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Factors affecting the formation of dark, thick epipedons beneath forest vegetation, Michigan, USA

R. J. SCHAETZL
Department of Geography, Michigan State University, East Lansing, Michigan 48824-1115, USA

SUMMARY

Extremely gravelly, coarse-textured soils (frigid Udorthents and Rendolls) with different thicknesses of Oa+A horizon sequences were studied to identify factors that have influenced their genesis. These well-drained, forested soils occur on geomorphic surfaces that range in age from 3200 to 6000 years BP. The soils all have more than 500 g kg^{-1} coarse fragments by mass; most contain less than 300 g kg^{-1} fine earth. In the lower solum of most pedons, content of cobbles increases and amount of fine earth decreases. Most coarse fragments are dolomite and chert.

Thick, gravelly Oa and A horizons are weakly correlated with parent material characteristics such as high pH and carbonate contents. Organic matter concentrations in, and thicknesses of, upper horizons are enhanced by an abundance of coarse clasts, as soils with the most gravel exhibited the thickest and darkest epipedons. Relatively high amounts of crystalline clasts in the fine gravel fraction, as well as feldspathic minerals in the fine sands, also appear to promote the development of mollic epipedons.

INTRODUCTION

With the exception of some Aquolls, most Mollisols have developed primarily under grassland vegetation or intermixed periods of forest/grassland dominance (Walker, 1966; Fenton, 1983; Smith, 1986). Afanas'yeva (1966), however, reported the occurrence of Chernozems under oak forest in the steppes of the USSR. Thus, well-drained, forested soils with dark, thick A horizons often have experienced a period of grassland dominance in the past (Geis *et al.*, 1970) or exist under a relatively open canopy with grasses in the understorey (Al-Barrak & Lewis, 1978; Nimlos & Tomer, 1982). Many Mollisols, in which the epipedon is thought to have formed under forest, also have abundant cobbles and gravels in the profile (e.g. Gaikawad & Hole, 1965; Small *et al.*, 1990).

Certain well-drained, forested soils on Bois Blanc Island in Lake Huron, Michigan have dark, thick epipedons, often underlying thick Oa horizons that have abundant gravels. The soils have formed in extremely gravelly and cobbly sediments, the coarse fragments of which are predominantly calcareous. Other soils on Bois Blanc Island lack these dark surface horizons, despite being formed in gravelly parent materials on the same geomorphic surfaces, and under similar vegetation cover. At present, there is little or no herbaceous vegetation in the understorey, and paleovegetation reconstructions for the island do not recognize a period of grassland dominance. It was assumed, therefore, that the dark epipedons formed under the influence of forest vegetation. The purpose of this paper is to document and characterize these forest soils, and to ascertain which soil-forming factors promote the development of thick, dark epipedons in this environment.

STUDY AREA

Bois Blanc Island, in north-western Lake Huron, USA (Fig. 1), was deglaciated approximately 11 200 years BP (Larsen, 1987). Most surficial deposits are gravelly, wave-worked glacial drift. The

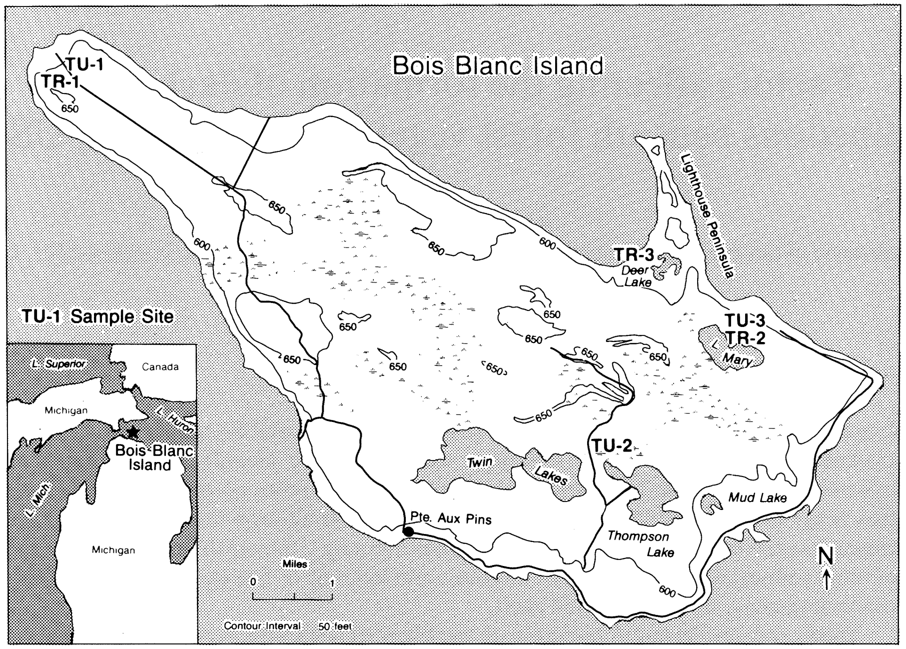

Fig. 1. The study area: Bois Blanc Island, Michigan. Pedon locations are labelled as in Tables 1 and 2.

course clasts are primarily dolomite and chert, reflecting the influence of the local dolomite bedrock. The fine earth fraction is strongly calcareous. Soils examined in this study are classified within loamy-skeletal and sandy-skeletal particle-size families.

