color (lower value) than the others may be due in part to residual Fe coatings, and in part to the presence of a higher organic matter content, with its characteristic lower chroma, since in weakly developed podzol-like soils distinguishing between an A horizon with low organic matter content and an E horizon with high organic matter content is often difficult. The E horizons on the Algoma surface are also characteristically thin.

Table 4. Mean highest-value E subhorizon Munsell value and chroma for pedons on four surfaces.[a]

Surface	Block 1		Block 2		Block 3		Total	
	Value	Chroma	Value	Chroma	Value	Chroma	Value	Chroma
Algoma	4.0	1.9	nd[b]	nd	nd	nd	4.0	1.9
Nipissing	4.6	1.9	5.0	2.0	nd	nd	4.8	1.9
Battlefield	4.6	1.8	4.8	2.0	4.3	2.0	4.6	1.9
Main Algonquin	4.6	2.0	4.6	1.9	nd	nd	4.6	1.9

a: N = 9 for all blocks except Algoma Block 1 (N = 10) and Battlefield Block 3 (N = 6). Hue is 10YR for all observations.
b: no data

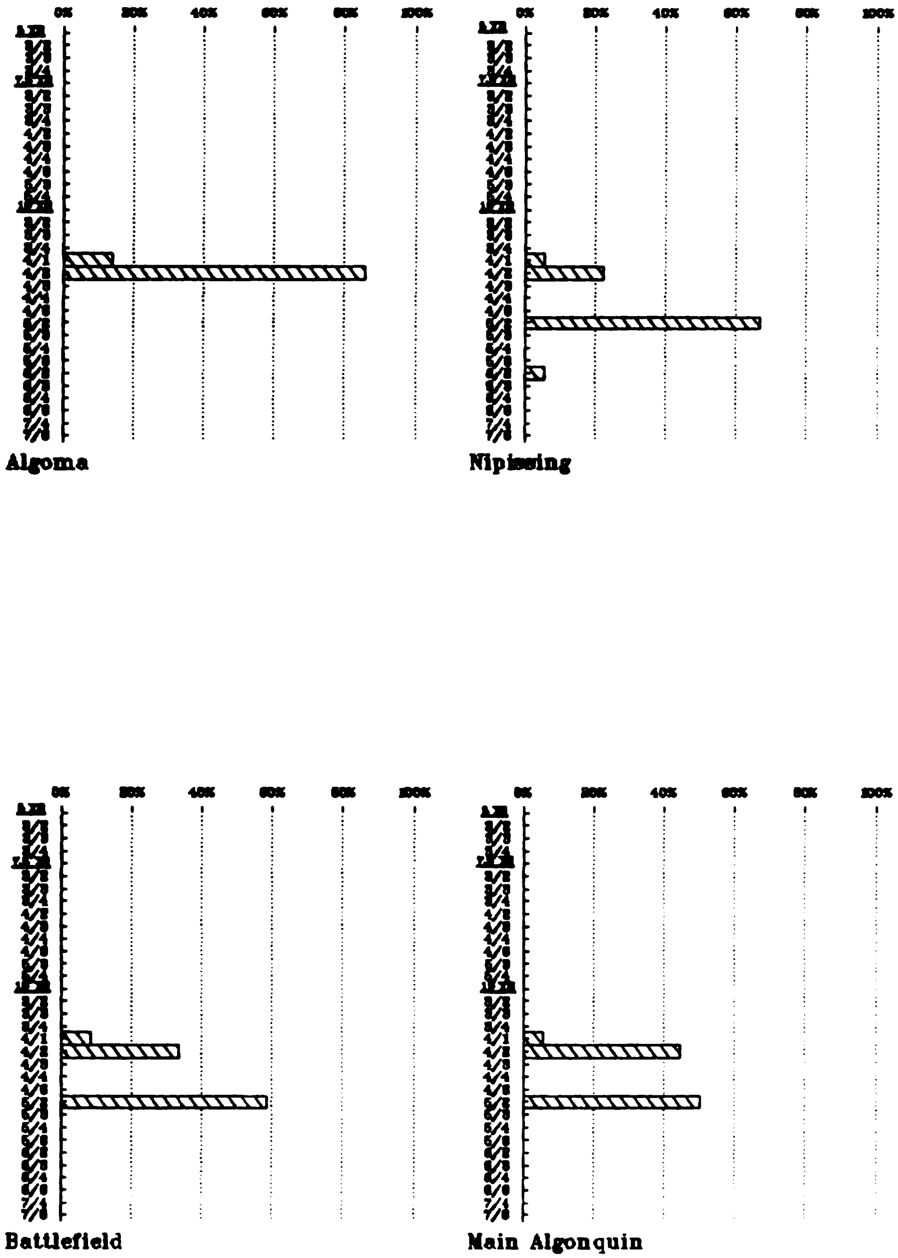

Figure 4. Distribution frequency of highest-value E subhorizon Munsell colors in pedons on four surfaces.

__B horizon color.__ Although on the Algoma surface all B subhorizons had hues of 10YR, the percentage of pedons with the lowest-value[5] B subhorizon[6] redder than 10YR increases with surface age (Figure 5). On the three oldest surfaces, the mean value of the darkest B subhorizon decreases with surface age (Table 5). 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 (see Table 17 in Appendix 2) (Evans and Cameron 1985). The B horizons of the older surfaces also have generally lower chromas, (Table 5), which can be attributed to the accumulation of organic C in the B horizon (Table 17 in Appendix 2) (Evans and Cameron 1985).

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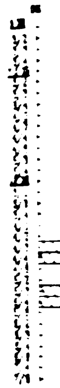
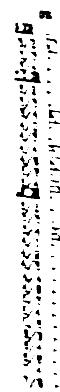
[5] To avoid confusion in terminology, hereafter the term "lowest value" will be considered synonymous with "darkest."

[6] The lowest-value B subhorizon was used in the following calculations in order to show the largest possible contrast in value between the E and B subhorizons. Often the lowest-value B subhorizon was also the B subhorizon with the reddest hue.



Algoma



Batt

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pedons

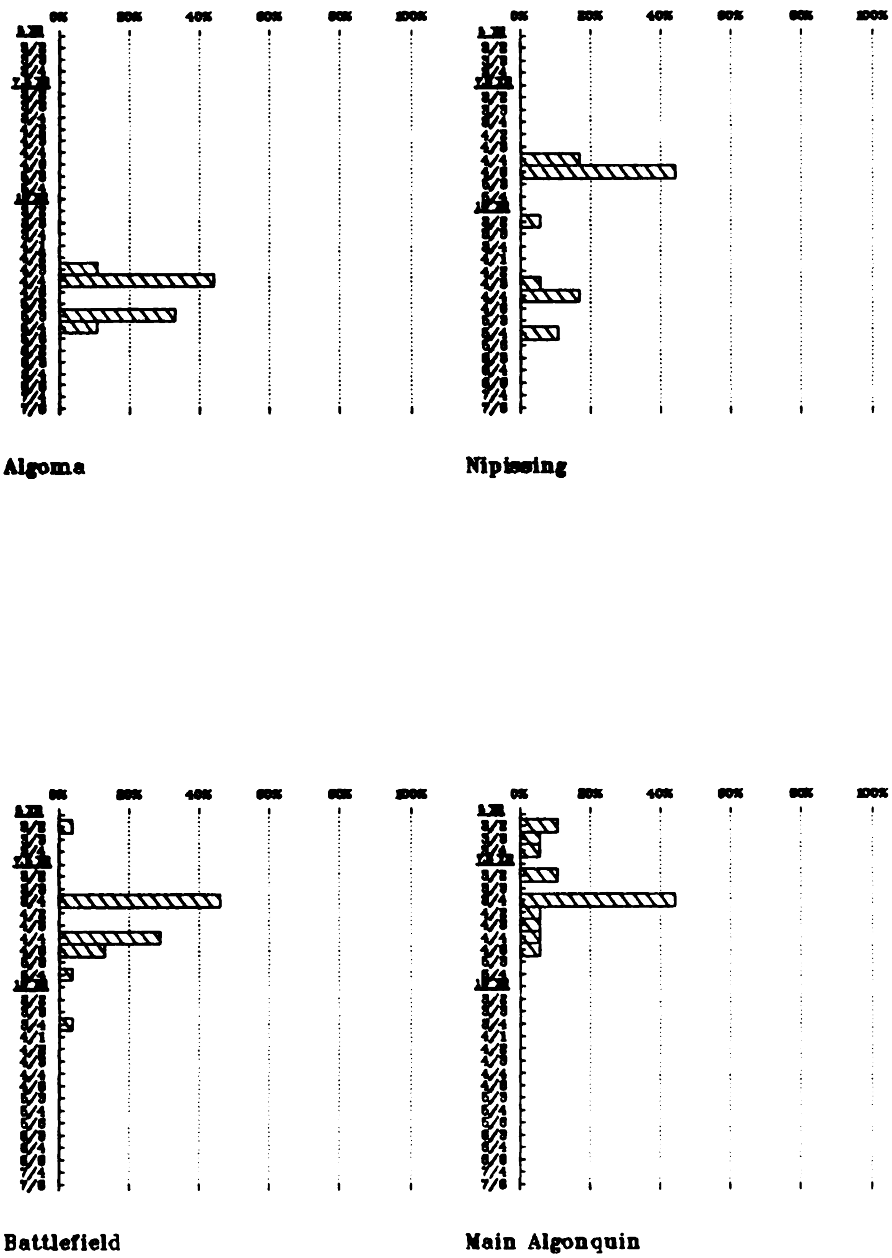

Figure 5. Distribution of lowest-value B subhorizon Munsell colors in pedons on four surfaces.

Table 5. Mo
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Surface
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Main Algonq

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Table 5. Mean lowest-value B subhorizon Munsell value and chroma in pedons on four surfaces.[a]

Surface	Block 1		Block 2		Block 3		Total	
	Value	Chroma	Value	Chroma	Value	Chroma	Value	Chroma
Algoma	4.4	3.6	nd[b]	nd	nd	nd	4.4	3.6
Nipissing	4.2	4.3	3.9	5.2	nd	nd	4.1	4.8
Battlefield	3.0	3.8	3.8	4.4	3.8	4.3	3.5	4.2
Main Algonquin	3.2	3.9	3.2	3.0	nd	nd	3.2	3.4

a: N = 9 for all blocks except Algoma Block 1 (N = 10) and Battlefield Block 3 (N = 6).
b: no data

Ortstein. The number of pedons in which ortstein (Bsm or Bhsm material) was encountered in at least one subhorizon increases with increasing surface age (Table 6). Ortstein was absent in all soils on the Algoma surface, while 11 % of the pedons on the Nipissing surface showed weakly developed ortstein in at least one B subhorizon. Ortstein was noted in 42 % of the pedons on the Battlefield surface and in 89 % of the pedons on the Main Algonquin surface. In general, the strength of ortstein cementation also tended to increase with surface age.

Table 6. Percent of pedons containing ortstein.[a]

Surface	Block 1	Block 2	Block 3	Total
	--------------%	---------------		--%--
Algoma	0	nd[b]	nd	0
Nipissing	22	0	nd	11
Battlefield	44	56	17	42
Main Algonquin	89	89	nd	89

a: N = 9 for all blocks except Algoma Block 1 (N = 10) and Battlefield Block 3 (N = 6).
b: no data

<u>Depth to the top of the B horizon.</u> The depth to the top of the B horizon is very similar for the Nipissing, Battlefield, and Main Algonquin surfaces; an analysis of variance test shows no significant difference between the means between these surfaces. The mean and median depths do, however, appear to increase slightly with surface age (Table 7).

On the Algoma surface, the B horizon is encountered at much shallower depths than on the other surfaces (Figure 6). The narrow range and shallow depths to the top of the B horizon on the Algoma surface can be attributed to the weak development of the E horizon, and probable lack of boundary irregularities, on that surface. 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 as well.

On the three older surfaces, the range of depths to the B horizon is quite large within each surface, and varies little among the surfaces. It should be noted, however, that these data were derived by augering, and the depth recorded as the depth at the point of augering only. Therefore, it is likely that the large range of depths to the B horizon on these surfaces is partly due to boundary irregularity and waviness, which cannot be detected by augering alone.

The increase in depth to the B horizon is statistically significant when regressed against surface age ($p < 0.001$), but due to the wide scatter on each surface, the association is very weak (adj. $r^2 = 0.13$), indicating that surface age explains only a small part of the variability in the depths to the top of the B horizon on these surfaces (Figure 7).

Previous chronosequence studies of Spodosols and Podzolic soils in sandy materials have reported conflicting results regarding the trend in depth to the top of the B horizon. Franzmeier and Whiteside (1963a) felt that, following an initial rapid increase in depth for the first 3000 years of development, the trend either leveled off or reversed, so 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 of Podzols in Denmark, because the thickness of the E horizon did not increase with increasing age and the B horizon grew upward into the eluvial horizon. In Finland, however, a positive correlation existed between soil age and the thickness of the E horizon (Koutaniemi *et al.* 1988). The findings of the present study appear to most closely parallel those of Koutaniemi *et al.* 1988.

Table 7. Depth to the top of the B horizon in pedons on the four geomorphic surfaces.

Surface	N	Mean[a]	Median	Minimum	Maximum	St. Dev.
		cm				
Algoma	10	13 y	15	0[b]	20	6.23
Nipissing	18	26 z	25	10	60	11.50
Battlefield	24	28 z	25	8	60	10.90
Main Algonquin	18	31 z	30	15	50	9.03

a: Numbers within the column followed by the same letter are not significantly different (LSD, $p < 0.05$).

b: Indicates pedon contains no B horizon. In pedons that contain a B horizon, the minimum depth to the top of the B horizon is 8 cm.

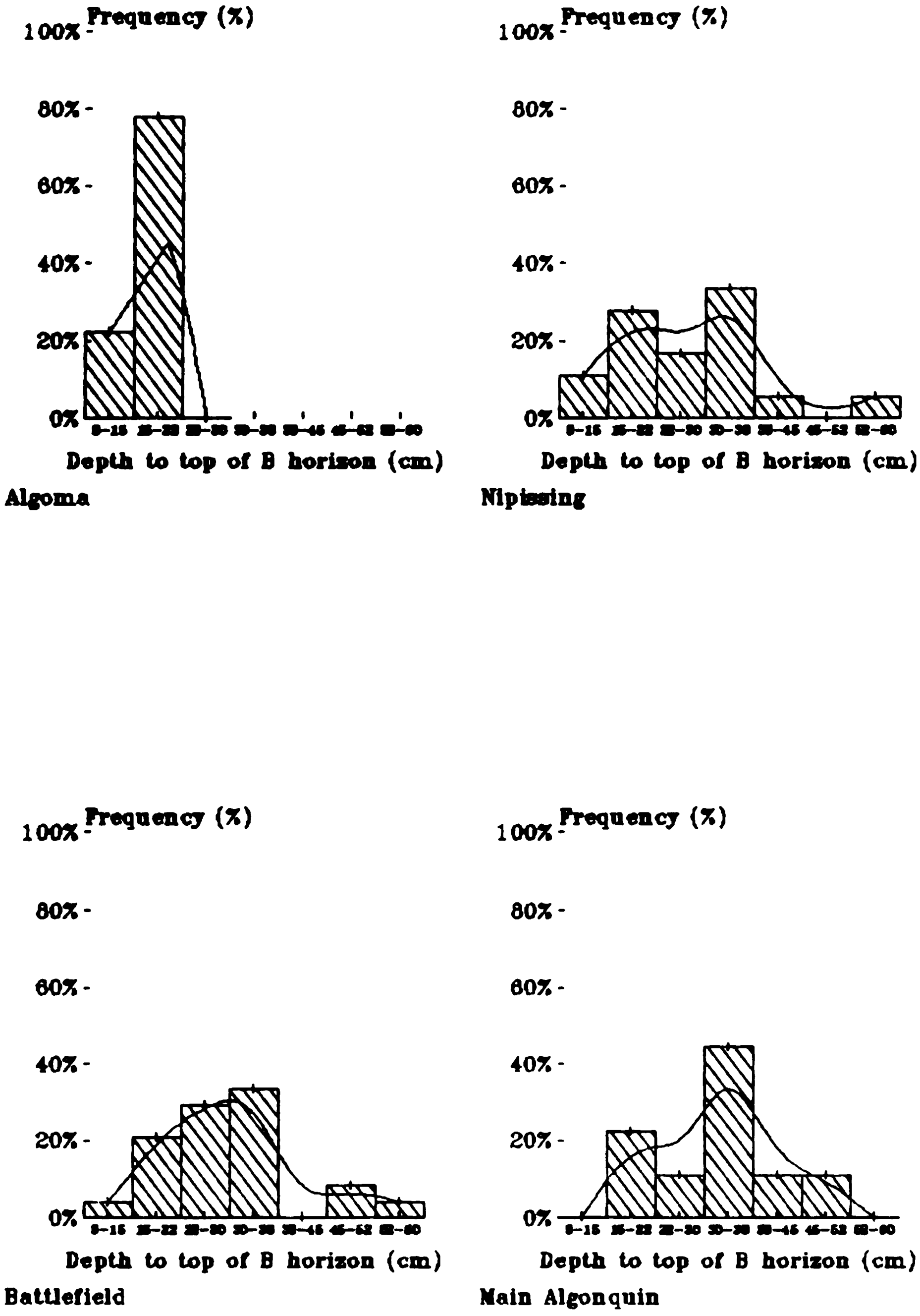

Figure 6. Frequency of depths to B horizon on four surfaces.

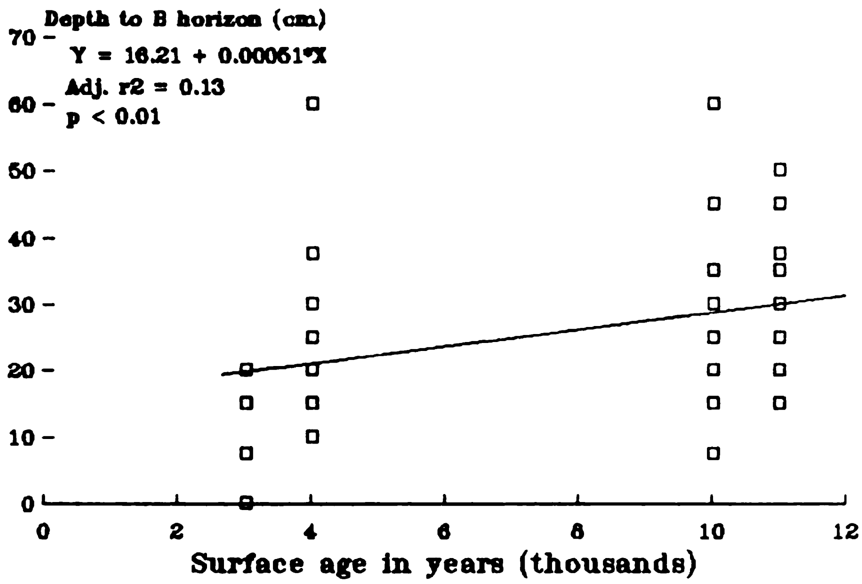

Figure 7. Depths to the B horizon as a function of surface age.

<u>B horizon thickness.</u> B horizon thickness mean, median, minimum, and maximum increase with increasing surface age (Table 8). The analysis of variance test, however, shows no statistically significant ($p < 0.05$) differences in means between the Algoma and Nipissing surfaces or the Battlefield and Main Algonquin surfaces. Because the members of each pair of surfaces are separated by only a small age difference, some similarity in soil development is expected. The increase in B horizon thickness is statistically significant when regressed against surface age ($p < 0.0001$), and, although the association is rather weak (adj. $r^2 = 0.39$), the relationship is stronger than it was for eluvial horizon thickness (Figure 8). An increase in B horizon thickness with increasing surface age has also been reported for a chronosequence of river terrace podzols in northern Finland (Koutaniemi *et al.* 1988).

Table 8. B horizon thickness on the four geomorphic surfaces.

Surface	N	Mean[a]	Median	Minimum	Maximum	St. Dev.
		cm				
Algoma	10	43 y	45	18	60	13.96
Nipissing	18	49 y	50	23	75	14.45
Battlefield	24	66 z	70	40	90	15.88
Main Algonquin	18	75 z	73	53	110	13.94

a: Numbers within the column followed by the same letter are not significantly different (LSD, $p < 0.05$).

