Alpine Soils on Mount Mansfield, Vermont, USA: Pedology, History, and Intraregional Comparison

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The highest summits in the northeastern United States rise above the tree line and support unique islands of alpine tundra. Little is known about the properties and history of these soils and it is unclear how soils compare between the separate, isolated alpine areas. As a step toward addressing this oversight, the physical and chemical properties of alpine soils were investigated on Mt. Mansfield, the highest mountain in Vermont. Soil thickness was determined through probing, while profile development and horizon properties were investigated in 31 excavations. Soil covers ~85% of the study area, with profiles averaging 18 cm thick (maximum of 58 cm). Most profiles demonstrate a variation on the sequence Oi, Oa and/or A, Bw or Bs (rare), AC, and Cr horizons. The Oa horizons contain significantly more C, Ca, K, and Mg than A horizons, and have higher cation exchange capacity and base saturation, and lower pH values. Almost one-third of the profiles were classified as Histosols, while nearly twice as many were classified as Entisols. Only two profiles contained cambic horizons and were classified as Inceptisols. Histosols contained significantly more K and organic matter than Entisols. A buried *Sphagnum* layer, radiocarbon dated to approximately 1000 yr ago, suggests a wetter climate at that time. Mount Mansfield soils are generally thinner than those described from nearby alpine environments and contain more organic matter with higher C/N ratios. Intraregional comparison suggests that carefully selected soil profiles could be combined to form litho-, climo-, and biosequences to elucidate controls on soil formation in these environments.

Abbreviations: BS, base saturation; CEC, cation exchange capacity.

PEDOLOGY

Despite their relatively low elevations, the highest summits in the northeastern United States rise above the tree line and support pockets of alpine tundra. These environments are remnants of an extensive tundra landscape that dominated the region immediately after deglaciation in the latest Pleistocene. Climatic amelioration into the Holocene and retreat of the Laurentide Ice Sheet displaced the ranges of tundra plants upward and northward until they were separated from the tundra environments of the Arctic (Zwinger and Willard, 1972; Miller and Spear, 1999). Today these pockets of alpine tundra are islands floating high above the typical forests of the northern Appalachians.

The total area of alpine habitat in the northeastern United States is approximately 34 km^2 (Kimball and Weihrauch, 2000). Almost a third of this area is located in the Presidential Range around Mt. Washington (1917 m) in New Hampshire, and almost a fifth of it is found on Mt. Katahdin (1605 m) in northern Maine. The remainder is scattered on lower summits in Maine (Longfellow Mountains), New Hampshire (White Mountains), Vermont (Green Mountains), and northern New York (Adirondack Mountains), shown in Fig. 1a. Despite their limited extent, these tundra communities have considerable value for habitat diversity and natural history; the plants germane to these environments are found nowhere else in the northeastern United States, and a few species [e.g., *Potentilla robbinsiana* (Lehm.) Oakes ex Rydb.] are found nowhere else in the world (Cogbill, 1993; Brumback et al., 2004).

The uniqueness of these environments has been recognized for more than a century, dating back to the pioneering work of Scudder (1874), Pease (1924), and Antevs (1932), and considerable attention has been focused on their botanical diversity (e.g., Bliss, 1963; Zwinger and Willard, 1972; Ketchledge and Fitzgerald, 1993). Alpine meteorology has also been studied extensively on Mt. Washington, where a weather observatory has been staffed continuously since 1932 (Mount Washington Observatory, 2007), and on Whiteface Mountain in New York (Atmospheric Sciences Research Center, 2007).

In contrast, very little work has been done on alpine soils in the northeastern United States, and much of it remains unpublished. Bockheim (1968) and Bockheim and Struchtemeyer (1969) studied soils on Saddleback Mountain in the Longfellow Mountains of western Maine and attempted to decipher relationships between soils and vegetation. Harries (1965) investigated the distribution of soils in the Presidential Range. In northern New York, Witty (1968) and Witty and Arnold (1970) mapped the distribution of soils on Whiteface Mountain and presented descriptions of soils above the tree line. Finally, Ketchledge (e.g., Ketchledge et al., 1985) studied alpine soils elsewhere in the Adirondack Mountains and devised strategies for stabilizing areas that were eroding under the assault of recreational impacts.