Immediately following deglaciation (11 200–10 000 years BP; Larsen, 1987), the island lay in proglacial Lake Algonquin, which had a higher lake level than does the modern Lake Huron because low outlets were blocked by the retreating ice sheet. Isostatic rebound and emergence of lower outlets produced several episodic changes in lake level during the following 8000 years (Zumberge & Potzger, 1956; Larsen, 1987). Thus, three to possibly five or more shoreline features are evident on the island.

The climate of Bois Blanc Island is cool, humid continental. Soils are in the frigid soil temperature regime (Soil Survey Staff, 1975). National Weather Service, 1951–1980 data for Cheboygan, a station 10 km south of the island, indicate that July is the warmest month (20.0°C). Mean January and February temperatures are both −7.2°C. Temperatures on the island are moderated more than those inland because of the effects of the lake. The primary influence of the surrounding water is to retard warming in spring and to offset cooling in autumn. The mean annual precipitation totals are 712 mm at Cheboygan, with late summer being the wettest period. Snow cover is extensive from late November to March.

Pre-settlement forests on the island were logged in the late 1800s, and abundant fires occurred in the cut-over wood. Second-growth forest now covers most of the island. Dry sites are dominated by northern hardwood forests including species such as sugar maple (*Acer saccharum* Marsh.), white birch (*Betula papyrifera* Marsh.), eastern white pine (*Pinus strobus* L.), balsam fir, (*Abies balsamea* L.), yellow birch (*Betula alleghaniensis* Britt.), basswood (*Tilia americana* L.) and trembling aspen-

METHODS

Unpublished map sheets (Mackinac County soil survey; Soil Conservation Service) were used to locate mapping units of well-drained soils that have dark epipedons of varying thickness. Approximately 20 pedons were examined in the field, and five representative pedons from this larger set were eventually sampled and described by horizon (Soil Survey Staff, 1951). Bulked samples (approximate mean sample mass 3 kg) of the combined gravel and fine earth fraction were taken for laboratory characterization. These very gravelly soils were sampled by removing entire, large volumes of each horizon, beginning at the surface and continuing downward, and subdividing each sample to achieve the standard sample mass of 3 kg. Cobbles (>75 mm diameter) were removed from the sample before transporting to the laboratory; their content was estimated from the larger sample. Standard laboratory analysis included: (i) pipette method for particle-size analysis (PSA) of the fine-earth fraction, (ii) pH in H_2O (2:1 ratios for mineral soils and 8:1 ratios for organic horizons), (iii) organic matter (OM) content of the fine-earth fraction by loss-on-ignition at 550°C, (iv) inorganic carbon content, determined gravimetrically with $HCl–FeCl_2.4H_2O$ (Kilmer & Alexander, 1949; Lee *et al.*, 1972; Sheldrick, 1984; further detail is provided in Schaetzl, 1991). OM content was converted to organic carbon (OC) percentages by $OC=OM/1.9$ (Broadbent, 1953). Because bulk density was not measured the OC data are gravimetric. Analyses were run in duplicate for all samples with the exception of PSA; only mean values are reported. Weighted profile contents of OC, $CaCO_3$ and other constituents were calculated by the following equation: $\Sigma(ZH_t)/D$, where Z is the soil constituent under consideration, H_t is the thickness of a horizon in cm, D is the total thickness (cm) over which the analysis was performed, and the summation (Σ) is calculated for all horizons in the given thickness.

The mineralogy of the fine and very fine sand separate was determined by X-ray diffraction (XRD). Whole soil samples were pretreated with H_2O_2, shaken overnight, and wet sieved to isolate the 50–250 μm fraction. This fraction was then dried and ground to pass a 50 μm sieve before placing as a dry mount on glass slides for XRD analysis. The samples were X-rayed with Cu Kα radiation at 35 kV and 15 mA (Schaetzl, 1991). The XRD data were quantified by comparing peak heights for dolomite (d-spacing: 0.289 nm), plagioclase feldspar (0.318 nm), and orthoclase feldspar (0.325 nm) with that of quartz (0.334 nm; cf. Berry 1987). Mean values of three X-ray analyses per sample are reported. Fine gravels (2–6.3 mm diameter) were separated into acidic igneous (e.g. granite, rhyolite) and basic igneous (basalt, gneiss) fractions by hand, and the masses were compared with a third fraction, composed primarily of carbonaceous clasts, chert and quartz (cf. Lee *et al.*, 1972).