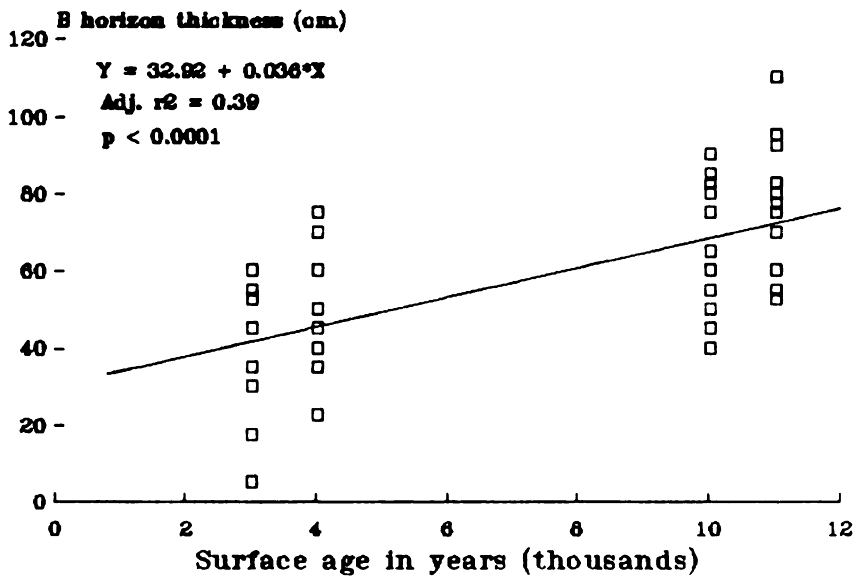

Figure 8. B horizon thickness as a function of surface age.

Solum thickness. Although the range of solum thickness values is quite broad within a single surface (Figure 9), the minimum, maximum, and mean values all increase with surface age (Table 9). Regression of solum thickness against surface age produces an equation significant at $p < 0.001$, with an adjusted r^2 of 0.49 (Figure 10), indicating that the sola become thicker with increasing surface age. Similarly, the thickness of both E and B horizons was positively correlated with soil age in sandy soils of northern Finland (Koutaniemi *et al.* 1988); i.e., older soils had thicker sola. Analogous findings have been reported for chronosequences of soils in calcareous sands, where the solum thickness is reflected by depth of leaching. For example, in dune soils of southern Michigan, the zone leached of carbonates had deepened to 2 m in 1000 years (Olson 1958). On beach ridges near Hudson Bay, Ontario, a strong linear relationship was found between depth of leaching and age (Protz 1984).

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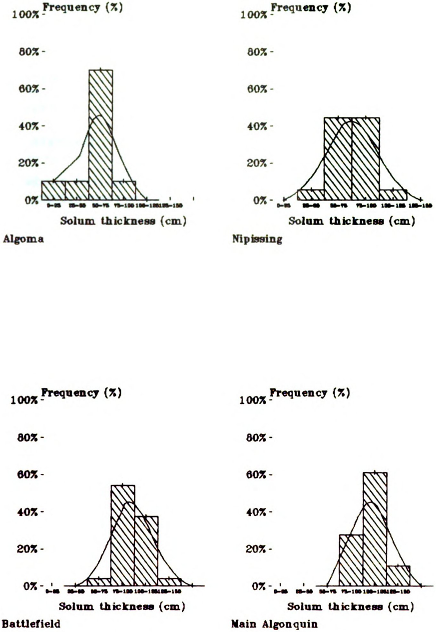

Figure 9. Distribution of solum thicknesses by surface.

Table 9.

Surface
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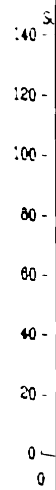

Figure 10.

Table 9. Solum thickness on the four geomorphic surfaces.

Surface	N	Mean[a]	Median	Minimum	Maximum	St. Dev.
		cm				
Algoma	10	52 w	55	5	75	19.65
Nipissing	18	73 x	70	40	105	18.80
Battlefield	24	95 y	90	60	135	18.55
Main Algonquin	18	106 z	105	90	135	12.55

a: Numbers within the column followed by the same letter are not significantly different (LSD, $p < 0.05$).

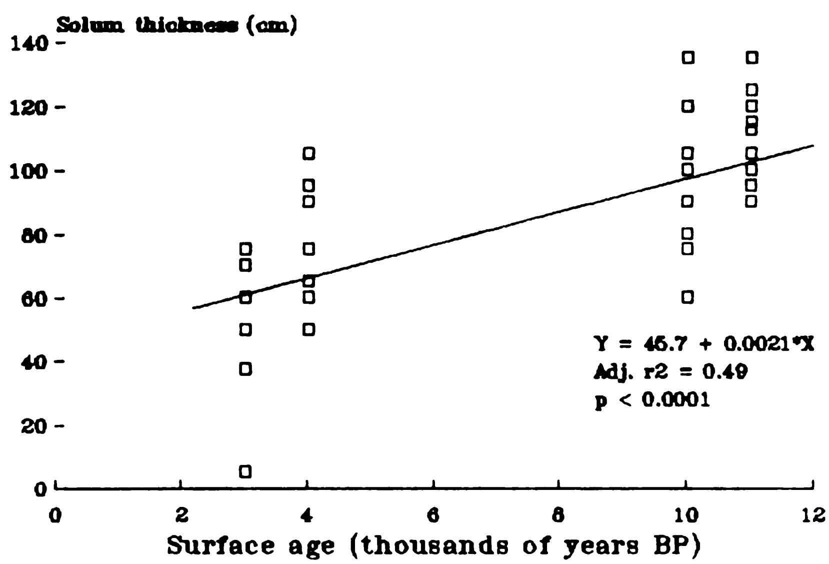

Figure 10. Solum thickness as a function of surface age.

<u>POD Index.</u> The mean, median, and maximum values of the POD Index values (Schaetzl and Mokma 1988) increase with surface age (Table 10). The range and standard deviation of POD Index values also increase, in part due to the large number of pedons with a POD Index of 0 on the younger surfaces (Figure 11). A POD Index of 0, however, can actually denote a wide range of development. A regression of POD Index against surface age produces an equation significant at $p < 0.0001$, but due to a wide variability on each surface, and the lower limit to the POD Index of 0, the adjusted r^2 is only 0.23 (Figure 12). POD Index values of 0-2 merely suggest that the soils are non-Spodosols, in this case, Entisols. 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 (Entic Haplorthods), and 6 or greater (Typic Haplorthods). Thus, based on the POD Index value, the pedons on these four surfaces have been assigned to a probable Soil Taxonomy classification (Soil Survey Staff 1975) (Table 11). Using this method, the percentage of pedons classified as Spodosols increases with surface age.

Table 10. POD Index on four surfaces, by surface.

Surface	N	Mean[a]	Median	Minimum	Maximum	St. Dev.
Algoma	10	0.0 x	0	0	0	0.00
Nipissing	18	2.2 x,y	2	0	8	2.16
Battlefield	24	4.3 y	4	0	14	3.87
Main Algonquin	18	7.6 z	6	0	18	6.07

a: Numbers within the column followed by the same letter are not significantly different (LSD, $p < 0.05$).

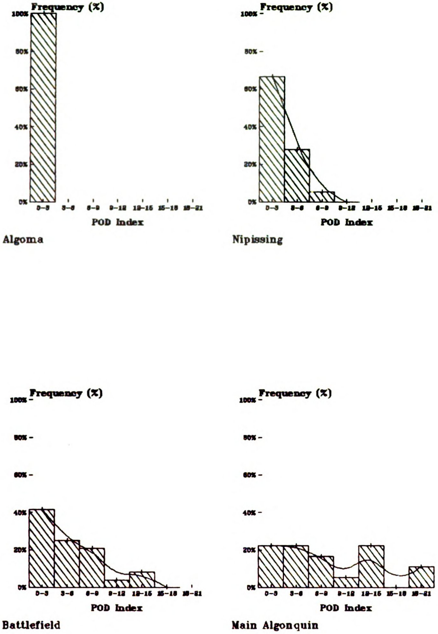

Figure 11. Distribution of POD Index values of pedons on four surfaces.

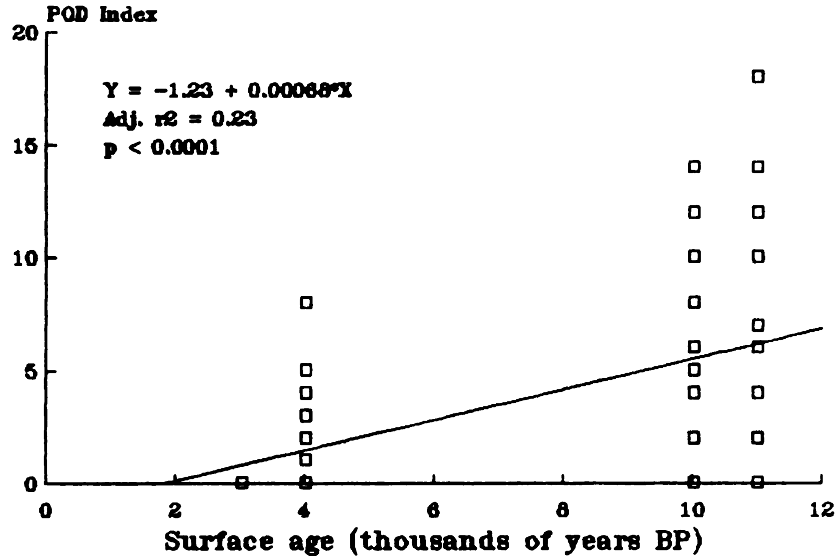

Figure 12. POD Index as a function of surface age.

Table 11. Probable Soil Taxonomy classifications of pedons on four surfaces based on POD Index values (Schaetzl and Mokma 1988).

Soil Taxonomy: POD Index value[a]:		Udipsamment 0-2		Entic Haplorthod 2-6		Typic Haplorthod 6+	
	N	Freq.	%	Freq.	%	Freq.	%
Algoma	10	10	100	0	0	0	0
Nipissing	18	12	67	10	56	1	6
Battlefield	24	10	42	14	58	8	33
Main Algonquin	18	4	22	7	39	10	56

a: Due to the overlap in POD Index values between classification categories, totals will add up to greater than 100 %.

Soil variability on each surface

An examination of the variability statistics for depth to the top of the B horizon reveals that variability decreases slightly with surface age, both in terms of the coefficient of variation (CV) and the standard deviation (SD) (Table 12). The 75th percentile-25th percentile range is similar for all surfaces, and the minimum-maximum range shows no relationship with surface age. The large discrepancy between the maximum-minimum range and the 75th percentile-25th percentile range for the three oldest surfaces illustrates that most of the variability comes from a few outlier pedons on these surfaces. The decreasing trend with surface age of CV for B horizon thickness is accompanied by inconclusive patterns for SD, minimum-maximum range, and 75th percentile-25th percentile range (Table 13). Variability in B horizon thickness on a surface does not appear to be strongly related to surface age.

The CV and the SD exhibit a decreasing trend with surface age for solum thickness (Table 14), indicating that solum thickness is less variable on the older surfaces. In contrast, although the CV for the POD Index also decreases with surface age, the SD and range measures for the POD Index show an increasing trend (Table 15). The large range and SD of the POD Index on the Main Algonquin surface reflect the disparity in soil development on the surface, from weak Entisol morphology to strongly developed Spodosol morphology. The contrast in range between the younger and older surfaces is exacerbated because a POD Index value of zero can denote quite a broad range of weak soil development, as would be expected on the younger surfaces.

Table 12. Variability in depth to the top of the B horizon by surface.

Surface	N	CV	SD	Range Max-Min	Range 75th Ptl-25th Ptl
		-------------------	------------------ cm	------------------	------------------
Algoma	10	1.2	6.23	20	8
Nipissing	18	1.1	11.50	50	10
Battlefield	24	1.0	10.90	53	8
Main Algonquin	18	0.8	9.03	35	10

Table 13. Variability in B horizon thickness by surface.

Surface	N	CV	SD	Range Max-Min	Range 75th Ptl-25th Ptl
		-------------------	------------------ cm	------------------	------------------
Algoma	10	0.32	13.9	42	18
Nipissing	18	0.29	14.4	52	25
Battlefield	24	0.24	15.9	50	28
Main Algonquin	18	0.19	13.9	57	10

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Table 14. Variability in solum thickness by surface.

Surface	N	CV	SD	Range Min-Max	Range 25th ile-75th ile
		cm			
Algoma	10	0.95	19.7	70	10
Nipissing	18	0.65	18.8	65	30
Battlefield	24	0.50	18.6	75	25
Main Algonquin	18	0.30	12.6	45	18

Table 15. Variability in POD Index by surface.

Surface	N	CV	SD	Range Min-Max	Range 25th ile-75th ile
Algoma	10	0[a]	0	0	0
Nipissing	18	0.97	2.16	8	4
Battlefield	24	0.91	3.87	14	5
Main Algonquin	18	0.80	6.07	18	10

a: All POD Index values were zero on this surface. Thus, CV and SD values are meaningless.

The decreasing variability of the depth to the top of the B horizon and the solum thickness with increasing age is counter to the expectation that soils on the older surfaces would be more variable than those on the younger surfaces, whereas the variability of the POD Index appears to support it. The difference between these two conditions may be a consequence of the types of observations in question. Perhaps depth to the B horizon and solum thickness have a greater dependence than the POD Index on subtle parent material effects that diminish with surface age.

Based on the variability of POD Index values on these surfaces, it appears that variability in soil development increases with surface age, probably because weakly developed soils are found on the older surfaces alongside more strongly developed soils. On the younger surfaces, most soils are weakly developed. Increasing variability of soil development on the older surfaces may reflect the increasing importance with surface age of micro-scale factors of soil development (Alexander 1986), including microclimate (Macyk *et al.* 1978; Alexander 1986), microrelief (Lag 1951; Schaetzl 1990), leaf litter distribution (Rowe 1955; Alexander 1986), and the influence of individual trees (Gersper and Holowaychuk 1970a, b; Crampton 1982; Boettcher and Kalisz 1990). Pedoturbation processes also affect soil development on a micro-scale (Hole 1961; Johnson *et al.* 1987) and may contribute to the increased variability in degree of soil development with time.

Laboratory study

I chose a modal pedon from each surface in order to supplement the field data and to study the physical and chemical characteristics of the soils. These modal pedons were described in detail (see Appendix 1), and sampled by genetic horizon. The samples were analyzed in the laboratory. Two pedons were sampled on the Algoma surface because the presence in the first Algoma pedon (referred to as Algoma-1) of a 2Ab horizon overlying a 2C horizon of coarser sand suggested that the Algoma-1 pedon might have formed in eolian sands. Results of the laboratory analysis are presented and discussed below; data for individual horizons are presented in Appendix 2.

Particle size analysis

Data on coarse fragment distribution and sand size particles have been examined as indicators of the extent of parent material uniformity in these pedons, whereas silt and clay depth functions have been studied to asses pedogenesis.

Coarse fragments. The five pedons have low coarse fragment contents; only the Nipissing pedon has a subhorizon within the solum (Bs2) that contains > 3% coarse fragments by weight (Table 16). Relatively high percentages of coarse fragments are also found in the Algoma-2 2C (60.7%) and Nipissing BC (21.6%) and C (10.6%) horizons.

Sand size particles. The five pedons developed in sandy materials; all subhorizons contain greater than 92% sand-sized particles (Table 26 in Appendix 2). The textures of all subhorizons meet the USDA "sand" textural class designation (Soil Survey Staff 1981). Medium sand is the modal sand size fraction in all subhorizons of these pedons, but the ratio of coarse/fine sand can vary arbitrarily between horizons within a single pedon (Table 16). The sand fraction distribution

(weighted mean of all horizons) for each pedon (Figure 13) indicates that the fine sand fraction makes up a larger component of the Algoma-1, Battlefield, and Main Algonquin pedons, while coarse sand is more prominent in the Algoma-2 and Nipissing pedons. The sand fraction distribution of the C horizon[7] for each pedon, representing relatively unweathered parent material, has a similar pattern (Figure 14).

It would be possible, based on coarse fragment content and sand size distribution within these five pedons, to recognize a number of lithologic discontinuities, e.g., between the Nipissing Bs3 and BC horizons, or between the Algoma-2 Bs and C horizons. The development of these soils in lacustrine sands suggests, however, that the depositional environment was one of varying energy states that may have affected the size distribution within the sand fraction, but not necessarily its mineralogical or chemical properties, or the ultimate pedogenic pathways of the soils. Mineralogy has been shown to be similar across sand-sized particles in a podzol soil developed in fluvioglacial materials in Denmark (Nørnberg 1980). Therefore, despite some differences in particle size distribution, the depositional environment for the parent materials of these pedons is probably uniform. Lithologic discontinuities, therefore, have not been indicated, except where they were obvious and identifiable in the field.

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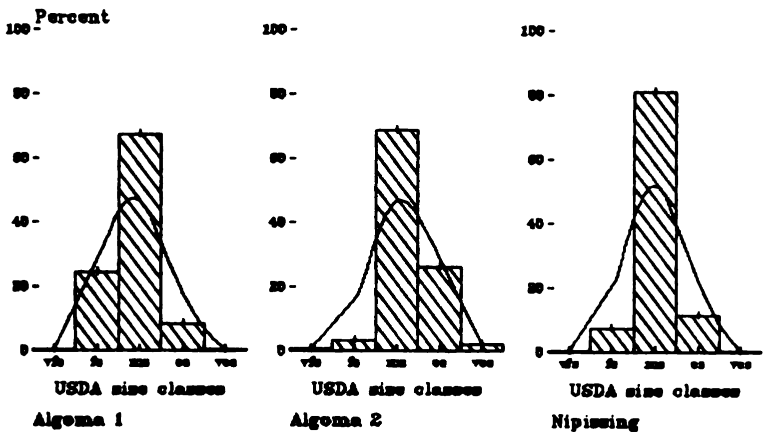
[7] For pedons with a lithologic discontinuity (e.g., Algoma-1), the C horizon used is the C horizon in the first (upper) parent material.

Table 16. Coarse fragments and sand size particles in five pedons.

	Coarse fragments >2.0 (mm)	VCS 2.0-1.0 (mm)	CS 1.0-0.5 (mm)	MS 0.5-0.25 (mm)	FS 0.25-0.1 (mm)	VFS 0.1-0.05 (mm)	CS/FS
	% of total	----------% of fine earth----------					
Algoma-1							
E	0.1	0.6	4.4	56.1	38.7	0.1	0.11
Bs	0.0	0.0	4.8	71.2	24.0	0.1	0.20
C	0.0	0.0	3.2	68.9	27.8	0.1	0.12
2Ab	0.1	0.1	8.7	66.9	24.2	0.2	0.36
2C	0.4	0.8	15.4	65.6	17.9	0.3	0.86
Algoma-2							
E	0.3	0.4	23.5	73.5	2.4	0.1	9.79
Bs	0.2	0.2	20.1	76.4	3.2	0.1	6.28
C	1.9	0.9	32.4	64.4	2.2	0.1	14.73
2C	60.7	8.2	36.2	48.8	4.9	2.0	7.39
Nipissing							
A	0.6	0.7	9.4	80.5	8.8	0.5	1.07
E1	0.1	0.2	7.9	81.4	9.9	0.7	0.80
E2	0.3	0.1	6.3	82.2	10.9	0.4	0.58
Bs1	2.3	0.1	7.1	85.1	7.6	0.2	0.93
Bs2	7.6	0.3	11.7	83.2	4.8	0.1	2.44
Bs3	2.3	0.9	22.0	73.2	3.8	0.1	5.79
BC	21.6	1.2	14.4	75.7	8.4	0.3	1.71
C	10.6	0.2	4.8	84.2	10.6	0.2	0.45
Battlefield							
A	0.5	0.8	10.0	65.3	22.8	1.1	0.44
E1	0.4	0.5	10.1	67.0	21.1	1.3	0.48
E2	2.1	0.6	7.8	64.0	26.7	0.9	0.29
Bs1	1.6	0.8	7.6	64.9	26.0	0.7	0.29
Bs2	0.9	0.6	7.2	63.2	28.7	0.4	0.25
Bs3	0.4	0.5	7.1	62.5	29.6	0.3	0.24
BC	0.4	0.5	3.9	46.2	48.5	0.9	0.08
C1	0.2	0.3	2.8	56.1	40.4	0.5	0.07
C2	0.1	0.5	14.2	72.8	12.4	0.1	1.15
Main Algonquin							
A	0.6	0.8	13.5	77.2	7.6	0.9	1.78
E	0.9	0.3	11.1	79.6	8.2	0.8	1.35
Bhs	0.5	0.2	7.2	81.9	10.4	0.3	0.69
Bs1	0.2	0.1	4.4	81.5	13.8	0.2	0.32
Bs2	0.0	0.1	3.8	81.9	14.1	0.2	0.27
BC	0.0	0.1	2.3	71.6	25.9	0.1	0.09
C	0.7	0.4	6.3	68.9	24.0	0.3	0.26



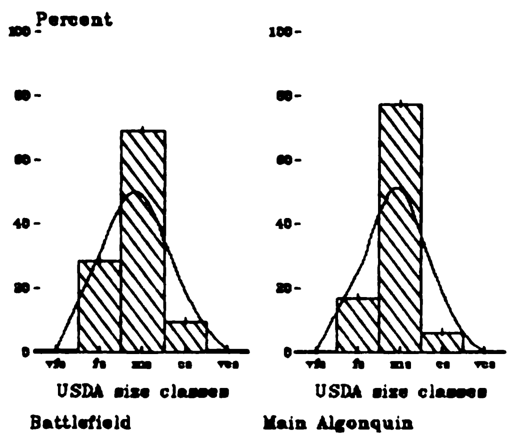


Figure 13. Frequency distribution of sand size particles, weighted mean of all horizons, for five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.