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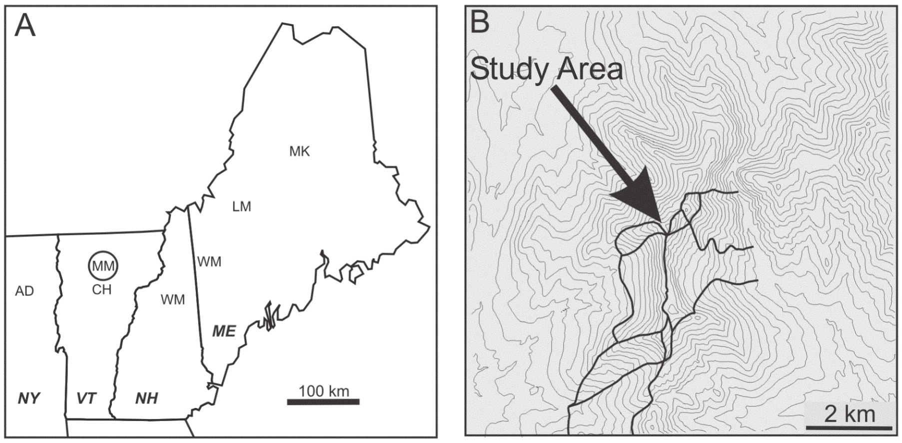

Fig. 1. Location of the Mt. Mansfield study site: (A) the northeastern United States with major alpine areas including Mt. Mansfield (circled MM) and Camels Hump (CH) in Vermont, the Adirondack Mountains (AD) in New York, the White Mountains (WM) in New Hampshire and Maine, and the Longfellow Mountains (LM) and Mt. Katahdin (MK) in Maine; (B) the topography of Mt. Mansfield (50-m contours) and the major trails on the mountain. The arrow points to the study area on the West Chin (area highlighted in Fig. 4). The elevation of the summit is 1337 m.

These studies remain valuable for the information they present about these poorly understood soils, yet many questions remain about alpine soils in this region. Moreover, no attempt has been made to present a unified understanding of soils from the isolated alpine zones in different states, despite the fact that close reading of the prior studies suggests the presence of intriguing differences. For instance, Witty and Arnold (1970) reported that Histosols were abundant on Whiteface Mountain, yet Bockheim and Struchtemeyer (1969) reported predominately mineral soils from Saddleback Mountain, and Bliss and Woodwell (1965) described two pedons with podzol-like characteristics on Mt. Katahdin. Furthermore, all of these studies were of a reconnaissance nature and presented details of only a few profiles from each area. As a result, little is known about the range of variability of these soils across more local scales.

Addressing these issues requires a dedicated program of field and laboratory investigations designed to elucidate the properties of soils in each of the major alpine areas of the northeastern United States. This study, which focused on Mt. Mansfield in the Green Mountains of Vermont, represents an initial step in this process. Mount Mansfield was selected because its summit ridge supports the largest alpine zone in the Green Mountains, and because alpine soils in Vermont have not been investigated previously, making it unclear how these soils compare with those documented from other northeastern alpine areas. The main objectives of the study were (i) to examine the morphology, properties, and distribution of alpine soils on Mt. Mansfield, (ii) to classify the soils using the U.S. Soil Taxonomy, (iii) to assess the age of the soils through radiocarbon dating, and (iv) to compare these soils with those reported from other alpine environments in the northeastern United States.

SETTING

Mount Mansfield is the highest mountain in Vermont, rising to an elevation of 1337 m. The mountain is comprised of an elongate north–south ridge paralleling the trend of the Green Mountains. The Laurentide Ice Sheet completely overran the mountain with erosive, warm-based ice, as evidenced by striations and glacial erratics on the summit ridge (Christman, 1959). The bedrock is predominantly chlorite- and mica-bearing schist of the Cambrian Underhill Formation (Christman, 1959).

Vegetation in the alpine zone is a mixture of *Carex bigelowii* Torr. ex Schwein. and *Vaccinium uliginosum* L. (Bowley, 1978). *Ledum groenlandicum* (Oeder) Kron & Judd is dominant in more poorly drained areas. Individual *Abies balsamea* (L.) Mill. and *Picea rubens* Sarg. trees with a krummholz growth form are locally present, although their abundance decreases steadily with distance above the tree line. The transition from continuous forest to *Vaccinium* heath and *Carex* meadow occurs at elevations as low as 1200 m on the northwestern shoulder of the mountain. This notably low tree line, anomalous even within the context of the northeastern United States, is apparently due to a combination of severe winter conditions, including pervasive icing (Ryerson, 1990), and edaphic factors.

The National Weather Service maintains a cooperative reporting station on the mountain. The mean annual temperature at the station during the period 1973 to 1992 was 1.4°C, the mean annual precipitation was ~1800 mm, and the mean maximum snow depth was 215 cm. It should be noted, however, that these measurements were taken within the uppermost subalpine forest, and are not necessarily representative of conditions in the alpine zone a few hundred meters higher (c.f. Bowley, 1978).