RESULTS AND DISCUSSION

Most well-drained soils with dark, thick epipedons on Bois Blanc Island can be classified into one of two subgroups. Typic Udorthents have an ochric epipedon and an A–C or an A–Bw–C profile where the Bw horizon is too coarse-textured to classify as a cambic horizon. In some Udorthents an incipient E horizon is evident as bleached and uncoated sand grains in the lower A horizon, but it is rarely laterally continuous. Typic Rendolls resemble Udorthents morphologically but lack indications of eluviation, and have a thicker, dark surface horizon that qualifies as a mollic epipedon (Soil Survey Staff, 1975). Either type of profile may have a thick, gravelly Oa horizon above the ochric or mollic epipedon. Both soil types are found on stable uplands and are extremely permeable. The pedons studied are located on the two lowest (youngest) geomorphic surfaces on the island, the Nipissing and Algoma lake terraces, which date between 6000 years BP and 3200 years BP (Larsen, 1987). Field observations suggest that variability in soil development (i.e. degree of horizonation and content of organic matter) is greater within the two geomorphic surfaces than between them.

The type of forest cover is similar on all sites at present, although *Thuja occidentalis* L. is slightly more abundant on some sites, especially near pedons TR-1 and TU-2. *A. saccharum* and *B. papyrifera* dominate the young (35–55 years), second-growth forests on most sampled sites. Changes in paleovegetation during the Holocene period for the upper midwest reflect general climatic amelioration and drying (Webb & Bryson, 1972; Webb *et al.*, 1983), and there is no reason to expect that the changes in climate and vegetation were not synchronous across the island. The

Table 1. Physical characteristics of soils

Horizon	Depth (cm)	Colour moist (dry)	Particle size (mm) distribution (g kg^{-1} of whole soil)					Particle size (μm) distribution (g kg^{-1} of fine earth)				Particle size class[b]
			> 75[a]	19.1–75	12.7–19.1	2.0–12.7	< 2	250–2000	50–250	2–50	< 2	
Typic Rendoll, sandy-skeletal, frigid[c] (TR-1)												
Oi	31–29		0	0	0	0	0	nd[d]	nd	nd	nd	
Oe	29–21	N 2/0	0	0	0	0	0	nd	nd	nd	nd	
Oa	21–0	10YR 2/1 (10YR 3/2)	50	570	142	161	77	nd	nd	nd	nd	exgr frag
A	0–16	10YR 2/2 (10YR 3/2)	550	313	45	63	29	183	97	664	56	exco sil
AC	16–35	10YR 3/2 (10YR 5/1 and 5/2)	600	184	29	133	55	733	25	211	31	exco lcs
C	35–60	10YR 7/2	400	190	47	166	196	924	28	44	4	exgr cs
Typic Rendoll, sandy-skeletal, frigid[c] (TR-2)												
Oi	2–0		0	0	0	0	0	nd	nd	nd	nd	
A	0–12	10YR 2/1 (10YR 3/1)	50	385	169	170	226	807	65	89	39	exgr s
B/A	12–26	7.5YR 4/2 (B) 7.5YR 3/0 (A) (10YR 5/3 mixed)	150	200	91	249	310	835	67	59	38	exgr s
Bw	26–63	10YR 4/3	500	187	48	126	139	856	66	59	19	exco s
C	63–70	10YR 5/2	550	102	47	214	87	506	128	315	51	exco csl
Typic Rendoll, sandy-skeletal, frigid[c] (TR-3)												
Oi	19–18		0	0	0	0	0	nd	nd	nd	nd	
Oe	18–16	2.5YR 2.5/0 (7.5YR 3/2)	0	0	0	0	0	nd	nd	nd	nd	
Oa	16–0	7.5YR 2/0 (7.5YR 2/0)	0	343	233	365	59	nd	nd	nd	nd	exgr frag
A	0–25	10YR 3/2 (10YR 5/2)	0	348	320	309	23	510	32	26	432	exgr sC
C	25–69	10YR 4/2	400	105	87	245	164	858	21	108	13	exgr s

Typic Udorthent, sandy-skeletal, frigid[c] (TU-1)												
Oi	3–0		0	0	0	0	0	nd	nd	nd	nd	
A1	0–13	10YR 2/1 & /2 (10YR 3/2)	0	216	98	193	492	796	29	127	49	vgr lcs
A2	13–18	10YR 2/2 & /3 (10YR 5/2)	0	122	96	293	490	848	19	103	30	vgr ls
2C	18–70	10YR 5/3	600	73	42	136	150	873	16	100	11	exco s
Typic Udorthent, sandy-skeletal, frigid[c] (TU-2)												
Oi	2–0		0	0	0	0	0	nd	nd	nd	nd	
A1	0–10	7.5YR 2/0 (10YR 3/1)	200	107	65	355	273	827	40	98	35	exgr ls
A2	10–22	10YR 2/1 (10YR 4/2)	50	129	143	402	276	820	33	97	50	exgr ls
Bw	22–55	10YR 4/2	100	209	115	278	298	849	52	77	22	exgr s
2C	55–71	10YR 5/3	600	56	73	155	116	617	207	151	26	exco ls
Typic Udorthent, sandy-skeletal, frigid[c] (TU-3)												
Oi	19–18		0	0	0	0	0	nd	nd	nd	nd	
Oe	18–16	5YR 2.5/2 (10YR 3/2)	0	0	0	0	0	nd	nd	nd	nd	
Oa	16–0	10YR 2/1 (10YR 4.5/1)	0	459	255	240	45	nd	nd	nd	nd	exgr frag
A	0–17	10YR 2/1 (10YR 4/1 & /2)	250	565	76	91	18	228	167	153	453	exgr C
Cd	17–54	10YR 7/2 & /3	500	303	61	84	51	866	10	88	36	exco cs

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[a] Estimated. [b] Particle size class abbreviations: ex = extremely, v = very, co = cobbly, gr = gravelly, frag = fragmental, c = coarse, f = fine, s = sand(y), si = silt(y), C = clay, l = loam(y). [c] The mineralogy families of the pedons were not determined. [d] No data.