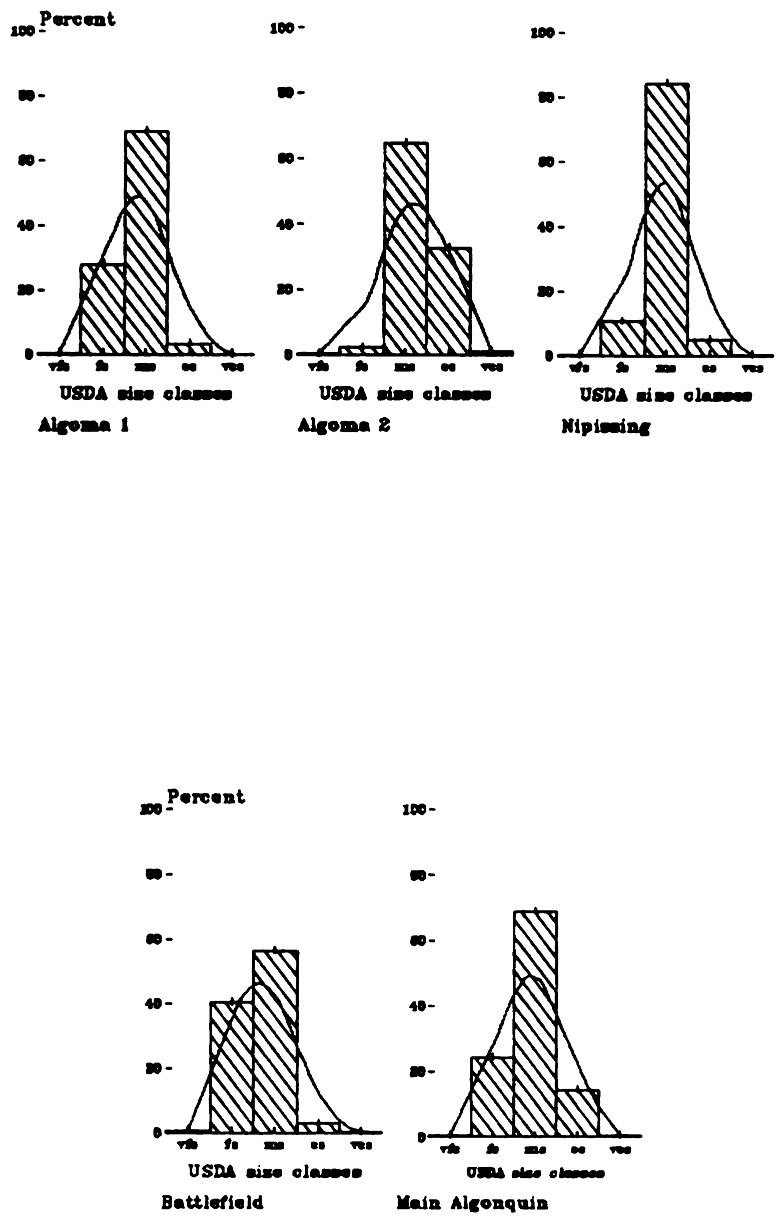


Figure 14. Frequency distribution of sand size particles in the C horizon for five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.

Silt. In these soils, maximum silt content is found in near-surface horizons; silt content decreases gradually with depth[8] (Figure 15). The percentage of silt in the near-surface horizons increases with surface age. The depth functions of the very fine sand fraction are similar, although not as pronounced (Table 16).

Reports of similar particle size distributions in sandy Spodosols and Spodosol-like soils are common (Franzmeier 1962; Jauhiainen 1973; Nørnberg 1977; Wang *et al.* 1978; Nørnberg 1980; Protz 1984; Wang and McKeague 1986), although the mechanisms that produce such distributions have not been well studied. Franzmeier (1962), Protz (1984), and Nørnberg (1977; 1980) suggested that silt-rich surface horizons are the result of *in situ* physical weathering of sand-sized particles. Increased physical weathering near the surface could produce a silt fraction that decreases with depth. This trend would likely get more pronounced with age of the soil. Jauhiainen (1973) felt that silt-enriched surface horizons could be best attributed to eolian inputs shortly after surface formation. It is also possible that eolian inputs to the surface are a gradual, continuing process (Yaalon and Ganor 1973). Further study is necessary to fully explain the origin of the increase in near-surface horizon silt.

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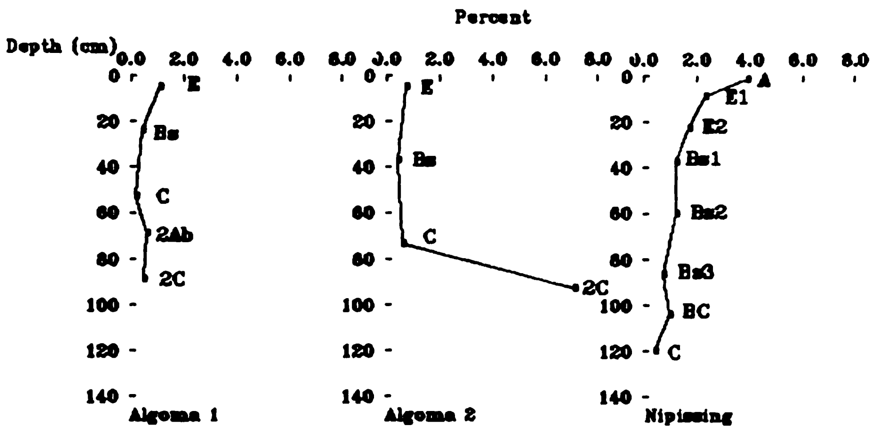
8 The relatively high silt content of the Algoma-2 2C horizon (7.1%) occurs below a lithologic discontinuity.



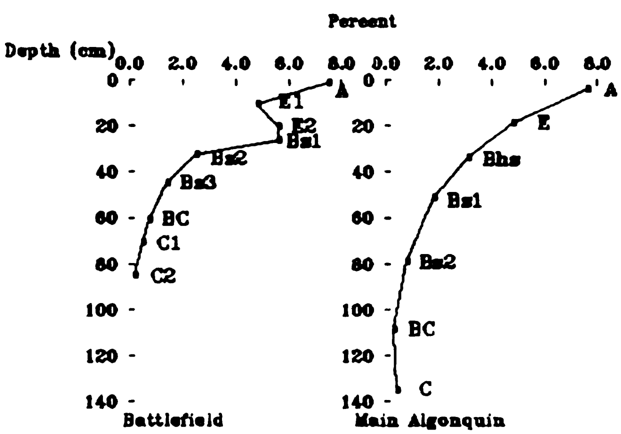
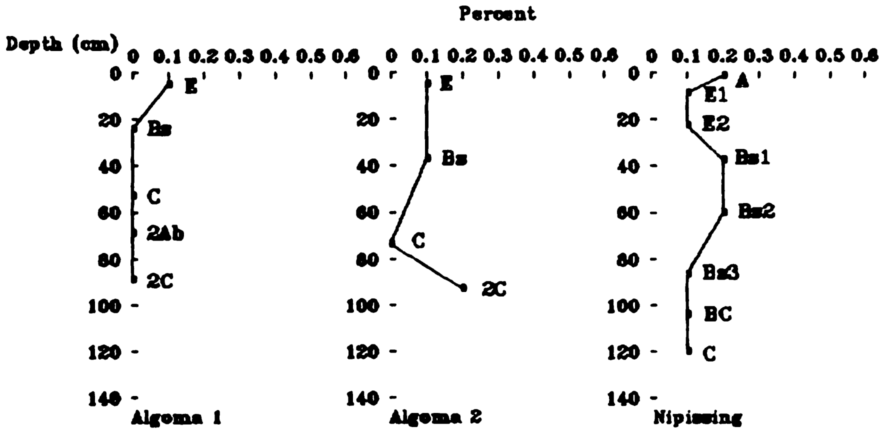


Figure 15. Percent silt with depth for five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.

Clay. Clay content of all pedons is minimal; all subhorizons have clay contents below 0.5%. Nevertheless, the distribution of clay with depth reveals a subsurface maximum (usually in the upper B horizon) that gets larger with surface age (Figure 16). Due to the extremely low clay content of these soils and the small number of pedons examined, the significance of this finding is questionable. Nevertheless, both Franzmeier (1962) and Nørnberg (1977) found similar distributions of clay with depth in the older members of Spodosol chronosequences, suggesting that small amounts of translocated clay may accumulate in the B horizons of Orthods. Clay illuviation, of varying degrees, has also been reported in other Spodosols (Brydon 1965; Wang and Rees 1980; Stanley and Ciolkosz 1981; Padley *et al.* 1985).



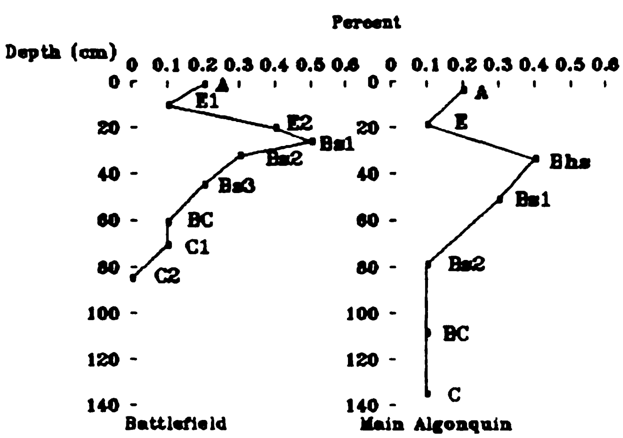


Figure 16. Percent clay with depth for five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.

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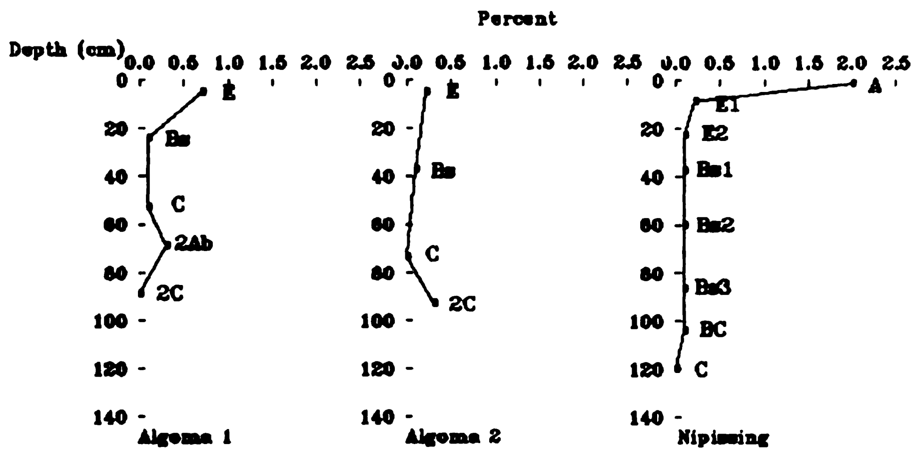
Organic C

The highest OC content in all pedons occurs in the O horizon (or A horizon, if no O horizon is present). Percent OC in the surface horizon shows no discernable trend with surface age (Table 17 in Appendix 2). Subsurface OC maxima are found in the upper B horizons of the Battlefield and Main Algonquin pedons, but are absent in the Nipissing and Algoma pedons (Figure 17). The slight increases in OC content in the Algoma-1 2Ab and Algoma-2 2C horizons are likely due to additions of organic matter prior to the deposition of the overlying parent materials.

The distribution of ODOE with depth is similar to the OC distribution (Figure 18), although the A horizon maxima are not as large and the B horizon maxima are more pronounced. In the Battlefield pedon, however, the ODOE of the Bs2 horizon (0.223) is much greater than that of the Bs1 (0.090), despite the fact that the OC contents of the Bs1 and Bs2 horizons are alike (0.8%). The ODOE distribution, therefore, does not simply mirror the OC distribution; ODOE maxima do, however, apparently coincide with the horizons containing the largest and most strongly cemented ortstein fragments (Appendix 1).

Since the ODOE test was designed as an indication of the presence of fulvic acids (Daly 1982), the high ODOE of some horizons would suggest that fulvic acids, which may be important in the translocation of Fe and Al in Spodosols (DeConinck 1980), have accumulated in ortstein-rich horizons. The magnitude of the ODOE B horizon maximum increases with surface age, suggesting increasing deposition of fulvic acids in the B horizon with time. An association of fulvic acids with ortstein has been reported for Spodosols in Florida, and was attributed to the greater reactivity of fulvic acids as opposed to humic acids (Lee *et al.* 1988a, b). The significance of the relationships between ODOE, OC, and ortstein in well-drained Michigan Spodosols is unclear from the present data and merits further study.



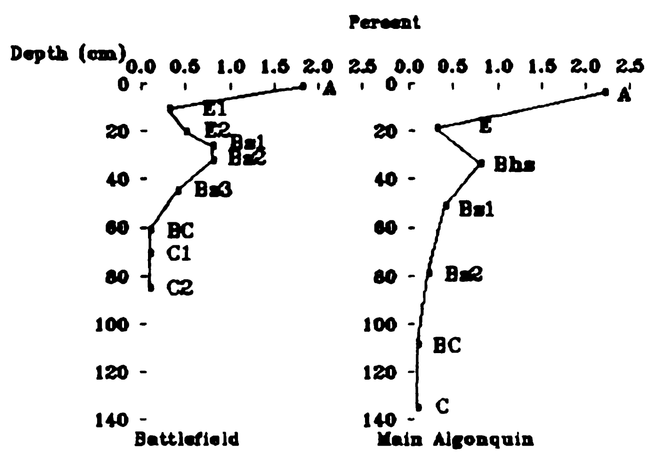
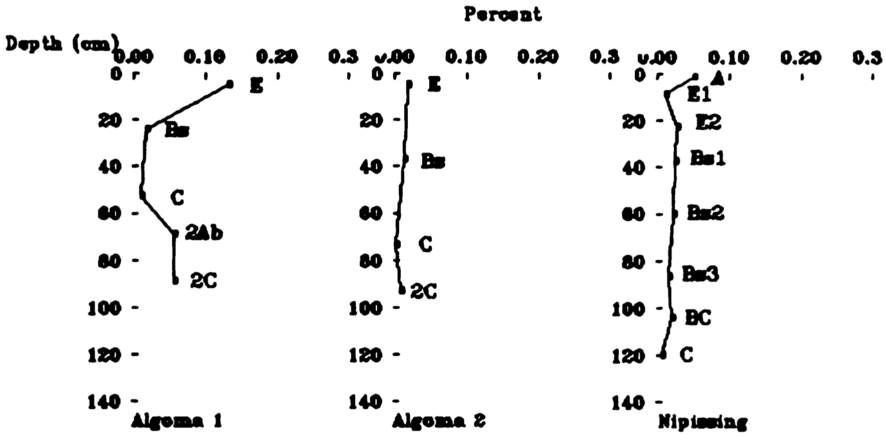


Figure 17. Percent OC with depth for five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.



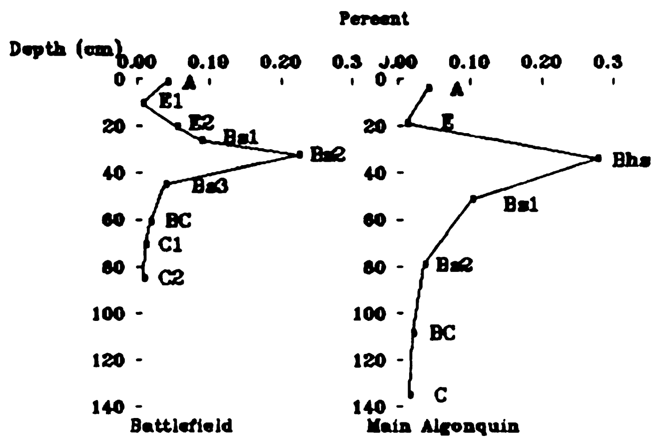


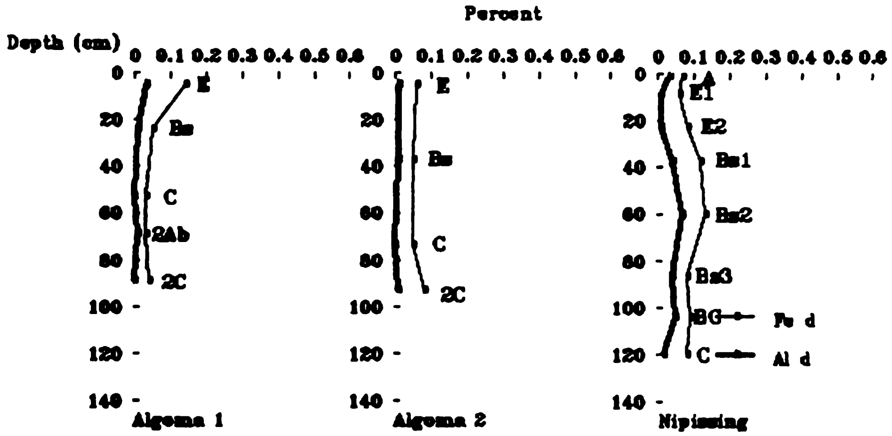
Figure 18. ODOE with depth for five pedons: Algoma-1, Algoma-2, Nipissing, Battlefields, and Main Algonquin.

Fe and Al

The accumulation of Al, along with OC, in the B horizon is important to the identification of spodic materials; Fe accumulation is often also a characteristic feature of spodic materials (Soil Survey Staff 1975). In the present study, three different Fe and Al extractants were used: (1) citrate-bicarbonate-dithionite (CBD), (2) Na pyrophosphate, and (3) acid ammonium oxalate. These extractants attack different forms of the Fe and Al present in the soil (McKeague 1978), and are discussed below (p. 99).

Fe_d and Al_d. The depth functions for both CBD-extractable Fe (Fe_d) and Al (Al_d) in these pedons show B horizon maxima that tend to increase with surface age (Figure 19), although the maximum Al_d content of the Main Algonquin Bs1 subhorizon (0.18%) is less than that of the Bs2 subhorizon in the Battlefield soil (0.29%).

Fe_o and Al_o. The general patterns of oxalate-extractable Fe (Fe_o) and Al (Al_o) with depth in these soils are similar to those of Fe_d and Al_d: the E horizon minimum and B horizon maximum (and the B-E horizon contrast) tend to increase in magnitude with surface age (Figure 20), except that, as with Al_d, the Battlefield Al_o maximum (Bs2 subhorizon) exceeds that of the Main Algonquin pedon (Bs1 subhorizon). The Al_o concentration of the C horizon, which is often the minimum or near-minimum value within a pedon, also increases slightly with surface age, from the Algoma-1 (0.00%) to the Main Algonquin (0.08%).



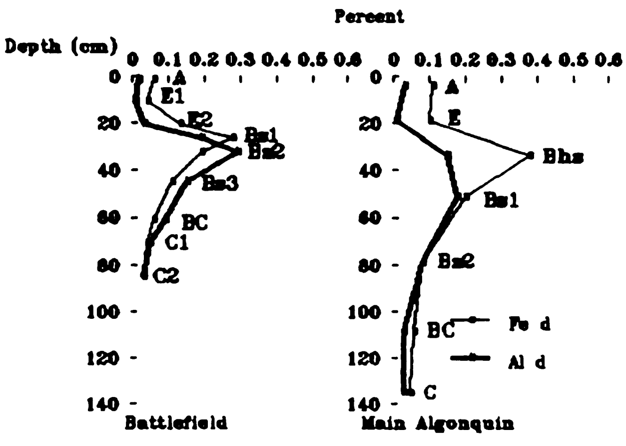
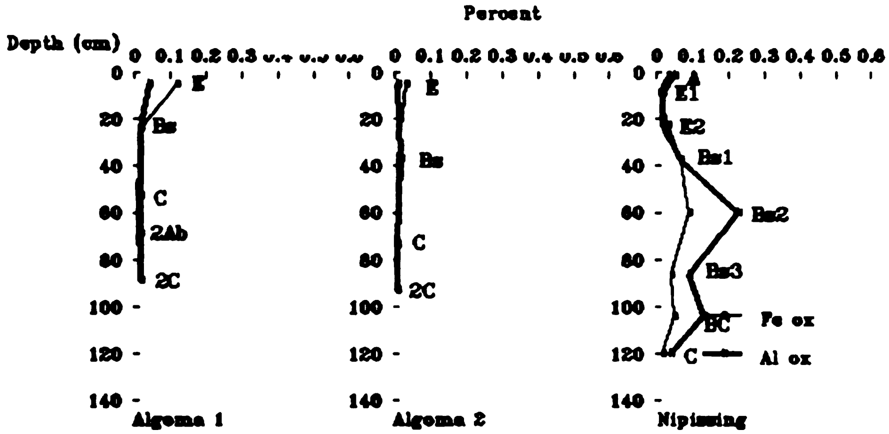


Figure 19. Percents Fe_d and Al_d as a function of depth for five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.