Table 2. Chemical and mineralogical properties of soils

pH H_2O	Organic carbon (g kg^{-1})	Inorganic carbon	Gravels in 2–6.3 mm fraction[a] Granitic (g kg^{-1})	Basaltic	Ratio of XRD peak heights × 100 Dolomite/ quartz	Plagioclase/ quartz	Orthoclase/ quartz
5.27	nd	nd	0	0	nd	nd	nd
5.72	435	nd	0	0	nd	nd	nd
7.44	226	nd	14	8	nd	nd	nd
7.65	131	58	tr[b]	tr	95.7	3.7	2.0
7.82	40	53	10	17	49.3	21.0	2.7
8.37	9	36	20	58	66.0	10.3	5.3
5.56	nd	nd	0	0	nd	nd	nd
5.68	61	5	tr	tr	0	4.0	1,3
6.89	19	1	tr	tr	0.3	41.0	4.0
7.81	12	8	tr	6	9.0	5.3	7.0
8.04	17	32	11	17	33.7	3.3	9.3
5.13	nd	nd	0	0	nd	nd	nd
4.67	412	nd	0	0	nd	nd	nd
5.80	280	nd	10	tr	nd	nd	nd
7.86	61	28	12	tr	56.3	1.0	0.3
8.21	8	23	14	38	55.3	3.3	4.0

)							
6.29	nd	nd	0	0	nd	nd	nd
7.78	54	1.0	57	139	112.7	16.0	5.0
7.76	27	2.5	42	94	277.0	30.3	13.0
8.11	11	4.3	51	96	nd	nd	nd
)							
6.16	nd	nd	0	0	nd	nd	nd
7.05	116	0.6	13	7	0.3	5.7	5.0
7.21	36	0.4	11	13	3.7	8.7	7.0
8.02	14	1.3	31	48	33.3	14.3	11.7
8.09	11	2.4	45	25	30.7	5.3	6.3
)							
5.36	nd	nd	0	0	nd	nd	nd
5.07	379	nd	0	0	nd	nd	nd
5.31	227	nd	0	0	nd	nd	nd
7.38	106	6	0	0	3.3	0	1.0
7.82	27	7	tr	tr	55.0	1.3	4.3

of gravels with igneous or metamorphic lithology, further subdivided into granitic and basaltic subgroups. The remainder of the gravels are clasts of quartzite or sandstone. [b]Trace amounts (< 1 g kg^{-1}).

soils do not exhibit continuous E horizons—a feature that might be expected under many well-drained forested soils.

Among these soils, the factors of vegetation, climate, time, and relief did not appear to influence strongly the types of profile that have developed. All sampled pedons were located on level, upland surfaces under similar vegetation assemblages. Therefore, the effects of parent material were analysed to provide insight into their genesis.

In several respects, parent materials are similar among the pedons studied. With two exceptions, all mineral horizons are extremely gravelly or cobbly and contain less than 500 g kg^{-1} fine earth (Table 1), with most horizons containing less than 300 g kg^{-1} fine earth. Many pedons exhibit a large increase in cobble content in the C and lower B horizons, below a lithologic discontinuity. The lithology of the gravel and cobble fraction is primarily dolomitic, with many chert fragments in the smaller gravel and coarse sand fractions. Crystalline clasts typically comprise less than 50 g kg^{-1} by mass of the gravel fraction (Table 2). Based on reaction and inorganic carbon data of the C horizon, the original parent material probably contained abundant free lime (Tables 2 and 3). Despite the similarities in the parent materials of the soils examined, there remain subtle differences that may be cited as possible explanations for varying genesis of these soils, of which melanization plays a major role.

Melanization processes and considerations

In part because of the affinity of humus for clays and colloidal-sized particles, the amount of clay in soils is often positively correlated with OC content (Greenland, 1965; Mortland, 1970; Anderson *et al.*, 1975*b*; Nichols, 1984). Similarly, because clay content and water-holding capacity are positively correlated, an association is often found in dry climates between clay content and increased organic matter production. In the cool, humid climate of Bois Blanc Island, a weak positive relationship exists between weighted profile OC content and clay content (to 30 cm), but statistical significance is not attained ($r = 0.42$, $P = 0.26$).

Melanization and the accumulation of OC is favoured in parent materials rich in bases, especially in calcium, which allow stable Ca–humus complexes (Ca–humates) to form (Smith *et al.*, 1950; Gaikawad & Hole, 1965; Anderson *et al.*, 1975a). Calcium stimulates the decomposition of fresh organic materials, but may slow the decomposition of humus and highly humified substances (Kononova, 1961). It promotes OC accumulation in upper horizons by precipitating certain organic substances, and inhibits advanced decomposition and biodegradation by tying up functional (especially phenolic and carboxylic) groups of organic molecules, thereby suppressing microbial attack (Bochter & Zech, 1985). Although all the soils under study have presumably formed in calcareous parent materials, the pedons with the greater OC contents (weighted profile mean values) appear to have the thickest surface horizons with the largest inorganic carbon contents (Table 3). This finding is in agreement with the general correlation between lime-rich parent materials and mollic epipedons worldwide (e.g. Nørnberg *et al.*, 1985). The statistical correlation between weighted inorganic carbon (free $CaCO_3$) content (to 50 cm) and weighted OC content (to 30 cm) is positive but not significant ($r = 0.15$, $P = 0.71$), probably because most of the calcium is complexed with organic molecules, and is not present as free lime.

Melanization and the accumulation of organic matter with depth in these soils appear to be promoted by large contents of coarse fragments, as indicated by the strong correlation between weighted OC and coarse fragment content ($r = 0.93$, $P < 0.001$). Here, the melanization process is primarily manifested as organic coatings on skeletal soil particles or structural units (e.g. McKeague, 1971; Broersma & Lavkulich, 1980; Protz, 1983), but also it is documented by the presence of clay–humus complexes (Anderson *et al.*, 1975b), and silt-sized, plasmic aggregates of humus, more or less independent and devoid of skeletal grains (St Arnaud & Whiteside, 1964; DeConinck *et al.*, 1974).

Humic substances that exist as coatings on particles larger than 2 mm diameter are not routinely included in laboratory determinations of the amount of OC in a soil sample. Thus, organic matter determinations from dark horizons with large contents of coarse fragments may underestimate the actual amount of organic matter present because these organic coatings (organs, organans; Protz, 1983), ubiquitous in the upper horizons of all soils studied, are not included in the analysis.

Table 3. Comparative soil data

	Udorthents with thin Oa + A horizons[a,b]	Udorthents with thick Oa + A horizons[a]	Rendolls
Number of pedons in sample	2	4	3
Type of epipedon	ochric	ochric	mollic
Oa + A horizon thickness[c]	6	25	41
Cobbles (g kg^{-1})[d,e]	100	247	313
Fine earth (g kg^{-1})[d,e]	208	157	125
Inorganic carbon (g kg^{-1})[d]	11	17	26
pH of C horizon	8.13	7.96	8.21
Granite + basalt in fine gravel fraction (g kg^{-1})[d,e]	21	561	258
Feldspar/quartz peak height ratio (× 100)[d,f]	17.3	10.7	12.6

[a] Includes data from pedons described in Schaetzl (1991). [b] Includes data from a Eutrochrept pedon, described in Schaetzl (1991) that morphologically is similar to the Udorthent pedons described in this study. [c] In cm. Includes transitional (e.g. AC) and mixed (e.g. B/A) horizons. [d] Weighted mean values of pedons in the same subgroup are combined and an overall subgroup mean is determined for the profile. [e] Includes data from horizons that morphologically resemble A horizons but, because they have > 20 g kg^{-1} organic carbon, classify as Oa horizons. [f] Sum of plagioclase/quartz and orthoclase feldspar/quartz peak heights (see Table 2).

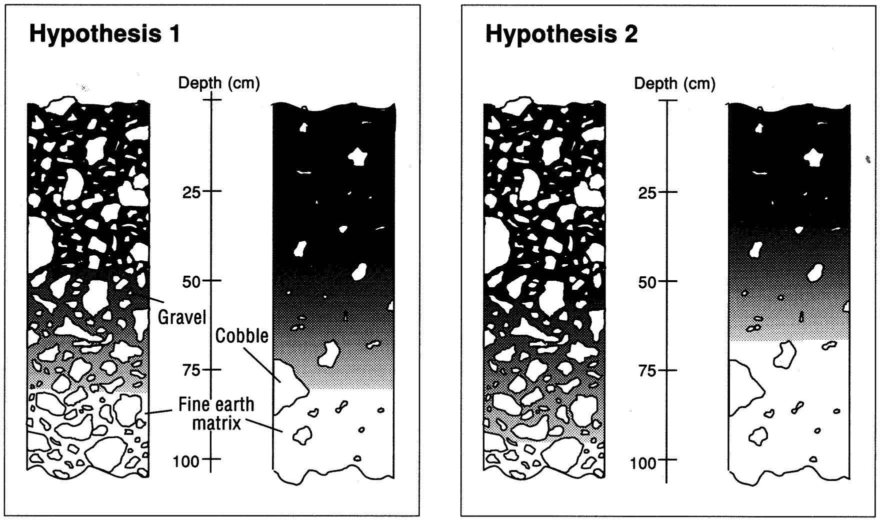

Fig. 2. Differing hypotheses of organic matter content and accumulation in two pedons that differ in content of coarse fragments. Darkness of shading is intended to correlate with amount of organic carbon in the fine earth fraction.