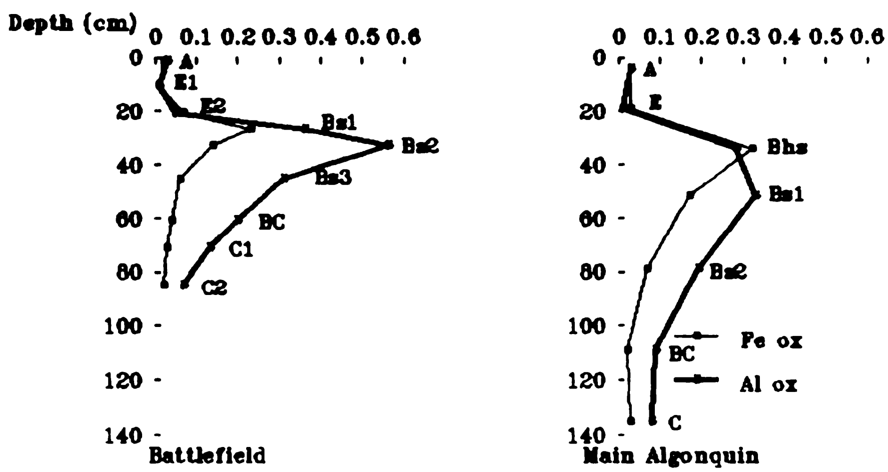


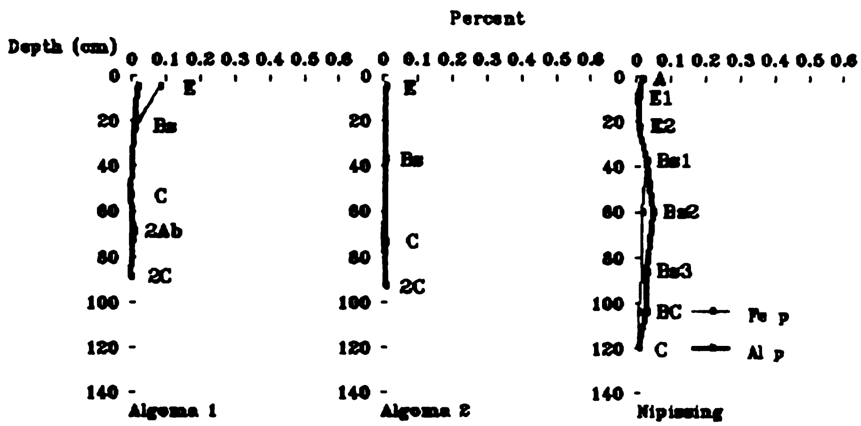
Figure 20. Percents Fe_o and Al_o as a function of depth for five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.

<u>Fe_p and Al_p.</u> Although pyrophosphate-extractable Fe (Fe_p) and Al (Al_p) concentrations in these soils are lower than the corresponding CBD- or oxalate-extractable concentrations, the shapes of the depth functions are similar (Figure 21). The Fe_p and Al_p maxima increase in magnitude with surface age, although, as before, the strongest Al_p maximum is found in the Bs2 subhorizon of the Battlefield pedon. As with the other extractants, the C horizon concentration of Al_p was slightly greater in the Main Algonquin pedon (0.03%) than in the Battlefield pedon (0.02%) or the Algoma and Nipissing pedons (0.01%).

<u>General Fe and Al patterns.</u> The shapes of the depth function curves for Fe and Al, for all three extractants, indicate that these elements have been translocated from the upper horizons and deposited in the B horizon. B horizon maxima and B-E differences are generally lacking in the Algoma pedons, but increase in strength with surface age for the remaining three surfaces.

Although the depth function patterns for extractable Al are similar to those for Fe, the Al maxima occur slightly deeper than the Fe maxima (Figures 19 - 21). The Fe maxima are found in the uppermost B subhorizon on the three oldest surfaces (Nipissing Bs1, Battlefield Bs1, and Main Algonquin Bhs), while the Al maxima occur in the second B subhorizon (Nipissing Bs2, Battlefield Bs2, and Main Algonquin Bs1). Al maxima deeper than Fe maxima have been reported in Spodosols of the northern Appalachians (DeKimpe and Martel 1976), Japan (Mizota 1982), Finland (Koutaniemi *et al.* 1988), and Quebec (Kodama and Wang 1989), as well as in northern Michigan (Franzmeier and Whiteside 1963b). This trend may occur because Al is more mobile than Fe, and thus initially deposited at greater depth (Mizota 1982), or because the Al is remobilized after deposition and moved farther down in the profile (DeKimpe and Martel 1976).

The phenomenon of deeper Al deposition within the profile may also account for the slight increase in C horizon Al content with surface age (Koutaniemi 1988): small amounts of the relatively colorless Al may be translocated deeper than is Fe, forming a horizon, slightly enriched in Al, that visually appears to be C horizon material. In sandy Spodosols, laboratory criteria may consistently place the B-C horizon boundary deeper than color and other field-based morphologic attributes would indicate (Schaetzl 1987). Thus the field-identified C horizon could actually exhibit small accumulations of Al.



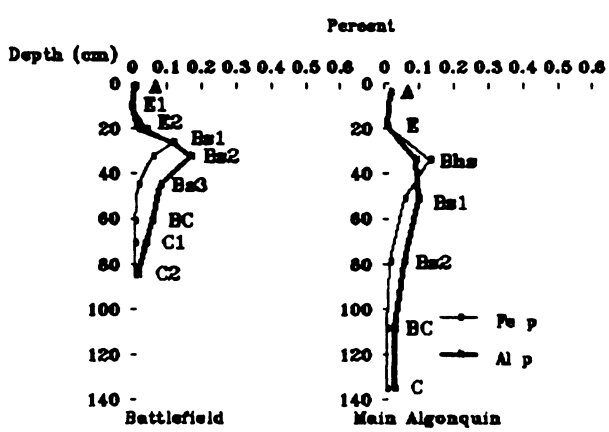


Figure 21. Percents Fe_p and Al_p as a function of depth in five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.

Comparisons between the three extractants. The three Fe and Al extractants employed affect different forms of these elements (Table 17). Due to these differences, various combinations and ratios of extractable Fe and Al data can be used to distinguish the forms of Fe and Al affected by pedogenesis.

Fe_d concentrations in the five pedons are higher than either Fe_o or Fe_p concentrations; Al_d concentrations are higher than Al_p, but lower than Al_o concentrations (Figures 19 - 21). The ratios of inorganic amorphous Fe to organic amorphous Fe $[(Fe_o - Fe_p) / Fe_p]$ and of inorganic amorphous Al to organic amorphous Al $[(Al_o - Al_p) / Al_p]$ have been used by other researchers to examine Fe and Al distributions in Spodosols (Evans and Wilson 1985; Wang *et al.* 1986), but have been shown to vary widely between Spodosols developed under different environments. High ratios (> 3) in the B horizons of some Spodosols were associated with the presence of pedogenic imogolite in these horizons, while low ratios (< 1) were associated with its absence (Wang *et al.* 1986). In the present study, these ratios are generally < 1 in the eluvial horizons, and increase with depth (Figure 22). Although the values of these ratios are not as large as those reported by Wang *et al.* (1986), a substantial proportion of the Fe and Al in the B horizons of these pedons is not organically bound. A consistent pattern with surface age is lacking, however. Although it was not the intent of this study to search for imogolite, the high proportion of inorganic amorphous Fe and Al in the Spodosol pedons would suggest that pedogenic imogolite may be present and that the imogolite content of these soils should be evaluated.

Because imogolite and proto-imogolite allophanes involved in the translocation process must be an amorphous inorganic Al compound, one method of estimating the imogolite content of a soil is the difference between Al_o and Al_p (Figure 23). The distribution of the amorphous inorganic Al with depth in the Algoma pedons reveals that little imogolite has formed in these pedons. In the pedons on the older surfaces, however, more imogolite is present, and its distribution with depth is typical of translocated materials. The accumulation of imogolite in the B horizon, as for all extractable Fe and Al, is greatest in the Battlefield pedon.

The ratio Fe_o/Fe_d ("activity ratio") has been used as a relative measure of the crystallinity of free iron oxides (Blume and Schwertmann 1969); a low Fe_o/Fe_d ratio would indicate a large proportion of crystalline Fe oxide compounds, and, therefore, low Fe mobility in the soil (Singer *et al.* 1978). The Nipissing, Battlefield, and Main Algonquin pedons have Fe_o/Fe_d ratio depth functions quite typical of Spodosol-like soils (Figure 24) (Blume and Schwertmann 1969), with pronounced maxima in the Bs horizons. The magnitude of this maximum, which is absent in the Algoma pedons, increases with surface age. Fe_o/Fe_d values for the pedons of the present study fall within a range of 0.2 - 0.8. This compares favorably with values of 0.3 - 1.19 (mean 0.87) in Humo-Ferric Podzols of northern Ontario (Evans and Wilson 1985) and a value of 1.0 for a well-drained Spodosol Bh horizon in Germany (Blume and Schwertmann 1969). Singer *et al.* (1978) suggested that Fe_o/Fe_d values of $>$ 0.3 in any Spodosol horizon indicated the horizon was currently undergoing pedogenesis and that the Fe and Al were mobile, while a ratio of 0.1 revealed less pedogenetic activity. Low Fe_o/Fe_d values may indicate that little of the extractable Fe is in the form of crystalline oxyhydroxides (Evans and Wilson 1985), while the higher values encountered in the present study would signify that a substantial proportion of the Fe present is amorphous in form, especially in the illuvial horizons.

In the present study the shape of the Al_o/Al_d ratio depth functions mirror those of Fe_o/Fe_d, although the values of the ratios are much higher, due to high Al_o values (Figure 25). The proper interpretation for the Al_o/Al_d ratio is unclear, but one possible interpretation would be analogous to that for Fe_o/Fe_d ratios. Al_o and Al_d values were found to be highly correlated in Ontario Spodosols, suggesting that the two extractants dissolved similar materials (Evans and Wilson 1985).

Table 17. Forms of Fe and Al affected by three extractants.

Extractant	Forms of Fe and Al			Reference
	Crystalline*	Amorphous		
		Organic bound	Inorganic bound	
CBD	x	x	x	McKeague and Day 1966; McKeague 1978
Oxalate		x	x	McKeague and Day 1966; Singer and Janitzky 1986
Pyrophosphate		x		McKeague 1978

*Includes crystalline primary Fe oxides (e.g., finely-divided hematite and goethite), but not Fe and Al in silicate materials (McKeague 1978).

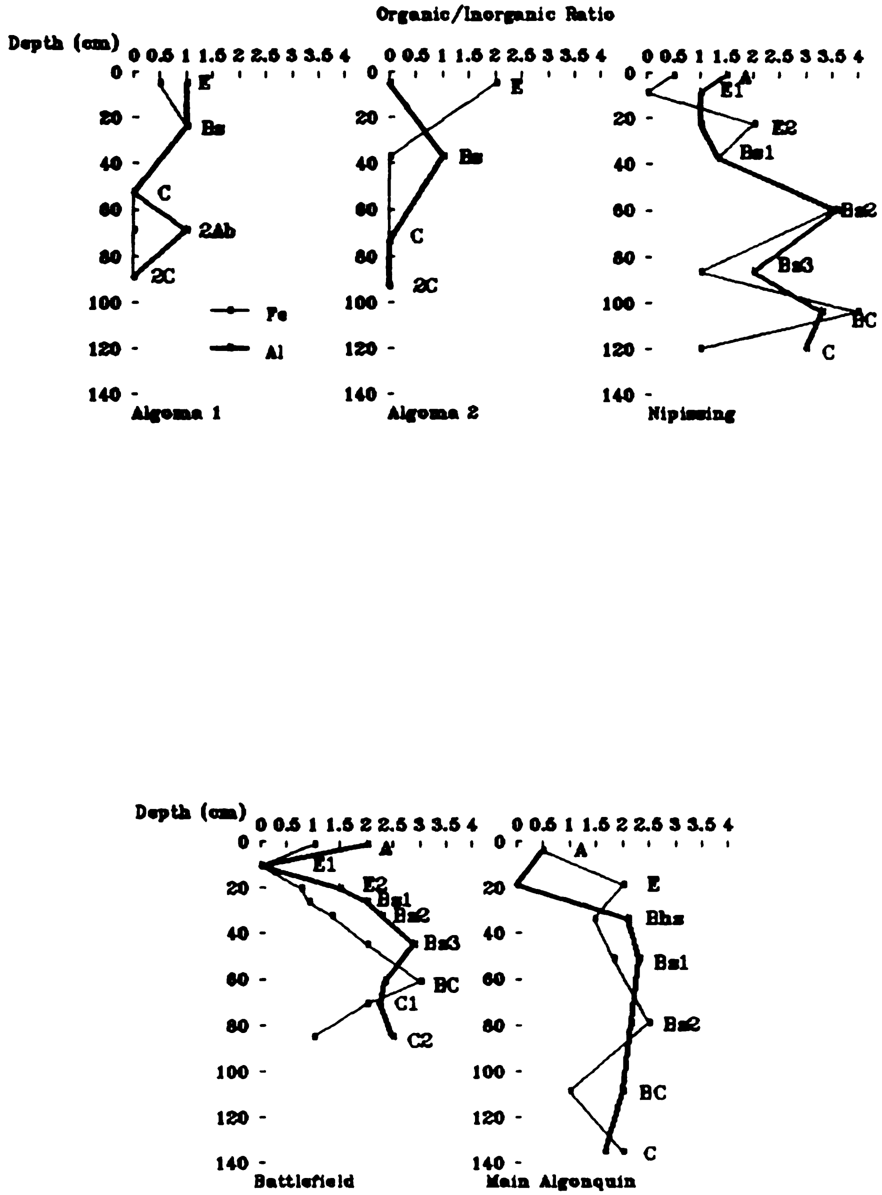


Figure 22. Ratios of inorganic amorphous Fe to organic amorphous Fe [$(Fe_o - Fe_p) / Fe_p$] and inorganic amorphous Al to organic amorphous Al [$(Al_o - Al_p) / Al_p$] as a function of depth in five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.

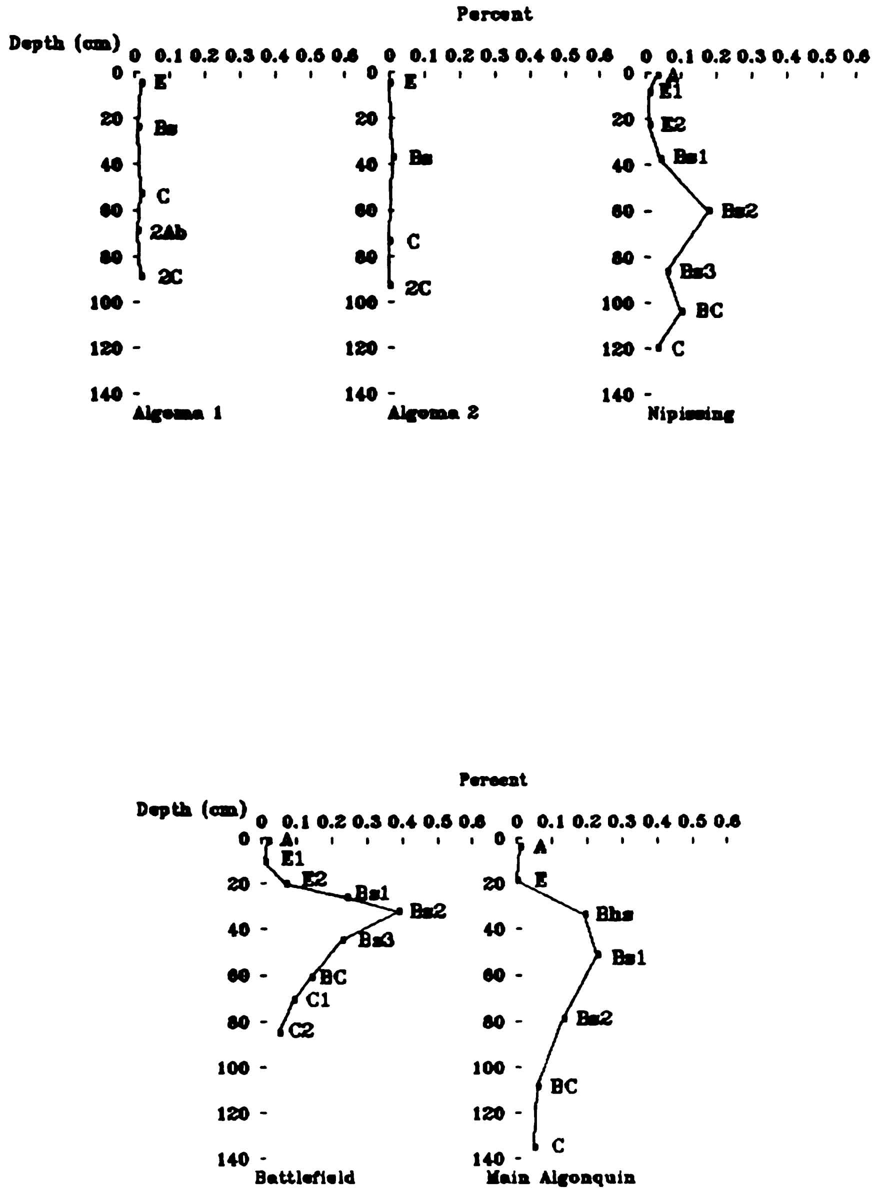


Figure 23. Inorganic amorphous Al ($Al_o - Al_p$) as a function of depth in five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.

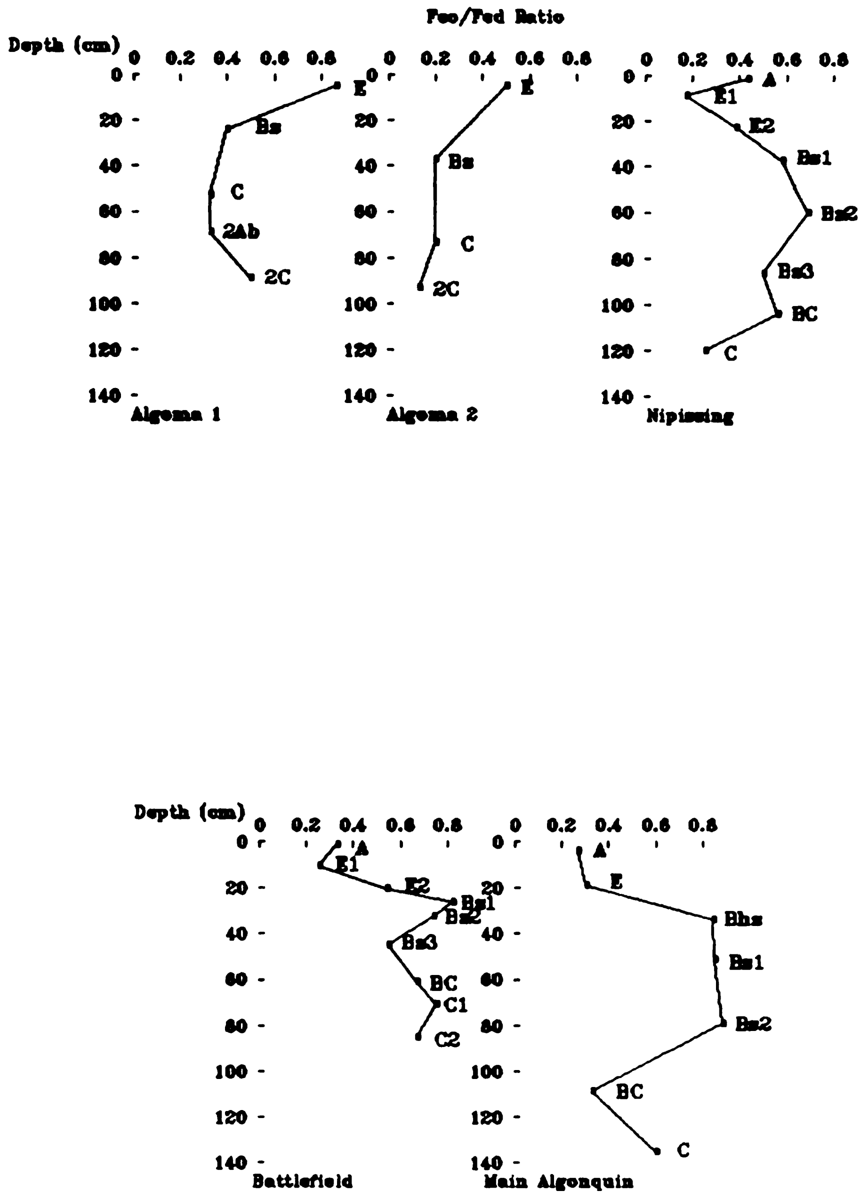


Figure 24. Fe_o/Fe_d ratios as a function of depth in five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.

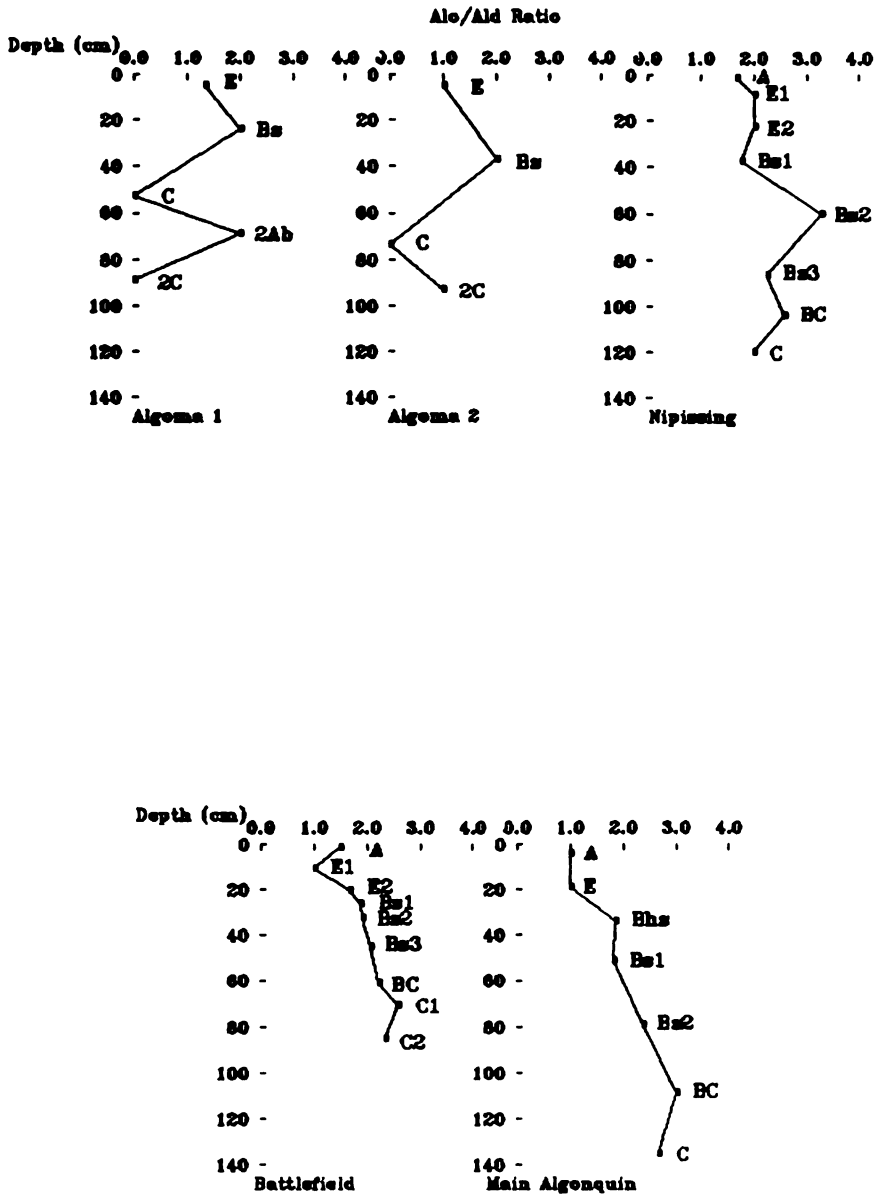
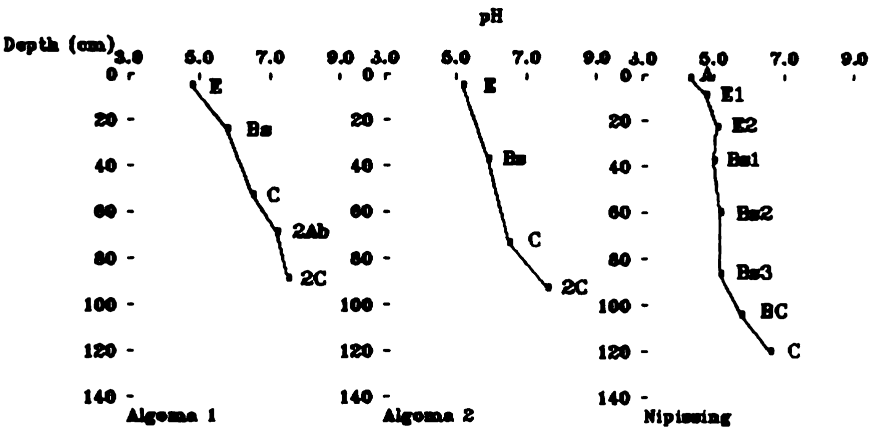


Figure 25. Al_o/Al_d ratios as a function of depth in five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.

pH

In the pedons examined, pH increases with depth (Figure 26). The Algoma pedons are less acidic than the other soils, but in general acidity is unrelated to surface age, either for the surface horizons or for the parent material. Franzmeier and Whiteside (1963b), similarly, found that pH increased with depth in northern Michigan, but with no pattern with respect to soil age. A comparable result was obtained for Spodosols of northwest Finland (Jauhiainen 1973). In contrast, Nørnberg (1977) reported that surface pH generally decreased with increasing soil age in uncultivated Spodosols of Denmark. In Finland, B horizon pH values were weakly correlated with soil age, whereas A horizon pH values decreased with soil age (Koutaniemi *et al.* 1988).



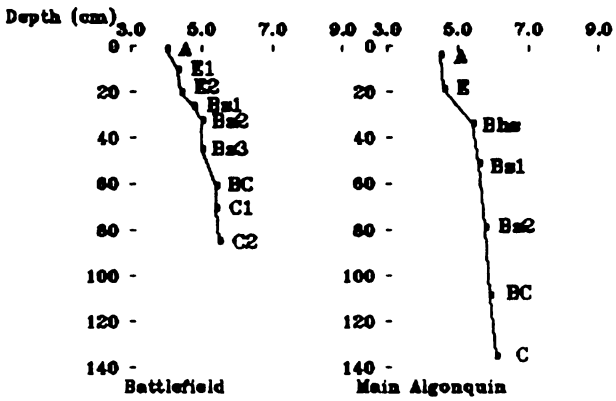


Figure 26. pH (2:1 H_2O) as a function of depth in five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.

Chronofunctions

Linear regression techniques applied to chronosequence data may quantify the changes in a particular soil property or properties in the sequence over time (Bockheim 1980). When the sample size is small and few surfaces are available for study, the chronofunctions may be difficult to interpret. The present study, involving only five pedons on four surfaces, is not ideal in this regard; therefore, caution should be exercised in generalizing the results presented below.

Chronofunctions have been developed for the five pedons using three types of data: (1) weighted profile means[9] of a soil property ; (2) weighted B horizon means[10] of a soil property; and (3) indices of soil development. Three regression models were attempted: (1) linear ($Y = a + bX$); (2) single-logarithmic [$Y = a + (b \ln X)$]; and (3) logarithmic [$\ln Y = a + (b \ln X)$], following Bockheim (1980). Other models were not attempted, even though some researchers have found them to be superior in some cases (e.g., Sondheim *et al.* 1981; James 1988). The small sample size in the present study does not provide enough data points to realistically evaluate the statistical fit of nonlinear models.

Chronofunctions using weighted profile means

One method of comparing soil properties between pedons is to use the mean value for all horizons in a profile, weighted for (multiplied by) horizon thicknesses in cm (Table 18). Linear regression analyses reveal

[9] Weighted profile means calculated according to the following formula: $\Sigma (P * H) / \Sigma H$, where P = soil property for a given horizon and H = horizon thickness in cm.

[10] Weighted B horizon means were calculated analogously to weighted profile means.

a positive correlation between OC and most extractable Fe and Al properties with surface age; however, neither pH nor ODOE showed a statistically significant relationship ($p < 0.05$) to surface age (Table 19). A graph for Fe_p, shown as an example of a significant regression equation, is presented in Figure 27. Of the three models used, the single logarithmic model appears to provide the best results, in terms of significance level and the amount of variability explained (adj. r^2). Nevertheless, the results from all models attempted are equally acceptable, so that no one model is clearly superior.

Table 18. Weighted profile and B horizon means for Fe, Al, OC, ODOE, and pH, for each surface.

Soil Property	Algoma-1	Algoma-2	Nipissing	Battlefield	Main Algonquin
Weighted profile mean:					
% Fe_d	0.049	0.057	0.095	0.086	0.115
% Al_d	0.006	0.008	0.038	0.087	0.071
% Fe_p	0.010	0.008	0.016	0.023	0.028
% Al_p	0.001	0.010	0.027	0.051	0.049
% Fe_o	0.026	0.012	0.049	0.055	0.078
% Al_o	0.022	0.015	0.101	0.177	0.147
% OC	0.12	0.13	0.14	0.36	0.36
ODOE	0.041	0.012	0.019	0.041	0.052
pH	6.52	6.29	5.36	5.44	5.55
% Al_o-% Al_p	0.017	0.005	0.073	0.138	0.099
Weighted B horizon mean:					
% Fe_d	0.05	0.05	0.11	0.16	0.17
% Al_d	0.01	0.01	0.05	0.18	0.13
% Fe_p	0.01	0.01	0.02	0.05	0.05
% Al_p	0.01	0.01	0.04	0.10	0.08
% Fe_o	0.02	0.01	0.07	0.11	0.14
% Al_o	0.02	0.02	0.15	0.36	0.26
% OC	0.1	0.1	0.1	0.6	0.4
ODOE	0.018	0.014	0.020	0.079	0.095
% Al_o-% Al_p	0.010	0.010	0.109	0.257	0.180

Table 19. Regression equation data for plots of weighted profile properties (Y) versus surface age (X).

Property	Model[a]	a	b	adj. r^2
% OC	1	0.026	0.000032	0.98***
% OC	2	-1.44	0.19	0.97**
% OC	3	-9.04	0.86	0.98***
ODOE	1	0.014	0.0000031	0.36
ODOE	2	-1.12	0.018	0.31
ODOE	3	-9.13	0.65	0.26
pH	1	6.37	-0.000087	0.22
pH	2	10.85	-0.59	0.33
pH	3	2.60	-0.10	0.32
% Fe_d	1	0.047	0.0000053	0.45
% Fe_d	2	-0.21	0.034	0.51
% Fe_d	3	-6.36	0.44	0.51
% Al_d	1	-0.011	0.0000086	0.84*
% Al_d	2	-0.42	0.054	0.90**
% Al_d	3	-18.71	1.76	0.75*
% Fe_p	1	0.0024	0.0000021	0.90**
% Fe_p	2	-0.093	0.013	0.92**
% Fe_p	3	-10.85	0.78	0.88*
% Al_p	1	-0.0055	0.0000053	0.83*
% Al_p	2	-0.20	0.034	0.89**
% Al_p	3	-23.05	2.18	0.38
% Fe_o	1	0.0092	0.0000056	0.67
% Fe_o	2	-0.26	0.035	0.71*
% Fe_o	3	-11.29	0.93	0.55
% Al_o	1	-0.01	0.000017	0.75*
% Al_o	2	-0.81	0.11	0.82*
% Al_o	3	-15.96	1.54	0.68
% Al_o-% Al_p	1	-0.095	0.00012	0.67
% Al_o-% Al_p	2	-0.61	0.079	0.76*
% Al_o-% Al_p	3	-18.44	1.77	0.56

[a] Regression models:
1. (Linear) $Y = a + bX$;
2. (Single logarithmic) $Y = a + (b \ln X)$;
3. (Logarithmic) $\ln Y = a + (b \ln X)$.

* significant at 0.05 level
** significant at 0.01 level
*** significant at 0.001 level

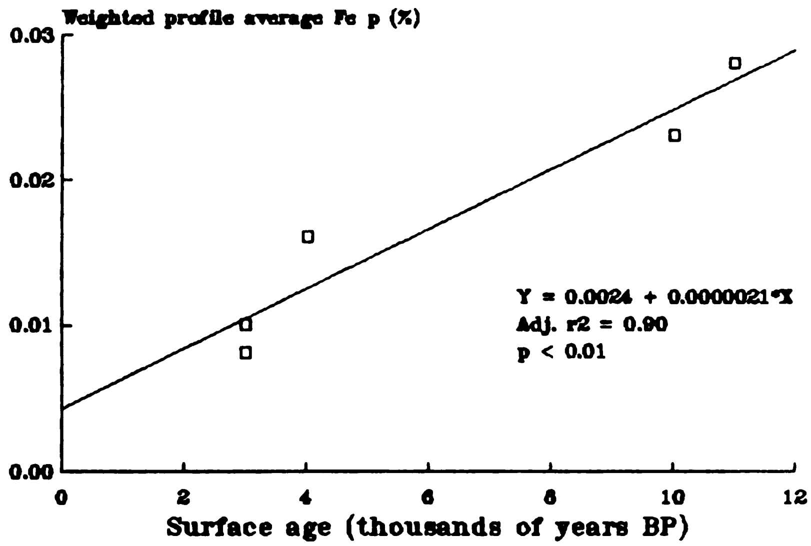


Figure 27. Weighted profile average Fe_p as a function of surface age.

Chronofunctions using weighted B horizon means

Weighted B horizon means[11] are used to compare the differences in B horizon content of the various soil chemical properties between pedons. Because depth functions (Figures 17 - 21) reveal that the concentrations of Fe, Al, and OC data change more predictably in the B horizon than in other horizons, the weighted B horizon mean should more closely reflect changes in soil development than the weighted profile mean. Weighted profile means may characterize changes in soil OC content more effectively than weighted B horizon means, however, because the strongest OC maximum in a profile generally occurs at the surface, rather than in the B horizon. Additionally, sampling of surface OC content is complicated by the presence of O horizons above the A horizons, which may distort the data for OC.

Weighted B horizon means, when used in linear regression models with surface age, yield significant equations ($p < 0.05$) for all Fe, Al, and OC properties studied and for all models attempted (Table 18). Fe_p shows an especially good fit in these models (Figure 28). All models provide acceptable r^2 values for all properties investigated, and no one model consistently gives the highest r^2 values (Table 20).

Levels of statistical significance and r^2 values for equations derived from weighted B horizon means were higher than for the corresponding equations derived from weighted profile means, except in the case of OC. The weighted profile mean for OC, although influenced by B horizon OC content, is probably most sensitive to surface OC concentrations, as explained above.

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[11] Weighted B horizon means are weighted means of all B subhorizons, not including BC horizons.

Table 20. Regression equation data for plots of weighted B horizon properties (Y) versus surface age (X).

Property	Model[a]	a	b	adj. r^2
% OC	1	-0.070	0.000053	0.78*
% OC	2	-2.54	0.33	0.79*
% OC	3	-12.92	1.31	0.90**
ODOE	1	-0.015	0.0000097	0.99***
ODOE	2	-0.46	0.059	0.96**
ODOE	3	-15.25	1.38	0.98***
% Fe_d	1	0.023	0.000014	0.85*
% Fe_d	2	-0.63	0.087	0.91**
% Fe_d	3	-9.80	0.87	0.82*
% Al_d	1	-0.04	0.000018	0.85*
% Al_d	2	-0.89	0.110	0.89*
% Al_d	3	-20.36	2.00	0.85*
% Fe_p	1	-0.0038	0.00000517	0.98***
% Fe_p	2	-0.24	0.032	0.99***
% Fe_p	3	-14.36	1.23	0.96**
% Al_p	1	-0.012	0.0000097	0.86*
% Al_p	2	-0.48	0.061	0.91**
% Al_p	3	-17.12	1.60	0.82*
% Fe_o	1	-0.013	0.000013	0.86*
% Fe_o	2	-0.65	0.084	0.90**
% Fe_o	3	-16.42	1.56	0.70*
% Al_o	1	-0.051	0.000034	0.77*
% Al_o	2	-1.70	0.22	0.85*
% Al_o	3	-19.02	1.94	0.73*
% Al_o-% Al_p	1	-0.039	0.00024	0.75*
% Al_o-% Al_p	2	-1.22	0.16	0.82*
% Al_o-% Al_p	3	-21.41	2.16	0.69*

[a] Regression models:
1. (Linear) Y = a + bX;
2. (Single logarithmic) Y = a + (b ln X);
3. (Logarithmic) ln Y = a + (b ln X).

* significant at 0.05 level
** significant at 0.01 level
*** significant at 0.001 level

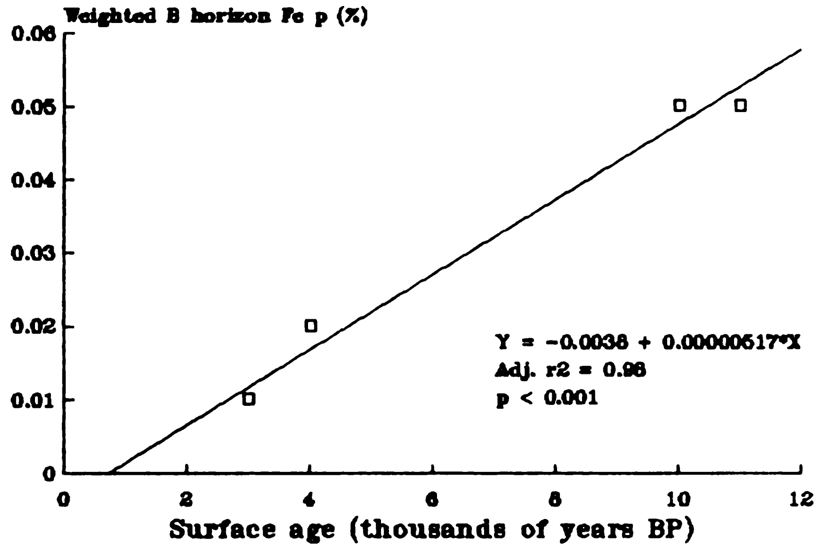

Figure 28. Weighted B horizon average Fe_p as a function of surface age.

Chronofunctions using indices of soil development

The pedons studied were also evaluated on five separate indices of soil development: (1) Walker and Green's (1976) Index of Profile Anisotropy (IPA); (2) Birkeland's (1984a) modification of the IPA (mIPA); (3) Duchaufour and Souchier's (1978) index of podzolization (K_{Al}); (4) the Harden (1982) Soil Development Index; and (5) Schaetzl and Mokma's (1988) POD Index. Of these, the IPA, mIPA, and Harden Index calculate separate values for various soil properties; the Harden Index provides a separate profile index as well. The K_{Al} is calculated from Al data only, although the particular Al extractant to be utilized for this index has not been specified (Duchaufour and Souchier 1978); it was calculated herein for both Al_d and Al_o. The various index values computed for the five pedons are listed in Table 21. The specific meaning and implications of the data in Table 21 will be examined below.

Table 21. Index values for five pedons: Algoma-1, Algoma-2, Nipissing, Battlefield, and Main Algonquin.

Property	Algoma-1	Algoma-2	Nipissing	Battlefield	Main Algonquin
Walker-Green Index of Profile Anisotropy (IPA):					
Fe_d	0.028	0.010	0.021	0.063	0.075
Al_d	0.009	0.003	0.038	0.074	0.055
Fe_p	0.020	0.003	0.008	0.025	0.029
Al_p	0.008	0.000	0.011	0.044	0.030
Fe_o	0.028	0.006	0.022	0.048	0.078
Al_o	0.005	0.005	0.059	0.147	0.105
pH	0.823	0.743	0.526	0.598	0.462
OC	0.185	0.100	0.282	0.368	0.437
ODOE	0.035	0.005	0.009	0.040	0.059
Modified Index of Profile Anisotropy (mIPA):					
Fe_d	0.028	0.010	0.016	0.074	0.090
Al_d	0.010	0.008	0.019	0.072	0.049
Fe_p	0.018	0.008	0.008	0.023	0.027
Al_p	0.008	0.000	0.014	0.044	0.027
Fe_o	0.026	0.005	0.025	0.051	0.069
Al_o	0.004	0.003	0.051	0.148	0.099
OC	0.180	0.100	0.338	0.444	0.486
ODOE	0.043	0.008	0.016	0.045	0.057
Duchaufour-Souchier (K_{Al}):					
Al_d	1	1	2.3	9.7	6
Al_o	1	1	4.6	11.2	8.3
Harden Index:					
Consistence	0	0	3	8	8
Melanization	14	37	33	25	50
pH	14	17	59	54	52
Rubification	2	3	30	18	37
Structure	0	0	2	5	5
Profile	30	57	127	110	152
POD Index:					
POD Index	0	0	0	2	7

<u>IPA and mIPA.</u> The IPA and mIPA were developed to measure the degree of profile anisotropy with regard to particular soil properties (Walker and Green 1976; Birkeland 1984a).[12] In the present study, both pedons Algoma-1 and Algoma-2 contain lithologic discontinuities below the first C horizon; in most cases the effect on the IPA and mIPA values appears to be minimal, and the results were included in the regression analysis.