Holding other factors constant, as the amount of coarse fragments increases, the soil volume in which melanization can operate and OC can accumulate, is diminished. Thus, if similar amounts of OM are incorporated into two soils, one with abudant coarse fragments and gravel, and one with few gravels or cobbles, two hypotheses are envisaged (Fig. 2). (1) OM will accumulate to approximately similar depths in both soils, but in the extremely gravelly soil the amount of organic matter in 'inter-gravel' void spaces and the thickness of organs on skeletal grains (*sensu* Broersma & Lavkulich, 1980) will be greater than in the soil with little gravel, resulting in larger concentrations of organic matter in the former. (2) The amount of OM in 'inter-gravel' void spaces of the upper horizons of the extremely gravelly soil will not differ markedly from that of the soil with little gravel, but the OM will be incorporated more deeply into the profile of the former.

In addition to volume considerations, surface area also varies with coarse fragment content. As the proportion of coarse fragments increases, the surface area onto which organs can be distributed concomitantly decreases. Broersma & Lavkulich (1980) showed that the weight of organic coatings per unit surface area decreased as the total surface area decreased in the coarser (silt, sand) soil fractions.

Finally, the amount of gravel in soils might affect infiltration and patterns of root development. Deeper rooting will facilitate melanization at greater depths, and thicker A horizons will develop. Yair and colleagues (Yair & Lavee, 1985; Yair, 1987; Yair & Berkowitz, 1989) explained increased biological productivity, biomass, and leaching in areas of rocky arid soils, by invoking deep infiltration of the scant desert rainfall. Similar reasoning could explain the thicker Oa and A horizon sequences in the most gravelly Bois Blanc soils. Even in the humid climate of Bois Blanc, periods of moisture deficiency may occur in these soils, as they contain so little fine earth and clay (Table 1). During and following such periods, infiltration events will penetrate deeper into the more gravelly soils, as for example pedon TR-1, which may allow increased root proliferation at depth, thereby facilitating deeper OC accumulation and the development of a mollic epipedon.

In pedon TR-1, organic matter has accumulated to such a degree that a thick (21 cm), extremely gravelly Oa horizon overlies the mollic epipedon. The Oa horizon morphologically resembles a mollic horizon, but the large OC content ($226\,\mathrm{g\,kg^{-1}}$) requires it to be classified as an organic horizon (Soil Survey Staff, 1975). Pedons TR-3 and TU-3 also have Oa horizons that cannot be easily discerned from A horizons in the field. The fine earth in the upper profile of the above soils is less than $80\,\mathrm{g\,kg^{-1}}$ (Table 1), and because the coarse fragment content is large, the amount of OC, although very high in the upper horizons, may be underestimated.

In pedons TR-2, TU-1 and TU-2, the amount of fine earth in the upper profile is 150 to $>500\,\mathrm{g\,kg^{-1}}$; thus, organic matter is disseminated throughout a greater soil volume and over a greater surface area than in other, more gravelly soils (Table 1; Fig. 2). The quantity of OM not measured because it exists as coatings on gravel or cobble surfaces is proportionately less in these soils, and the 'real' amount (concentration) of OC present therein may be closer to the values reported in Table 2 than it is for pedons TR-1, TR-3 and TU-3. This example illustrates how coarse fragments can assume importance in the overall accumulation as well as the determination of carbon content.

Finally, field observations of soil fauna in these soils indicated that earthworms were often small (<5 cm) and usually confined to Oa and A horizons. Two worms removed from Oa horizons were identified as *Aporrectodea trapezoides*, a species known to frequent a wide variety of habitats, including sandy soils (Reynolds, 1977). Because this species does not routinely produce surface casts, as do some *Lumbricus* species, burial of coarse fragments is not likely (Birch & Clark, 1953). The extremely cobbly, coarse-textured AC and 2C horizons in these soils may restrict the vertical movements of the fauna during certain periods, thereby limiting the degree to which OM is incorporated into the lower solum and increasing the OM content of near-surface horizons.

Organic matter production and parent material

The intensity of melanization depends on the rate at which organic matter production and incorporation into the solum exceeds decomposition and mineralization therein. Documented cases of mollic epipedons beneath forest vegetation, where a period of grassland dominance is not inferred, generally occur in frigid or colder soil temperature regimes (Nimlos & Tomer, 1982; Anderson *et al.*, 1975a), or on the cool side of mesic temperature regimes (Gaikawad & Hole, 1965), where mineralization rates are slow.