For the pedons in the present study, the index values for the IPA and mIPA are similar (Table 21), and, when regressed against surface age, yield similar equations (Table 22). Both the IPA and mIPA values for Fe_d, Al_d, Fe_o, and Al_o resulted in statistically significant equations, as did the IPA of Al_p and OC, suggesting that profile anisotropy (for these soil properties) increases with soil age. The regression of IPA for Fe_d exhibits a significant linear relationship with surface age, having the highest r^2 value of those examined (Figure 29).

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[12]Theoretically, at time = 0 a profile should be isotropic (Jenny 1941), and anisotropy and the indices should increase with time. The IPA measures deviation of a property from the weighted mean value for the profile, while the mIPA measures deviation of a property from the presumed parent material (here interpreted to be the C horizon) value of that property. The two indices are best applied to soils that have developed in a single parent material, as the presence of a lithologic discontinuity means that the soil was anisotropic even at time = 0. The mIPA and IPA may be used, though with caution, when the second parent material occurs in the C horizon only. For the mIPA, finding a suitable parent material value may be difficult if no C horizon in the first parent material is present, although a value can be derived by comparison with neighboring soils that have a (1)C horizon (Birkeland 1984a). IPA values can also be affected by the presence of a 2C horizon, due to skewing of the weighted profile mean used in its calculation. Interpretation of IPA and mIPA values for pedons containing a lithologic discontinuity may present difficulties, as the values will be artificially inflated.

K_{Al}. Duchaufour and Souchier's (1978) index of podzolization was developed to measure the degree of profile anisotropy in podzol soils using the plot of Al concentration against depth. When the K_{Al} values for the pedons in the present study are regressed against surface age, the resulting equations are statistically significant (Table 22; Figure 30), suggesting that the profiles become increasingly anisotropic with age, agreeing with the results from the IPA and mIPA indices, above.

Harden Index. The Harden Index was originally developed for use with soils in California (Harden 1982), although it has been applied to soils developed in a variety of climates (Harden and Taylor 1983; Birkeland 1984a). Nevertheless, this index is difficult to use with the soils of the present study, as it does not measure the critical diagnostic and pedogenic properties of these soils (Schaetzl and Mokma 1988). Only five of the eight soil properties employed in the Harden index are applicable to these soils: consistence, structure, melanization, pH, and rubification. Of these, only two, consistence and structure, yield statistically significant regression equations (Table 22). The consistence and structure values in this case probably reflect increasing cementation (ortstein) in the soils. For sandy soils such as these, structure and consistence are relatively unimportant properties, difficult to measure and conveying little information about the degree of soil development. Although the composite (profile) Harden Index values generally increase with surface age, the correlation is not statistically significant (Figure 31).

POD Index. The POD Index is an index of soil development based on morphological characteristics (Schaetzl and Mokma 1988). Because it was developed for soils similar to those of the present study, it provided a superior measure of soil development for these soils. The plot of this function, however, suggests that the POD Index values[13] increase non-linearly with time (Figure 32), which is why the equation resulting from the regression of POD Index values against time is not statistically significant (Table 22).

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[13] POD Index values used in regression equations here are the values calculated from the profiles actually described and sampled in soil pits. The median POD Index values obtained for each surface (the POD Index values of the auger hole at the location of the soil pit) can also be used in a regression equation. Regression equations obtained with these values are as follows:

POD Index = -1.51 + 0.0063 * surface age ($p < 0.01$, $r^2 = .89$)
POD Index = -31.03 + 3.90 * ln surface age ($p < .01$, $r2 = 0.91$)
ln POD Index = -7.23 + 0.953 *ln surface age ($p > 0.5$, $r2 = .84$)

These POD Index values show better fit in the linear chronofunctions than the values derived from the sampled pit profiles.

Table 22. Regression equation data for plots of index values (Y) versus surface age (X).

Property	Model[a]	a	b	adj. r^2
Walker-Green Index of Profile Anisotropy (IPA):				
Fe_d	1	-0.0032	0.0000069	0.92**
Fe_d	2	-0.32	0.042	0.89*
Fe_d	3	-13.30	1.15	0.74*
Al_d	1	-0.0060	0.0000067	0.72*
Al_d	2	-0.33	0.043	0.80*
Al_d	3	-18.93	1.76	0.60
Fe_p	1	0.0029	0.0000023	0.55
Fe_p	2	-0.10	0.014	0.50
Fe_p	3	-13.25	1.04	0.32
Al_p	1	-0.0072	0.0000042	0.79*
Al_p	2	-0.20	0.026	0.83*
Al_p	3	-14.43	1.20	0.90*
Fe_o	1	-0.0031	0.0000064	0.77*
Fe_o	2	-0.29	0.038	0.73*
Fe_o	3	-14.09	1.22	0.55
Al_o	1	-0.025	0.000014	0.77*
Al_o	2	-0.71	0.091	0.85*
Al_o	3	-22.95	2.27	0.71*
pH	1	0.80	-0.000027	0.33
pH	2	2.11	-0.17	0.40
pH	3	1.86	-0.27	0.40
OC	1	0.08	0.000031	0.78*
OC	2	-1.41	0.20	0.82*
OC	3	-8.17	0.79	0.65
ODOE	1	0.0012	0.0000046	0.53
ODOE	2	-0.20	0.027	0.46
ODOE	3	-13.81	1.16	0.33

[a] Regression models used:
1. (Linear) $Y = a + bX$;
2. (Single logarithmic) $Y = a + (b \ln X)$;
3. (Logarithmic) $\ln Y = a + (b \ln X)$.

* significant at 0.05 level
** significant at 0.01 level
*** significant at 0.001 level

Table 22. (Cont'd.).

Property	Model[a]	a	b	adj. r^2
Modified Index of Profile Anisotropy (mIPA):				
Fe_d	1	-0.011	0.0000089	0.93**
Fe_d	2	-0.41	0.054	0.88*
Al_d	1	-0.88	0.0000065	0.81*
Al_d	2	-0.32	0.041	0.84*
Fe_p	1	0.0053	0.0000019	0.63
Fe_p	2	-0.08	0.011	0.55
Al_p	1	-0.0051	0.0000038	0.69
Al_p	2	-0.19	0.024	0.75*
Fe_o	1	-0.0014	0.0000059	0.83*
Fe_o	2	-0.27	0.036	0.83*
Al_o	1	-0.028	0.000014	0.77*
Al_o	2	-0.72	0.091	0.83*
pH	1	1.07	-0.000036	0.00
pH	2	2.47	-0.19	0.00
OC	1	0.075	0.000038	0.75
OC	2	-1.75	0.24	0.82*
ODOE	1	0.0095	0.0000039	0.41
ODOE	2	-0.16	0.023	0.35

[a] Regression models used:
1. (Linear) $Y = a + bX$;
2. (Single logarithmic) $Y = a + (b \ln X)$;
3. (Logarithmic) $\ln Y = a + (b \ln X)$.

* significant at 0.05 level
** significant at 0.01 level
*** significant at 0.001 level

Table 22. (Cont'd.).

Property	Model[a]	a	b	adj. r^2
Duchaufour-Souchier Index (K_{Al}):				
Al_d	1	-1.35	0.00086	0.75*
Al_d	2	-42.22	5.40	0.79*
Al_d	3	-12.26	1.55	0.91**
Al_o	1	-1.28	0.0010	0.80*
Al_o	2	-51.56	6.63	0.87*
Al_o	3	-12.78	1.63	0.78*
Harden Index:				
Consistence	1	-2.34	0.00099	0.93***
Consistence	2	-49.17	6.19	0.98**
Melanization	1	22.01	0.0016	0.00
Melanization	2	-49.62	9.51	0.00
pH	1	16.91	0.0036	0.24
pH	2	-167.60	24.15	0.35
Rubification	1	1.81	0.0026	0.24
Rubification	2	-126.95	16.92	0.31
Structure	1	-1.41	0.00061	0.92**
Structure	2	-30.53	3.85	0.97**
Profile	1	36.97	0.0094	0.39
Profile	2	-423.88	60.61	0.47
POD Index:				
POD Index	1	-2.27	0.00066	0.65
POD Index	2	-31.58	3.90	0.58
POD Index	3			

[a] Regression models used:
1. (Linear) $Y = a + bX$;
2. (Single logarithmic) $Y = a + (b \ln X)$;
3. (Logarithmic) $\ln Y = a + (b \ln X)$.

* significant at 0.05 level
** significant at 0.01 level
*** significant at 0.001 level

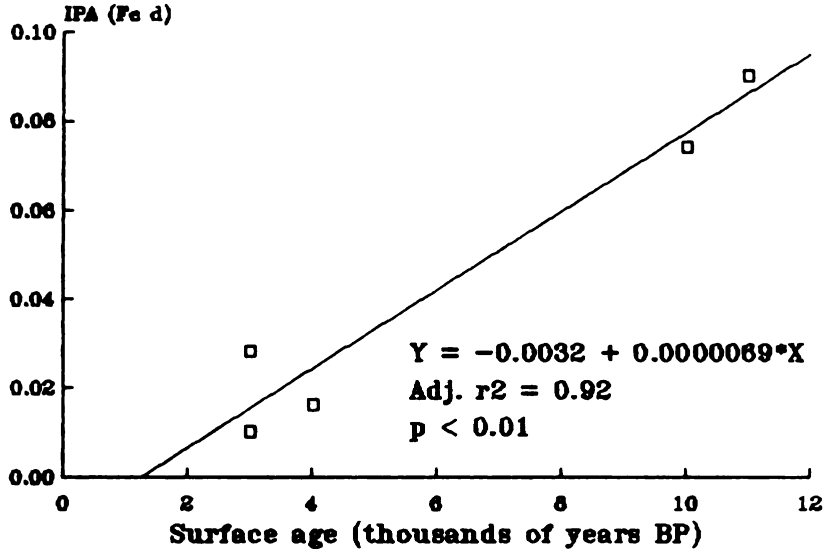

Figure 29. Index of Profile Anisotropy (IPA) for Fe_d as a function of surface age.

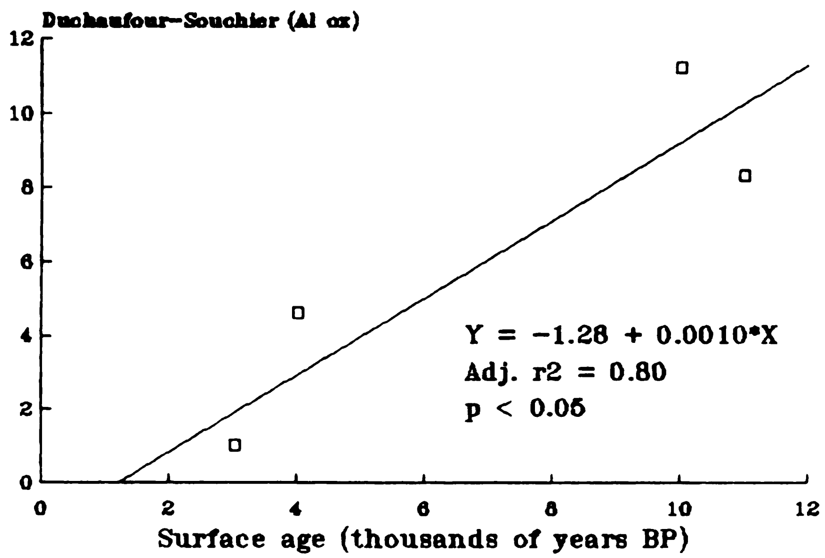

Figure 30. K_{Al} (Al_o) as a function of surface age.

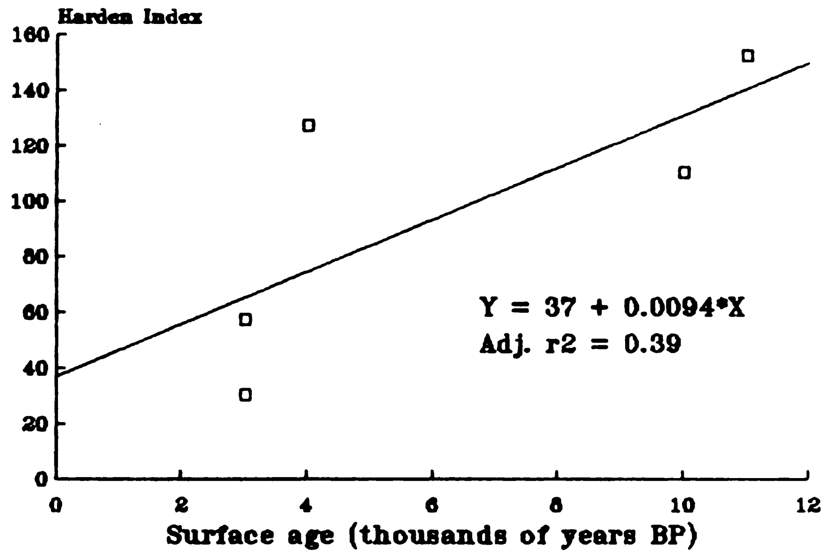

Figure 31. Harden profile index, calculated using 5 profile properties, as a function of surface age.

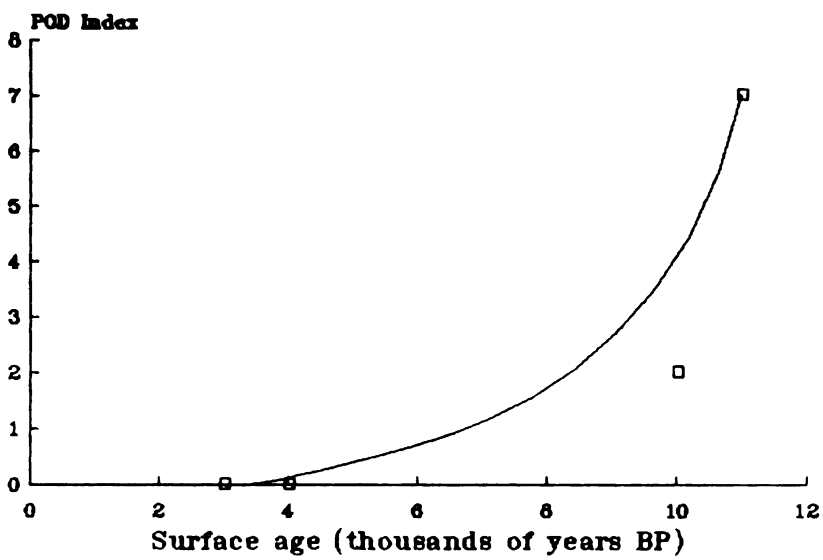

Figure 32. POD Index as a function of surface age.

Discussion of chronofunction results

The high proportion of statistically significant chronofunctions derived from weighted profile averages, weighted B horizon averages, and various soil development indices (Tables 19, 20, and 22) suggests that, in general, the soil properties examined do increase with surface age, and most can be adequately described with linear or logarithmic functions. The chronofunction model is not equally appropriate for all the soil properties examined, as no equations based on pH values are statistically significant, ODOE chronofunctions are significant only for weighted B horizon values, and POD Index appears to increase exponentially with time.

The correct interpretation of many specific chronofunctions is more difficult to discern. One aspect of concern with the linear and single logarithmic models attempted here is the large number of equations for which the y-intercept (a) is a negative value (Tables 19, 20, and 22): the implication of such a chronofunction is that at time = 0 the soil property in question would be expected to have a negative value. Clearly, a negative soil propery value is impossible; most probably the negative y-intercept value is an indication that soil development does not proceed in a strictly linear or logarithmic manner, and that extrapolation of these functions linearly backward to time = 0 is not valid. The line described by the regression equation may be an approximation of a more complex curve that is nearly linear over the range described by the present chronosequence (Richardson and Edmonds 1987). In some cases, as with many index values, the negative y-intercept value may imply that development of the property in questi

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question was delayed, or that the index value remained zero until the soil had developed for a period of time (e.g., Figure 29).

Of the three regression models attempted, the single logarithmic model ($Y = a + b \ln X$) appears to provide the most satisfactory, although not consistent, results. Bockheim (1980) similarly found that the single logarithmic model most consistently provided significant results and yielded the highest correlation coefficients with the wide variety of soils data used in his survey. Busacca (1987) found it to be the best model for describing many soil properties in a California chronosequence. The single logarithmic model provided the best representation of the relationship between time and certain properties of soils in Antarctica, but others were superior for different properties (Bockheim 1990).

Chronosequence studies that use chronofunctions have become increasingly common since the publication of Bockheim's (1980) study, but no single regression model has proven to be superior for all soil properties. Instead, most studies have found that more than one model may fit the data (e.g., Bockheim 1980; Muhs 1982; Harden 1982; Harden and Taylor 1983; Levine and Ciolkosz 1983; Birkeland 1984a; McFadden and Hendricks 1985; Busacca 1987; James 1988; Bockheim 1990). Such a conclusion is appropriate for the present study, as well.

SOIL CLASSIFICATION

Because the Soil Taxonomy criteria for the classification of Spodosols are currently undergoing revision, the pedons in this investigation have been classified both according to the current Soil Taxonomy criteria (Soil Survey Staff 1988) and the proposed revised criteria[1] (R. Rourke 1990, pers. comm.). The pedons in this chronosequence span a range of soil development in well drained sandy parent materials, and so their classifications should reflect a range of Spodosol and Spodosol-like categories.

Current Soil Taxonomy Criteria

A Spodosol is defined as a soil with a spodic or placic horizon (Soil Survey Staff 1988). Because placic horizons have not been noted in Michigan soils, the classification of these pedons as Spodosols depends on the presence of a spodic horizon. A spodic horizon is defined as a subsurface horizon, below 12.5 cm depth, that must either have a subhorizon > 2.5 cm thick that is continuously cemented (ortstein), or have a sandy particle size class with sand grains covered with cracked coatings, or meet certain chemical requirements (Soil Survey Staff 1988). No pedon of the current study had a continuously cemented subhorizon, nor were cracked coatings observed (Appendix 1),

[1] The proposed revised criteria have been used as they stand at the time of writing. It is likely, however, that the criteria will change further before they are finalized.

so chemical criteria have been used to determine the presence of a spodic horizon (Table 23). One of these criteria, the index of accumulation, requires determination of CEC, which was not calculated for these pedons, but due to the low (≤ 0.5 %) clay content and relatively thick B horizons, it is assumed that any of the pedons meeting the Fe and Al criteria would meet this criterion as well. In a study of similar northern Michigan soils, all soils meeting the sesquioxide and OC criteria also met the index of accumulation criterion (Padley *et al.* 1985). Therefore, it is likely that the CEC of the present pedons would satisfy the index of accumulation requirements if the Fe, Al, and OC requirements are also met. According to the Fe and Al criteria, the Battlefield and Main Algonquin pedons classify as Spodosols, but the Algoma pedons and the Nipissing pedon do not, and are thus classified as Entisols (Tables 23 and 24).

Table 23. Chemical criteria for identification of spodic horizons (Soil Survey Staff 1988).

Horizon	$(Fe_p + Al_p)$/clay	$(Fe_p + Al_p)/(Fe_d + Al_d)$
Algoma-1 (Bs)	.13	.33
Algoma-2 (Bs)	.31	.33
Nipissing (Bs2)	.35	.35
Battlefield (Bs1)	.51*	.51*
Main Algonquin (Bs2)	.61*	.50*

* Exceeds requirement for spodic horizon.

Table 24. Classifications of five pedons according to Soil Taxonomy (Soil Survey Staff 1988) and under the proposed revision to Soil Taxonomy (R. Rourke 1990, pers. comm.).

Pedon	Soil Taxonomy	Proposed revision	Series name, Family
Algoma-1	Typic Udipsamment	Typic Udipsamment	Grayling, mixed*, frigid
Algoma-2	Typic Udipsamment	Typic Udipsamment	Grayling, mixed*, frigid
Nipissing	Spodic Udipsamment	Spodic Udipsamment	Eastport, mixed*, frigid
Battlefield	Entic Haplorthod	Entic Haplorthod	Rubicon, sandy, mixed*, frigid
Main Algonquin	Typic Haplorthod	Typic Haplorthod	Kalkaska, sandy, mixed*, frigid

* Assumed. Mineral families were not directly determined.

Proposed Soil Taxonomy

Under the proposed definition, spodic soil materials must meet a color requirement or have a laterally continuous horizon at least 2.5 cm thick that is cemented by Fe or Al and meets a color requirement. One of five morphologic or chemical requirements must also be met. Since none of the current pedons has a cemented horizon (Appendix 1), the color requirement must be met in order to classify as Spodosols; the Battlefield and Main Algonquin pedons meet this criterion, but the others do not (Table 25). The Nipissing Bs2 horizon, however, meets the hue and value requirements, but lacks the required chroma (chroma of 8 instead of "< 6.5"). Of the five morphologic or chemical criteria, no pedon has a B subhorizon > 20 cm thick with value and chroma < 3.5 or sand grains with cracked coatings (Appendix 1). The three remaining criteria, B to E (or A or Ap) horizon ratios of Fe_o, Al_o plus $Fe_o/2$, and ODOE, are met by at least one B subhorizon in the Nipissing, Battlefield, and Main Algonquin pedons (Table 25). Therefore, the Battlefield and Main Algonquin pedons easily meet spodic materials criteria, but the Nipissing pedon does not, because it does not meet the first color requirement.

The proposed revision of Soil Taxonomy also defines criteria for a spodic materials intergrade, in which the horizons in question must meet all of the requirements for spodic materials except for the lower boundary requirement, an andic requirement, or the color requirement. The spodic materials intergrade color requirement states that the colors must be 10YR 4/4, 3/4, or 5/4 with or without the presence of an albic horizon. The Algoma-1, Algoma-2, and Nipissing pedons all have B horizons that meet this requirement (Table 25), but only the Nipissing pedon qualifies as a spodic materials intergrade because the Algoma pedons do not meet the other requirements for spodic materials. Therefore, the classification of the five pedons under the proposed revision is as presented in Table 24, assuming that the great group and subgroup criteria within the Spodosol order do not change. Note that the final classifications under the proposed revision are identical to those under the current Soil Taxonomy system.

The Battlefield and Main Algonquin pedons exhibit the morphological characteristics of Spodosols, and the Algoma pedons clearly are very weakly developed soils, as indicated by their classifications under both systems. The Nipissing pedon, on the other hand, exhibits many of the characteristics of the Spodosols and evidently is developing towards a Spodosol. Under the proposed revised criteria, the Nipissing pedon meets most of the requirements necessary to classify as a Spodosol. It has a B subhorizon (Bs2) with a matrix color of 7.5YR 5/8 (Table 25), one unit of hue redder than the E horizon; the color value of 5 also meets the proposed criteria. This pedon also meets the chemical criteria for B horizon Fe and Al content, demonstrating the illuvial nature of the Fe and Al in the B horizon. Because the Nipissing pedon demonstrates most of the morphological and chemical requirements of a Spodosol, it may be desirable not to disqualify it on the basis of B horizon chroma, but rather to classify it as an Entic Haplorthod. The only modification to the revised criteria that would be necessary is the elimination of the chroma requirement in item 1(a)(1) (R. Rourke 1990, pers. comm.), such that it would read: "1. Have a dominant hue in a B subhorizon that is 7.5YR or redder, and a value of less than or equal to 5.5 in the matrix."

Table 25. Color and chemical criteria for the identification of spodic materials and spodic materials intergrades in the proposed revision to Soil Taxonomy (R. Rourke 1990, pers. comm.).

Horizon	Color	Fe_o ratio[a]	$Al_o + Fe_o/2$ ratio[b]	ODOE value[c]	ODOE ratio[d]
Algoma-1 (Bs)	10YR 5/4#	0.2	0.3	0.018	.1
Algoma-2 (Bs)	10YR 4/4#	0.3	1.0	0.014	.8
Nipissing (Bs2)	7.5YR 5/8	9.0*	11.0*	0.022	1.8
Nipissing (Bs3)	10YR 5/4#	4.0*	4.4*	0.015	1.3
Battlefield (Bs2)	7.5YR 3/4*	14.0*	42.0*	0.223	31.9*
Main Algonquin (Bhs)	7.5YR 3/2*	10.7*	17.6*	0.277*	23.1*

a: The ratio of Fe_o in the B subhorizon to the overlying E, A, or Ap horizon; must be > 2 to meet criterion.

b: The ratio of $Al_o + Fe_o/2$ in the B subhorizon to the overlying E, A, or Ap horizon; must be > 2 to meet criterion.

c: The ODOE value of the B subhorizon; must be greater than 0.25 to meet criterion.

d: The ratio of ODOE in the B subhorizon to the overlying E, A, or Ap horizon; must be > 2 to meet criterion.

Meets criterion for spodic materials intergrade.

* Meets criterion for spodic materials.

CONCLUSIONS

General

In the present thesis I have investigated soil development in sandy soils of northern Michigan. The four surfaces of the chronosequence under study ranged in age from 3000 to 11,000 BP, with the modal taxonomic class on these surfaces varying from Typic Udipsamments to Typic Haplorthods. The differences observed among the soils on the four surfaces may be related to the passage of time (soil age), assuming that any environmental changes (e.g., vegetation, climate) that took place had a negligible effect on the pathway of development.

Morphological soil properties that showed differences among members of the chronosequence included E and B horizon color differentiation, increasing amounts of ortstein in the B horizon, and increasing thickness of both the eluvial horizon and the total solum, with greater surface age. Many of these morphological differences may be associated with the translocation of Al, OC, and Fe from eluvial to illuvial horizons in the soil profile. Chemical analysis of soil horizons demonstrated that the profile distributions of OC and extractable Fe and Al with depth were indicative of translocation, and that increasing amounts of these substances had become translocated with soil age. Maximum B horizon accumulation of Al occurred deeper in the profile than the maxima of either Fe or OC. The majority of the Fe and Al translocated in these soils appeared to be amorphous (non-crystalline); a substantial proportion of the amorphous sesquioxides in the illuvial horizons may not have been organically-bound. Physical analysis of the pedons in this study revealed increased silt contents in the surface horizons.

Chronofunctions were developed using linear regression techniques for profile weighted mean properties, weighted mean B horizon properties, and several indices of soil development. A single logarithmic model (logarithmic transformation of soil property data only) appeared to provide the best results, but a simple linear model and a logarithmic model (logarithmic transformation of both soil property data and surface age) also provided satisfactory results. Negative y-intercept values for many of the chronofunctions (e.g., those for weighted profile Fe and Al content) indicate either that an alternative, non-linear model would be superior or that the specific soil property or index value did not develop (becoming greater than zero) until after a given amount of time had passed.

Morphological soil properties studied with field-acquired auger data revealed a substantial amount of variability among the soils on a given surface, with CVs ranging from 0.30 to 1.20 for the properties investigated. Variability within a surface decreased with time for depth to the B horizon and solum thickness, but increased with time for POD Index values. Increasing variability of soil development reflected in POD Index values may result from the increasing importance of micro-scale factors of soil development with time.

<u>Comparison to studies by Franzmeier and Whiteside</u>

The present study compares well with that of Franzmeier and Whiteside (1963a, b) and Franzmeier *et al.* (1963). Their study site was located near the shore of Lake Huron in Cheboygan County, approximately 50 km to the east of the present study site. The surfaces utilized in both the present study and the Franzmeier and Whiteside studies were beach deposits resulting from the same lake stages in the Lake Michigan/Huron basin.[1] Franzmeier and Whiteside, however, chose as a fourth[2] member of the chronosequence an upland moraine of Valders (Greatlakean) age. Most distinctions between the two studies can be traced to difficulties introduced with the inclusion of the Valders site, which was located inland and about 45 km distant from the first three sites, and on sandy, morainic parent material with slightly different soil textures than that of the first three. The soils at the Valders site exhibited clay bands (lamellae) below the spodic B horizon, leading Franzmeier and Whiteside to conclude that with increased time for soil development, clay band formation would occur after spodic horizon formation, even in the sandy textured parent materials of the lake terrace sites. It is likely, however, that the clay bands of the Valders pedon were due to a higher parent material clay content, rather than to soil age.

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[1] Discrepancies in dates between the two studies result from changes in accepted ages of lake stages between the date of the Franzmeier and Whiteside (1963) study and that of the present study (1991).

[2] Franzmeier and Whiteside identified only one Algonquin surface, resulting in only three lake terrace surfaces, whereas I subdivided the Algonquin surface into Battlefield and Main Algonquin surfaces.

Of the physical and chemical analyses performed in both studies, results in the present study closely matched those for the lake terrace (three youngest) sites of Franzmeier and Whiteside. In particular, the development of subsurface maxima of Fe_d, Al_d, and OC were very similar in the two studies, as were patterns of pH depth functions, increasing silt content of the surface horizon, and clay translocation into Bs/Bhs horizons.

Chronofunctions in the present study were calculated by means of linear regression analysis, but, in the Franzmeier and Whiteside study, only graphical methods were used. Although the different methods make comparison between the two studies difficult, both studies show increases in OC (organic matter in Franzmeier and Whiteside), Fe_d, and Al_d in the profile with time. Franzmeier and Whiteside, however, observed a maximum quantity of Fe and Al in the third member of their sequence, equivalent to the Battlefield or Main Algonquin pedons in the present study, and then a slight decrease in profile sesquioxide content at the Valders site. This discrepancy is likely due to the influence of a different parent material at the Valders site.

Franzmeier and Whiteside investigated a number of physical, chemical, and micromorphological characteristics of the soils not included in the present study. Investigation into variability of the soils within each surface and the change in soil variability with time is unique to the present study.

RECOMMENDATIONS FOR FURTHER RESEARCH

Several observations of the present study introduce questions which cannot be answered adequately because of the nature of the data obtained, but which could be investigated in future research. Although the importance of the translocation of Al, OC, and Fe to the development of Spodosols has been widely investigated, the role of clay translocation into Bs/Bhs horizons has been little studied. The present data suggest that some clay translocation may have taken place in these soils, as did the data of Franzmeier and Whiteside (1963b). Nevertheless, the amount of clay present in the parent materials is so low that it is difficult to conclude that clay has been translocated. Further research is necessary to confirm this conclusion and to investigate the possible role of clay in the formation of spodic horizons in sandy parent materials.

Both the present study and that of Franzmeier and Whiteside (1963b) identified increases in silt content in surface horizons with increasing surface age. The source of the silt-sized particles, although assumed to be due to physical weathering of larger particles, cannot be conclusively identified. Further research, similar to the work of Nørnberg (1980) in Denmark, could be undertaken to compare the mineralogy of the silt with that of the sands in the surface horizons and with possible sources of eolian input to these soils, in order to determine the source(s) of the silt in these horizons.

The presence of higher than anticipated amounts of inorganic, amorphous forms of Fe and Al in the Bs/Bhs horizons of the Spodosols of the present study suggests that the role of imogolite, proto-imogolite, and allophane in spodic horizon formation (Farmer *et al.* 1980; Anderson *et al.* 1982; Childs *et al.* 1983; Farmer 1982, 1984) should be more fully investigated for the soils of northern Michigan. Mineralogical studies are necessary to confirm or disprove the presence of imogolite in these soils, to determine the extent of its role in the translocation of Fe and Al in these soils, and to evaluate the role of organic matter in the translocation of sesquioxides. Although Wang *et al.* (1986) were able to identify imogolite in some Canadian soils, it was not present in all Spodosols. If both inorganic amorphous forms of Fe and Al as well as organically-bound sesquioxides are being translocated in the Spodosols of Michigan, a better understanding of the spatial extent and environmental conditions under which both processes operate could be provided by further research.

Although soils within a single surface vary spatially with respect to such properties as color, pH, texture, and organic matter content (e.g., Campbell 1979, Wilding and Drees 1983), the relationship between that variability and soil development has not often been investigated. In the present study, variability of certain soil properties appeared to decrease with surface age, while others (POD Index values) increased. Further research into the variability of other soil properties, and the relationship of that variability both to surface age, as well as to micro-scale factors of soil development (Alexander 1986), is necessary.

One difficulty involved in the investigation of soil spatial variability is the large amount of data necessary for an adequate description of soil variability. An understanding of variability is necessary not only to an understanding of soil development, but also to those who use soil in agriculture and engineering. Spatial variability of soils must also be represented accurately in soil maps and in GIS data bases. Therefore, new methods of gathering data on soil variability that are reliable yet less laborious and costly, as well as less destructive of the soils studied, are necessary.

APPENDIX A

Modal pedon descriptions

Algoma-1.

Location: 150 m E of Little Sucker Creek, 35 m S of road; SW 1/4 of NW 1/4 of SE 1/4, section 29, T39N, R5W, Emmet County, Michigan.
Date of description: October 1, 1989.
Series: Grayling.
Subgroup classification: Typic Udipsamment.
Parent material: Lacustrine sands.
Physiography: Lake terrace; elevation about 180 m.
Topography: About a 5% west-facing slope.
Drainage class: Well drained.

Oa 5-0 cm. Black (N2/0) well decomposed hardwood leaf litter; weak medium granular structure; very friable; many fine and medium roots; abrupt smooth boundary.

E 0-10 cm. Dark grayish brown (10YR 4/2) sand; single grain; loose; common fine roots; very strongly acid; clear wavy boundary.

Bs 10-38 cm. Yellowish brown (10YR 5/4) sand; single grain; loose; few fine roots; few fragments weakly cemented strong brown (7.5YR 5.6) ortstein (1 cm in diameter), primarily between 20 and 23 cm depth; medium acid; gradual smooth boundary.

C 38-68 cm Light yellowish brown (10YR 6/4) sand; single grain; loose; few fine roots; few medium distinct dark brown (10YR 4/3) mottles; slightly acid; abrupt smooth boundary.

Ab 68-70 cm. Very dark grayish brown (10YR 3/2) sand; single grain; loose; common fine roots; neutral; abrupt smooth boundary.

2C 70-108 cm. Light yellowish brown (10YR 6/4) sand; single grain; loose; mildly alkaline; abrupt wavy boundary.

3C 108+ cm. Boulders and organic matter.

<u>Algoma 2.</u>

Location: 20 m E of Little Sucker Creek, 70 m N of road; NW 1/4 of NW 1/4 of SE 1/4, section 29, T39N, R5W, Emmet County, Michigan.
Date of description: October 7, 1989.
Series: Grayling.
Subgroup classification: Typic Udipsamment.
Parent material: Lacustrine sands.
Physiography: Lake terrace; elevation about 180 m.
Topography: Nearly level.
Drainage class: Well drained.

Oe 5-0 cm. Very dark brown (10YR 2/2) partially decomposed hardwood and cedar leaf litter.

E 0-10 cm. Dark grayish brown (10YR 4/2) sand; single grain; loose; many fine and medium roots; strongly acid; clear wavy boundary.

Bs 10-64 cm. Dark yellowish brown (10YR 4/4) sand; single grain; loose; few fine distinct yellowish brown (10YR 5/6) mottles; medium acid; gradual wavy boundary.

C 64-83 cm. Light yellowish brown (10YR 6/4) sand; single grain; loose; slightly acid; abrupt smooth boundary.

2C 83+ cm. Brown (10YR 5/3) extremely gravelly sand; single grain; loose; mildly alkaline.

Nipissing.

Location: 125 m S of intersection of trail and Lakeshore Drive; NW 1/4 of SE 1/4 of SE 1/4, section 24, T38N, R6W.
Date of description: September 4, 1989.
Series: Eastport.
Subgroup classification: Spodic Udipsamment.
Parent material: Lacustrine sands.
Physiography: Lake terrace; elevation about 189 m.
Topography: Nearly level.
Drainage class: Well drained.

A — 0-3 cm. Black (N 2/0) sand; weak medium granular structure; very friable; many fine roots; extremely acid; abrupt smooth boundary.

E1 — 3-15 cm. Dark gray (10YR 4/1) sand; single grain; loose; common medium roots; very strongly acid; clear smooth boundary.

E2 — 15-30 cm. Brown (10YR 5/3) sand; single grain; loose; common fine roots; stongly acid; clear smooth boundary.

Bs1 — 30-45 cm. Yellowish brown (10YR 5/6) sand; single grain; loose; few fine roots; few small fragments weakly cemented strong brown (7.5YR 4/6) ortstein, 5-10 mm diameter; very strongly acid; clear irregular boundary.

Bs2 — 45-75 cm. Strong brown (7.5YR 5/8) sand; single grain; loose; few medium roots; few small fragments weakly cemented ortstein, 5-10 mm diameter; strongly acid; gradual smooth boundary.

Bs3 — 75-98 cm. Yellowish brown (10YR 5/4) sand; single grain; loose; few fine roots; few chunks yellowish brown (10YR 5/6) weakly cemented ortstein, 10-20 mm diameter; strongly acid; gradual smooth boundary.

BC — 98-110 cm. Yellowish brown (10YR 5/6) gravelly sand; single grain; loose; few fine roots; few chunks weakly cemented ortstein, 5-10 mm diameter; medium acid; gradual smooth boundary.

C — 110+ cm. Light yellowish brown (10YR 6/4) sand; few prominent medium dark brown (10YR 4/3) mottles; single grain; loose; neutral.

Battlefield.

Location: 25m E of 25m N of bend in road (where road leaves section line), NW 1/4 of SW 1/4 of NW 1/4 of section 31, T38N, R5W, Emmet County, Michigan.
Date of description: September 3, 1989.
Series: Rubicon.
Subgroup classification: Entic Haplorthod.
Parent material: Lacustrine sands.
Physiography: Lake terrace; elevation about 208 m.
Topography: About a 5% north-facing slope.
Drainage class: Well drained.

Oi 5-0 cm. Moderately decomposed hardwood litter; abrupt smooth boundary.

A 0-3 cm. Black (10YR 2/1) sand, flecked with light brownish gray (10YR 6/2); weak granular structure; very friable; common fine roots; extremely acid; abrupt wavy boundary.

E1 3-18 cm. Dark grayish brown (10YR 4/2) sand; single grain; loose; common roots; extremely acid; clear wavy boundary.

E2 18-23 cm. Brown (7.5YR 4/2) sand; single grain; loose; common fine roots; extremely acid; abrupt wavy boundary.

Bs1 23-30 cm. Strong brown (7.5YR 4/6) sand; single grain; loose; common roots; very strongly acid; clear irregular boundary.

Bs2 30-35 cm. Dark brown (7.5YR 3/4) sand; very weak subangular blocky structure; very friable; chunks of ortstein 10-20cm diameter, black (N 2/0) or black (5YR 2.5/1) with streaks of dark brown (7.5YR 4/4) and dark yellowish brown (10YR 4/6) sand; single grain; massive; weakly to strongly cemented; few roots; very strongly acid; clear broken boundary. Horizon is discontinuous.

Bs3 35-55 cm. Yellowish brown (10YR 5/6) sand; single grain; massive; weakly cemented (ortstein); few roots; very strongly acid; gradual irregular boundary.

BC 55-67 cm. Brownish yellow (10YR 6/6) sand; single grain; massive; weakly cemented; strongly acid; gradual wavy boundary.

C 67+ cm. Light yellowish brown (10YR 6/4) sand; single grain; loose; strongly acid.

Main Algonquin.
Location: 100 m N and 50 m E of corner of Larks Lake Road and unnamed road; SW1/4 of SW1/4 of SW1/4 of section 31, T38N, R5W, Emmet County, Michigan.
Date of description: September 30, 1989.
Series: Kalkaska.
Subgroup classification: Typic Haplorthod.
Parent material: Lacustrine sands.
Physiography: Lake terrace; elevation about 222 m.
Topography: Nearly level.
Drainage class: Well drained.

A 0-8cm. Very dark gray (N 3/1) sand; single grain; loose; many fine and medium roots; very strongly acid; abrupt wavy boundary.

E 8-30cm. Grayish brown (10YR 5/2) sand; single grain; loose; common medium and fine roots; very strongly acid; abrupt wavy boundary.

Bhs 30-38cm. Dark brown (7.5YR 3/2) sand; weak medium subangular blocky structure; very friable; common fine roots; chunks of strongly cemented ortstein, black (5YR 2.5/1) 15-30 cm diameter in lower part of horizon; strongly acid; clear irregular boundary.

Bs1 38-65 cm. Strong brown (7.5YR 4/6) sand; single grain; loose; few fine roots; chunks of weakly to strongly cemented ortstein 15-23 cm in diameter; medium acid; clear wavy boundary.

Bs2 65-93 cm. Dark yellowish brown (10YR 4/6) sand; single grain; loose; few fine roots; few fragments weakly cemented ortstein; medium acid; gradual wavy boundary.

BC 93-125 cm. Yellowish brown (10YR 5/4) sand; single grain; loose; medium acid; gradual smooth boundary.

C 125+ cm. Light yellowish brown (10YR 6/4) sand; single grain; loose; medium acid.