Assuming that mineralization rates do not vary markedly among the profiles, differences in rates of OM production, both from above-ground (foliage, branches) and below-ground (root) sources, might account for some of the variability in OC content in these soils. OM production may be related to nutrient status of the parent materials, as other factors (moisture, climate, light, fire frequency, herbivory, etc.) are not noticeably dissimilar among these soils. Macro-nutrients such as

Ca and Mg are not lacking in these soils because of the abundance of dolomitic materials. The main non-organic sources of K and Fe, however, are the feldspar, mica and ferromagnesian mineral groups, all crystalline minerals whose abundance is not uniform among the soils. Crystalline clasts in the fine gravel fraction are most abundant in Rendolls and in Udorthents with thick, dark epipedons. Udorthents with abundant organic matter in the upper profile typically have in excess of 45 g kg^{-1} (weighted profile mean) crystalline gravels (pedons TU-1, TU-2, TU-3; Table 3). Conversely, Udorthents that either lack or have a thin, discontinuous A horizon contain less than 1 g kg^{-1} granitic or basaltic gravels in the profile (Schaetzl, 1991). Finally, as the feldspar/quartz ratios for the fine sand fraction increase, concomitant increases are observed in the thickness of the A (and A + O) horizons (Tables 2 and 3). Calculation of feldspar/quartz (F/Q) ratios from the raw data of Anderson *et al.* (1975a) also support this general finding. Their two Boroll pedons had weighted profile mean F/Q ratios of 0.54 and 0.22, whereas the two Boralf pedons studied had F/Q ratios of 0.20 and 0.11.

CONCLUSIONS

Although the small number of pedons sampled in this study does not allow rigorous statistical treatment of the data, simple correlation coefficients indicated that OC contents of the upper horizons of these gravelly, forested soils correlated with certain properties of the parent material. Weak, positive relationships occurred between weighted profile OC and contents of both clay and inorganic carbon. Strong correlation ($r = 0.93$) was observed between weighted profile OC and proportion of coarse fragments, even though the most gravelly soils have abundant, thick organic coatings on the gravels that are not included in traditional OC analyses. Had these organic coatings been included, the correlation might have been even stronger.

Large gravel and cobble amounts in these soils promote OC accumulation, possibly by decreasing available void spaces in a given thickness of soil. Gravelly soils will have less operable pedogenic volume in the upper profile than will less gravelly soils. This condition will cause organic matter to become more concentrated in the remaining space, other things being equal. Increased thickness of dark epipedons will also result because wetting fronts, and hence roots, may penetrate deeper in the more gravelly soils.

Finally, pedons that (i) contained a greater percentage of crystalline gravel clasts, and (ii) had higher feldspar/quartz ratios in the sand fraction, exhibited larger OC contents in upper horizons. This general relationship might result from the release of macronutrients during weathering of the crystalline gravels and feldspathic, sand-sized minerals, thereby increasing organic matter production in the forested ecosystem.

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REFERENCES

AFANAS'YEVA, Y.A. 1966. Thick Chernozems under grass and tree coenoses. *Soviet Soil Science,* 615–625.

AL-BARRAK, S. & LEWIS, D.T. 1978. Soils of a grassland–forest ecotone in eastern Nebraska. *Soil Science Society of America Journal* **42,** 334–338.

ANDERSON, J.U., BAILEY, O.F. & RAI, D. 1975a. Effects of parent materials on the genesis of Borolls and Boralfs in south-central New Mexico mountains. *Soil Science Society of America Proceedings* **39,** 901–904.

ANDERSON, J.U., SILBERMAN, D. & RAI, D. 1975b. Humus accumulation in a forested Haploboroll in south-central New Mexico. *Soil Science Society of America Proceedings* **39,** 905–908.

BERRY, M.E. 1987. Morphological and chemical characteristics of soil catenas on Pinedale and Bull Lake moraine slopes in the Salmon River Mountains, Idaho. *Quaternary Research* **28,** 210–225.

BIRCH, L.C. & CLARK, D.P. 1953. Forest soil as an ecological community with special reference to the fauna. *Quarterly Review of Biology* **28,** 13–36.

BOCHTER, R. & ZECH, W. 1985. Organic compounds in cryofolists developed on limestone under subalpine coniferous forest, Bavaria. *Geoderma* **36,** 145–157.

BROADBENT, F.E. 1953. The soil organic fraction. *Advances in Agronomy* **5,** 153–183.

BROERSMA, K. & LAVKULICH, L.M. 1980. Organic matter distribution with particle-size in surface horizons of some sombric soils in Vancouver Island. *Canadian Journal of Soil Science* **60,** 583–586.

DECONINCK, F., RIGHI, D., MAUCORPS, J. & ROBIN, A.M. 1974. Origin and micromorphological nomenclature of organic matter in sandy Spodosols. In *Soil Microscopy* (ed. G.K. Rutherford), pp. 263–280. The Limestone Press, Kingston, Ontario.

FENTON, T.E. 1983. Mollisols. In *Pedogenesis and Soil Taxonomy* (eds L.P. Wilding, N.E. Smeck & G.F. Hall), pp. 125–163. Elsevier, New York.

GAIKAWAD, S.T. & HOLE, F.D. 1965. Characteristics and genesis of a gravelly Brunizemic regosol. *Soil Science Society of America Proceedings* **29,** 725–728.

GEIS, J.W., BOGGESS, W.R. & ALEXANDER, J.D. 1970. Early effects of forest vegetation and topographic position on dark-colored, prairie-derived soils. *Soil Science Society of America Proceedings* **34,** 105–111.