Location of auger sampling sites

Algoma surface

Block 1

AL1-1: 35 m W of Little Sucker Creek, about 5 m N of road, Wilderness State Park, NW1/4 of NW1/4 of SW1/4, section 29, T39N, R5W, Emmet County, Michigan.
AL1-2: 25 m N of AL1-1.
AL1-3: 25 m W, 15 m S of bridge over Little Sucker Creek in SW1/4 of NW1/4 of SW1/4 of section 29, T39N, R5W.
AL1-4: 25 m E of bridge, 25 m N of road, in NE1/4 of NW1/4 of SW1/4 of section 29, T39N, R5W.
AL1-5: 25 m N of AL1-4.
AL1-6: 50 m NE of AL1-5.
AL1-7: 30 m S of road and 50 m E of AL1-4.[1]
AL1-8: 50 m S of road and 50 m E AL1-7.
AL1-9: 50 m E of AL1-8.
AL1-10: 50 m E of road, farther south.

Nipissing surface

Block 2[2]

NI2-1: 50 m S of the first telephone pole SW of the intersection of the trail leading to the N shore of Wycamp Lake and Lake Shore Drive, NE1/4 of SW 1/4 of NE1/4, section 26, T38N, R6W, Emmet County, Michigan.
NI2-2: 50 m S of NI2-1.
NI2-3: 50 m S of NI2-2.
NI2-4: 50 m W of NI2-2.
NI2-5: 50 m S of NI2-4.
NI2-6: 50 m S of NI2-5.
NI2-7: 50 m W of NI2-5.
NI2-8: 50 m S of NI2-7.
NI2-9: 50 m S of NI2-8.

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[1] Later sampled as Algoma-1.

[2] Block 1 was abandoned because of possible imperfect drainage.

Block 4[3]

NI4-1: 25 m S of the intersection of the trail leading SE towards Wycamp Lake and Lake Shore Drive, NW1/4 of SE1/4 of SE1/4, section 24 T38N R6W.
NI4-2: 50 m S of NI4-1.
NI4-3: 50 m S of NI4-2.[4]
NI4-4: 50 m W of NI4-2.
NI4-5: 50 m S of NI4-4.
NI4-6: 50 m W of NI4-4.
NI4-7: 50 m S of NI4-6.
NI4-8: 50 m S of NI4-7.
NI4-9: 50 m W of NI4-7.

Battlefield surface

Block 1

BA1-1: 25m SW of bend in road, NW1/4 of SE1/4 of NE1/4 of section 36, T38N, R6W, Emmet County, Michigan.
BA1-2: 50 m SW of BA1-1.
BA1-3: 50 m SW of BA1-2.
BA1-4: 50 m SE of BA1-1.
BA1-5: 50 m SW of BA1-4.
BA1-6: 50 m SW of BA1-5.
BA1-7: 50 m SE of BA1-4.
BA1-8: 50 m SW of BA1-7.
BA1-9: 50 m SW of BA1-8.

Block 2

BA2-1: 25 m E of 25 m N of bend in road (where the road leaves the section line), NW 1/4 of SW1/4 of NW1/4 of section 31, T38N, R5W, Emmet County, Michigan.[5]
BA2-2: 50 m E of BA2-1.
BA2-3: 50 m E of BA2-2.
BA2-4: 50 m S of BA2-1.
BA2-5: 50 m E of BA2-4.
BA2-6: 50 m E of BA2-5.
BA2-7: 50 m S of BA2-4.
BA2-8: 50 m E of BA2-7.
BA2-9: 50 m E of BA2-8.

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[3]Block 3 was abandoned because of possible imperfect drainage.

[4]Later sampled as Nipissing pedon.

[5]Later sampled as Battlefield pedon.

<u>Block 3</u>

BA3-1: 50 m S of road, SW1/4 of SE1/4 of NW1/4, section 36, T38N, R6W, Emmet County, Michigan.
BA3-2: 50 m S of BA3-1.
BA3-3: 50 m S of BA3-2.
BA3-1: 50 m W of BA3-1 (about 25 m from road)
BA3-5: 50 m S of BA3-4.
BA3-6: 50 m S of BA3-5.

Main Algonquin surface

<u>Block 1</u>

MA1-1: 50 m W of road, 50 m N of where clearcut area intersects with the road, NW1/4 of SW1/4, section 31, T38N, R5W.
MA1-2: 50 m W of MA1-1.
MA1-3: 50 m W of MA1-2.
MA1-4: 50 m N of MA1-1.
MA1-5: 50 m W of MA1-4.
MA1-6: 50 m W of MA1-5.
MA1-7: 50 m S of MA1-1.
MA1-8: 50 m W of MA1-7.
MA1-9: 50 m W of MA1-8.

<u>Block 3</u>[6]

MA3-1: near the corner of Levering Rd and smaller unnamed road, 25 m E of telephone pole at edge of forest and 75 m N of Levering Rd (50 m N of large beech tree), SW1/4 of SW1/4 of section 31, T38N, R5W.
MA3-2: 50 m N of MA3-1.
MA3-3: 50 m N of MA3-2.
MA3-4: 50 m E of MA3-1.
MA3-5: 50 m N of MA3-4.[7]
MA3-6: 50 m N of MA3-5.
MA3-7: 50 m E of MA3-4.
MA3-8: 50 m N of MA3-7.
MA3-9: 50 m N of MA3-8.

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[6] Block 2 was abandoned.

[7] Later sampled as Main Algonquin pedon.

APPENDIX B

Table 26. Physical analysis data.

Horizon	Depth (cm)	Coarse >2.0 (mm)	Fine Earth <2.0 (mm)	Sand 2.0-0.05 (mm)	Silt 0.05-0.002 (mm)	Clay <0.002 (mm)
	----% of	total sample----		-----% of	fine earth-----	
			Algoma 1*			
E	0-10	0.1	99.9	98.5	1.4	0.1
				99.1	0.8	0.1
Bs	10-38	0.0	100.0	99.5	0.4	0.1
				99.6	0.3	0.1
C	38-68	0.0	100.0	99.7	0.2	0.1
				99.7	0.2	0.1
2Ab	68-70	0.1	99.9	99.3	0.7	0.0
				99.4	0.5	0.1
2C	70+	0.4	99.6	99.4	0.5	0.1
				99.4	0.5	0.1
			Algoma 2			
E	0-10	0.3	99.7	99.2	0.7	0.1
				99.3	0.6	0.1
Bs	10-40	0.2	99.8	99.5	0.4	0.1
				99.5	0.4	0.1
C	40-83	1.9	98.1	99.4	0.5	0.1
				99.3	0.6	0.1
2C	83+	60.7	39.3	92.5	7.3	0.2
				92.6	7.2	0.2
				93.0	6.9	0.1

* Two numbers for each horizon represent replicated sample data.

Table 26 (cont'd.).

Horizon	Depth (cm)	Coarse >2.0 (mm)	Fine Earth <2.0 (mm)	Sand 2.0–0.05 (mm)	Silt 0.05–0.002 (mm)	Clay <0.002 (mm)
	----% of	total sample----		-----% of	fine earth-----	
			Nipissing			
A	0-3	0.6	99.4	95.6	4.3	0.1
				96.3	3.4	0.3
				95.9	3.9	0.2
E1	3-15	0.1	99.9	97.6	2.3	0.1
				97.6	2.3	0.1
E2	15-30	0.3	99.7	98.2	1.7	0.1
				98.3	1.6	0.1
Bs1	30-45	2.3	97.7	98.6	1.2	0.2
				98.5	1.3	0.2
Bs2	45-75	7.6	92.4	98.6	1.2	0.2
				98.6	1.2	0.2
Bs3	75-98	2.3	97.7	99.2	0.7	0.1
				99.3	0.6	0.1
BC	98-110	21.6	78.4	98.7	1.1	0.2
				99.0	0.9	0.1
C	110+	10.6	89.4	99.4	0.5	0.1
				99.6	0.3	0.1

Table 26 (cont'd.).

Horizon	Depth (cm)	Coarse >2.0 (mm)	Fine Earth <2.0 (mm)	Sand 2.0-0.05 (mm)	Silt 0.05-0.002 (mm)	Clay <0.002 (mm)
	----% of	total sample----		-----% of	fine earth-----	
			Battlefield			
A	0-3	0.5	99.5	92.0	7.9	0.1
				92.0	7.7	0.3
				92.8	6.9	0.2
E1	3-18	0.4	99.6	95.2	4.7	0.1
				95.0	4.8	0.2
E2	18-23	2.1	97.9	93.9	5.7	0.4
				94.1	5.5	0.4
Bs1	23-30	1.6	98.4	94.3	5.3	0.4
				94.0	5.5	0.5
				93.5	6.1	0.4
Bs2	30-35	0.9	99.1	97.3	2.4	0.3
				96.6	3.1	0.3
				97.7	2.1	0.2
Bs3	35-55	0.4	99.6	98.2	1.5	0.3
				98.6	1.2	0.2
BC	55-67	0.4	99.6	99.2	0.7	0.1
				99.2	0.7	0.1
C1	67+	0.2	99.8	99.5	0.4	0.1
				99.5	0.4	0.1
C2		0.1	99.9	99.7	0.2	0.1
				99.7	0.2	0.1

Table 26 (cont'd.).

Horizon	Depth (cm)	Coarse >2.0 (mm)	Fine Earth <2.0 (mm)	Sand 2.0-0.05 (mm)	Silt 0.05-0.002 (mm)	Clay <0.002 (mm)
	----% of	total sample----		-----% of	fine earth-----	
			Main Algonquin			
A	0-8	0.6	99.4	91.3	8.6	0.1
				92.4	7.3	0.3
				92.8	7.0	0.2
E	8-30	0.9	99.1	94.9	5.0	0.1
				95.4	4.5	0.1
				95.1	4.8	0.2
Bhs	30-38	0.5	99.5	96.6	3.0	0.4
				96.4	3.1	0.5
Bs1	38-65	0.2	99.8	97.7	1.9	0.3
				97.9	1.8	0.3
Bs2	65-93	0.1	99.9	98.9	0.9	0.2
				99.2	0.7	0.1
BC	93-125	0.0	100.0	99.7	0.2	0.1
				99.7	0.2	0.1
C	125+	0.7	99.3	99.6	0.3	0.1
				99.5	0.4	0.1

Table 27. Sand separates data.

Horizon	Very Coarse (2-1 mm)	Coarse (1-0.5 mm)	Medium (0.5-0.25 mm)	Fine (0.25-0.1 mm)	Very Fine (0.1-0.05 mm)
	-------------------	% of 2.0 - 0.05 mm	-------------------		
			Algoma 1*		
E	0.3	4.0	45.8	49.8	0.1
	0.8	4.9	66.5	27.7	0.1
Bs	0.0	5.0	74.0	20.9	0.1
	0.05	4.53	68.35	26.98	0.09
C	0.03	3.32	68.02	28.50	0.13
	0.04	3.06	69.69	27.12	0.09
Ab	0.04	7.67	65.91	26.22	0.16
	0.12	9.78	67.82	22.14	0.14
2C	0.79	14.92	63.66	20.43	0.20
	0.88	15.81	67.51	15.40	0.40
			Algoma 2		
E	0.38	24.99	72.09	2.39	0.15
	0.51	22.07	74.90	2.42	0.10
Bs	0.18	20.24	76.32	3.14	0.12
	0.13	20.01	76.51	3.26	0.09
C	0.79	32.06	64.81	2.18	0.14
	1.05	32.64	64.01	2.16	0.15
2C	7.36	33.35	51.24	5.60	2.45
	8.71	37.87	46.83	4.66	1.93
	8.63	37.30	48.22	4.34	1.52

* Two numbers for each horizon represent replicated sample data.

Table 16 (cont'd.).

Horizon	Very Coarse (2-1 mm)	Coarse (1-0.5 mm)	Medium (0.5-0.25 mm)	Fine (0.25-0.1 mm)	Very Fine (0.1-0.05 mm)
	-------------------% of 2.0 - 0.05 mm				-----------------
			<u>Nipissing</u>		
A	0.05	7.52	82.22	9.75	0.47
	1.17	9.59	79.70	8.67	0.86
	1.02	10.97	79.59	8.11	0.32
E1	0.25	7.40	81.85	9.82	0.68
	0.11	8.37	80.90	9.99	0.63
E2	0.15	6.05	82.50	10.87	0.42
	0.14	6.52	81.87	11.01	0.45
Bs1	0.11	7.37	84.97	7.37	0.17
	0.09	6.75	85.16	7.82	0.17
Bs2	0.29	10.95	83.84	4.88	0.04
	0.27	12.36	82.57	4.74	0.06
Bs3	0.88	21.67	73.42	3.96	0.07
	0.98	22.28	72.96	3.74	0.04
BC	1.12	13.25	75.71	9.53	0.39
	1.35	15.50	75.60	7.27	0.29
C	0.22	4.27	84.86	10.38	0.27
	0.26	5.36	83.53	10.77	0.09

Table 16 (cont'd.).

Horizon	Very Coarse (2-1 mm)	Coarse (1-0.5 mm)	Medium (0.5-0.25 mm)	Fine (0.25-0.1 mm)	Very Fine (0.1-0.05 mm)
	---------	---------% of 2.0	- 0.05 mm----	---------	---------
			Battlefield		
A	0.44	7.87	60.66	29.71	1.32
	0.89	11.63	70.02	16.46	1.00
	1.02	10.51	65.18	22.17	1.11
E1	0.34	9.28	69.02	20.09	1.26
	0.73	10.87	65.02	22.12	1.26
E2	0.57	8.10	66.45	23.90	0.98
	0.54	7.45	61.59	29.50	0.92
Bs1	0.74	7.74	64.29	26.44	0.79
	0.70	8.04	68.31	22.49	0.45
	0.89	7.11	62.15	29.05	0.81
Bs2	0.50	7.08	61.26	30.75	0.41
	0.61	7.11	63.95	27.93	0.40
	0.61	7.30	64.38	27.40	0.31
Bs3	0.54	7.76	67.10	24.37	0.23
	0.46	6.46	57.88	34.90	0.29
BC	0.41	3.93	49.88	44.97	0.80
	0.57	3.94	42.46	52.10	0.93
C1	0.25	2.52	57.44	39.31	0.49
	0.26	3.03	54.68	41.47	0.57
C2	0.63	14.60	72.74	11.95	0.08
	0.47	13.71	72.91	12.86	0.04

Table 16 (cont'd.).

Horizon	Very Coarse (2-1 mm)	Coarse (1-0.5 mm)	Medium (0.5-0.25 mm)	Fine (0.25-0.1 mm)	Very Fine (0.1-0.05 mm)
	-------------------	---% of 2.0 - 0.05 mm	---	---	-------------------
			Main Algonquin		
A	0.57	11.51	78.07	8.65	1.21
	0.91	14.48	76.84	6.92	0.84
	1.03	14.49	76.77	7.20	0.52
E	0.34	10.37	79.93	8.64	0.71
	0.32	11.69	79.18	7.94	0.87
	0.37	11.27	79.54	8.02	0.80
Bhs	0.17	7.16	82.15	10.21	0.30
	0.16	7.33	81.69	10.50	0.32
Bs1	0.09	4.09	80.08	15.42	0.33
	0.06	4.63	82.97	12.25	0.09
Bs2	0.05	4.33	80.65	14.68	0.29
	0.06	3.22	83.19	13.44	0.09
BC	0.06	2.08	70.14	27.68	0.04
	0.05	2.60	73.07	24.21	0.08
C	0.44	6.54	68.77	23.95	0.31
	0.45	6.09	69.11	24.07	0.27

Table 28. Chemical data.

Horizon	Dithionite extractable Fe	Dithionite extractable Al	Pyrophosphate extractable Fe	Pyrophosphate extractable Al	Oxalate extractable Fe	Oxalate extractable Al	pH	Organic C	ODOE
	%						(2:1 H_2O)	%	
					Algoma 1*				
O								33.3 36.1	
E	0.14	0.03	0.08	0.02	0.12	0.04	4.7 4.8	0.8 0.7 0.6	0.133
Bs	0.05	0.01	0.01	0.01	0.02	0.02	5.7 5.9	0.1 0.1	0.018
C	0.03	0.00	0.00	0.00	0.01	0.02	6.6 6.4	0.1 0.1 0.1	0.012
Ab	0.03	0.01	0.00	0.01	0.01	0.02	7.2	0.2 0.3	0.058
2C	0.04	0.00	0.00	0.00	0.02	0.02	7.7 7.3	0.0 0.0	0.056
					Algoma 2				
O								4.2 3.5	
E	0.06	0.01	0.01	0.01	0.03	0.01	5.3 5.2	0.2 0.1	0.018
Bs	0.05	0.01	0.01	0.01	0.01	0.02	6.0 5.9	0.1 0.1 0.1	0.014
C	0.05	0.00	0.00	0.01	0.01	0.01	6.4 6.6	0.0 0.1 0.1	0.003
2C	0.08	0.01	0.01	0.01	0.01	0.01	7.6 7.6	0.3 0.3 0.4	0.010

* Two numbers for each horizon represent replicated sample data.

Table 17 (cont'd).

Horizon	Dithionite extractable Fe	Dithionite extractable Al	Pyrophosphate extractable Fe	Pyrophosphate extractable Al	Oxalate extractable Fe	Oxalate extractable Al	pH	Organic C	ODOE
	%						(2:1 H_2O)	%	
				Nipissing					
A	0.07	0.03	0.02	0.02	0.03	0.05	4.2	2.3	0.050
							4.4	1.6	
								2.1	
E1	0.06	0.01	0.00	0.01	0.01	0.02	4.8	0.2	0.012
							4.9	0.2	
E2	0.08	0.01	0.01	0.01	0.03	0.02	5.1	0.1	0.028
							5.0	0.1	
								0.1	
Bs1	0.12	0.04	0.03	0.03	0.07	0.07	5.1	0.1	0.024
							4.9	0.1	
								0.1	
Bs2	0.13	0.07	0.02	0.05	0.09	0.23	5.2	0.1	0.022
							5.2	0.1	
								0.1	
Bs3	0.08	0.04	0.02	0.03	0.04	0.09	5.3	0.1	0.015
							5.1	0.1	
								0.1	
BC	0.09	0.05	0.01	0.03	0.05	0.13	5.8	0.1	0.021
							5.7	0.1	
C	0.08	0.02	0.01	0.01	0.02	0.04	6.6	0.0	0.006
							6.6	0.0	
								0.0	

Table 17 (cont'd).

Horizon	Dithionite extractable Fe	Dithionite extractable Al	Pyrophosphate extractable Fe	Pyrophosphate extractable Al	Oxalate extractable Fe	Oxalate extractable Al	pH	Organic C	ODOE
	%						(2:1 H_2O)	%	
					Battlefield				
O								18.9 20.5 20.3	
A	0.06	0.02	0.01	0.01	0.02	0.03	4.0	2.1 1.8 1.5	0.041
E1	0.04	0.01	0.00	0.00	0.01	0.01	4.9 4.4	0.3 0.2	0.007
E2	0.13	0.03	0.04	0.02	0.07	0.05	4.4 4.4	0.5 0.4	0.054
Bs1	0.28	0.19	0.12	0.12	0.23	0.36	4.8 4.7	0.8 0.8 0.8	0.090
Bs2	0.19	0.29	0.06	0.17	0.14	0.56	5.0 5.1	0.8 0.8 0.8	0.223
Bs3	0.11	0.15	0.02	0.08	0.06	0.31	5.0 4.9	0.4 0.3 0.3	0.039
BC	0.06	0.09	0.01	0.06	0.04	0.20	5.4 5.4	0.1 0.1	0.018
C1	0.04	0.05	0.01	0.04	0.03	0.13	5.5 5.3	0.1 0.1	0.011
C2	0.03	0.03	0.01	0.02	0.02	0.07	5.5 5.4	0.1 0.1	0.010

Table 17 (cont'd).

Horizon	Dithionite extractable Fe	Dithionite extractable Al	Pyrophosphate extractable Fe	Pyrophosphate extractable Al	Oxalate extractable Fe	Oxalate extractable Al	pH	Organic C	ODOE
	%						(2:1 H_2O)	%	
				Main Algonquin					
A	0.11	0.03	0.02	0.02	0.03	0.03	4.6 4.5	2.0 2.1 2.5	0.041
E	0.10	0.01	0.01	0.01	0.03	0.01	4.7 4.5	0.2 0.5 0.2	0.012
Bhs	0.38	0.15	0.13	0.09	0.32	0.28	5.4 5.5	0.8 0.7	0.277
Bs1	0.20	0.18	0.06	0.10	0.17	0.33	5.6 5.5	0.4 0.4	0.102
Bs2	0.08	0.08	0.02	0.06	0.07	0.19	5.9 5.8	0.2 0.2	0.036
BC	0.06	0.03	0.01	0.03	0.02	0.09	5.9 5.9	0.1 0.1	0.020
C	0.05	0.03	0.01	0.03	0.03	0.08	6.0 6.1	0.1 0.1	0.016

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