GREENLAND, D.J. 1965. Interaction between clays and organic compounds in soils. Part II. Adsorption of soil organic compounds and its effect on soil properties. *Soils and Fertilizers* **28,** 521–532.

KILMER, V.J. & ALEXANDER, L.T. 1949. Methods of making mechanical analysis of soils. *Soil Science* **68,** 15–24.

KONONOVA, M.M. 1961. *Soil Organic Matter.* Pergamon Press, Oxford.

LARSEN, C.E. 1987. Geological history of glacial Lake Algonquin and the upper Great Lakes. *Bulletin of the United States Geological Survey* **1801.**

LEE, G.B., JANKE, W.E. & HORN, M.E. 1972. Characteristics and genetic relationships of soils and soil parent materials in the Valderan drift region of eastern Wisconsin. *Proceedings of the 15th Conference Great Lakes,* pp. 371–382. International Association for Great Lakes Research.

MCKEAGUE, J.A. 1971. In Organic matter in particle-size and specific gravity fractions of some Ah horizons. *Canadian Journal of Soil Science* **51,** 499–505.

MORTLAND, M.M. 1970. Clay–organic complexes and interactions. *Advances in Agronomy* **22,** 75–117.

NICHOLS, J.D. 1984. Relation of organic carbon to soil properties and climate in the southern Great Plains. *Soil Science Society of America Journal* **48,** 1382–1384.

NIMLOS, T.J. & TOMER, M. 1982. Mollisols beneath conifer forests in southwestern Montana. *Soil Science* **134,** 371–375.

NØRNBERG, P., DALSGAARD, K. & SKAMMELSEN, E. 1985. Morphology and composition of three Mollisol profiles over chalk, Denmark. *Geoderma* **36,** 317–342.

PROTZ, R. 1983. Calcium-dominated organans in humic Podzols from the Hudson and James Bay lowlands of Ontario (Canada). *Geoderma* **30,** 171–177.

REYNOLDS, J.W. 1977. *The Earthworms (Lumbricidae and Sparganophilidae) of Ontario,* Life Science Miscellaneous Publication. Royal Ontario Museum, Toronto.

ST ARNAUD, R.J. & WHITESIDE, E.P. 1964. Morphology and genesis of a chernozemic to podzolic sequence of soil profiles in Saskatchewan. *Canadian Journal of Soil Science* **44,** 88–99.

SCHAETZL, R.J. 1991. A lithosequence of soils in extremely gravelly, dolomitic parent materials, Bois Blanc Island, Lake Huron. *Geoderma* **46,** 305–320.

SHELDRICK, B.H. (ed.) 1984. *Analytical methods manual.* Land Resource Research Institute Contribution No. 84-30. Ottawa, Canada.

SMALL, T.W., SCHAETZL, R.J. & BRIXIE, J.M. 1990. Redistribution and mixing of soil gravels by tree uprooting. *Professional Geographer* **42,** 445–457.

SMITH, G.D., ALLAWAY, W.H. & RIECKEN, F.F. 1950. Prairie soils of the upper Mississippi valley. *Advances in Agronomy* **2,** 157–205.

SMITH, G.D. 1986. *The Guy Smith Interviews: Rationale for Concepts in Soil Taxonomy,* Technical Monograph 11. Soil Management Support Services, Cornell University, Ithaca, NY.

SOIL SURVEY STAFF. 1951. *Soil Survey Manual,* USDA Handbook 18. US Government Printing Office, Washington, DC.

SOIL SURVEY STAFF. 1975. *Soil Taxonomy,* USDA Handbook 436. US Government Printing Office, Washington, DC.

WALKER, P.H. 1966. Postglacial environments in relation to landscape and soils on the Cary drift, Iowa. *Iowa Agriculuture and Home Economics Experiment Station Bulletin* **549.**

WEBB III, T., & BRYSON, R.A. 1972. Late- and post-glacial climatic change in the northern Midwest, USA: quantitative estimates derived from fossil pollen spectra by multivariate statistical analysis. *Quaternary Research* **2,** 70–115.

WEBB III, T., CUSING, E.J. & WRIGHT, H.E. 1983. Holocene changes in the vegetation of the Midwest. In *Late-Quaternary Environments of the United States,* Vol. 2, *The Holocene* (ed. H.E. Wright, Jr), pp. 142–165. University of Minnesota Press, Minneapolis, MN.

YAIR, A. 1987. Environmental effects of loess penetration into the Negev Desert. *Journal of Arid Environments* **13,** 9–24.

YAIR, A. & LAVEE, H. 1985. Runoff generation in arid and semi-arid zones. In *Hydrological Forecasting* (eds M.G. Anderson & T.P. Burt), pp. 183–220. Wiley, New York.

YAIR, A. & BERKOWITZ, S.M. 1989. Climatic and non-climatic controls of aridity: the case of the northern Negev of Israel. *Catena Supplement* **14,** 145–158.

ZUMBERGE, J.H. & POTZGER, J.E. 1956. Late Wisconsin chronology of the Lake Michigan basin correlated with pollen studies. *Bulletin of the Geological Society of America* **67,** 271–288.

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