The study site for this project is located southwest of the main summit of Mt. Mansfield, primarily within the West Chin Protected Area (Fig. 1b). Access to the West Chin is restricted to authorized researchers to reduce impacts on the tundra vegetation. Elevations within the study site average 1310 m, aspects range from 45 to 315°, and slopes range from 0 to 35°.

METHODS

Field investigations involved the determination of soil thickness through probing, and excavation, description, and sampling of soil profiles. Given the delicate nature of the vegetation and the designation of the field site as a protected area, considerable effort was made to minimize the impacts of this field campaign. Work was conducted after vegetation was dormant for the winter and at times when the number of people on the mountain was at a minimum to reduce the possibility of luring curious hikers off trail. Walking directly on the tundra vegetation was avoided whenever possible, and wooden boards were used as a platform to distribute weight when work was focused on a specific area.

To assess the variability of soil thickness, probings were made every 10 paces along a rough series of southwest–northeast-oriented traverses with a 1-cm-diameter graduated steel rod. A total of 303 probings were made in two separate field visits, generating a pattern of depth measurements that overlaps in the western half of the study area. The distribution of probings extended across the entire alpine zone (~150 by 150 m), which is bounded by krummholz, forest, and cliffs. Traverses were focused on either side of the corridor occupied by a major hiking trail to obviate the possibility of sampling soils impacted by foot traffic. Because the pacing pattern was intended to be as objective as possible, some measurements were inevitably made on bedrock outcrops; these were scored as a soil thickness of 0 cm. Clasts suspended in the soil profile could be distinguished from underlying bedrock because clasts could be depressed slightly, yielding a softer impact. The location of each probing was recorded with a global positioning system (precision ±2 m), and the dominant vegetation type was noted by visual inspection.

At every 10th site, a small pit was excavated by shovel to bedrock; full-size soil pits were not permitted in the protected area, so excavations were limited to approximately 20 by 20 cm. For each pit, horizons were described and measured, bulk samples (∼100 g) were collected, and smaller known-volume samples were taken for determination of bulk density (Soil Survey Laboratory Staff, 2004, Method 3B6a). If the 10th site fell on a bedrock outcrop, the excavation was made at the next probing site presenting measurable soil. Altogether, 303 thickness measurements were made, 31 profiles were described, 53 horizons were sampled for chemistry, and 52 samples were taken for bulk density determinations.

In the lab, Munsell color was determined for each of the samples, and organic content was calculated through loss-on-ignition at 400°C to allow comparison with prior studies from the region (Soil Survey Laboratory Staff, 2004, Method 5A). Fifty-three samples, representing 93% of the mineral soil and sapric O horizons, were sent to the University of Vermont Agricultural and Environmental Testing Laboratory for analysis. Other horizons were too thin to sample and analyze. Soil chemistry was determined by inductively coupled plasma–Ar emission spectrometry on a per-weight basis following extraction in 1.25 mol L^{-1} NH_4OAc. The P content was determined colorimetrically. Soil pH was determined in 0.01 mol L^{-1} $CaCl_2$ (Soil Survey Laboratory Staff, 2004, Method 4C1a1a4). Total C and total N were determined using a CHN analyzer; none of the soils exhibited a reaction with dilute HCl so the total C measurement is considered equivalent to organic C. Given the high organic contents of the samples, soil texture was only determined for the 10 largest samples using the hydrometer method (Gee and Bauder, 1986) following digestion in H_2O_2. Six *Abies* needles recovered from near the base of one of the profiles, and a buried peat layer recovered from another, were submitted to Beta Analytic (Miami, FL) for AMS radiocarbon dating.

The locations of the 303 soil probings were imported into ArcGIS and soil thickness within the study area was displayed with a series of graduated symbols. The distribution of probings was assessed with a nearest neighbor analysis in ArcGIS. High–low clustering of soil profiles by depth and vegetation type was assessed with the Getis-Ord General G test, and spatial autocorrelation was assessed with the Moran's *I* test, also in ArcGIS. Mean thicknesses of the soils under the major vegetation types were compared using a *t*-test after logarithmic transformation.

On the basis of the lab results, Oa horizons were distinguished from A horizons using a C content threshold of 20% (determined by the CHN analyzer), allowing profiles to be classified into soil orders (Soil Survey Staff, 2006). Potential correlations between soil orders and vegetation were assessed with Pearson's χ^2 test. Spatial clustering of soil orders was assessed with the Getis-Ord General G and Moran's *I* tests in ArcGIS. Mean values of nutrients (in kg m^{-3}) for Oa and A horizons were compared using the Mann–Whitney *U* test. Mean values of nutrients in the two most common soil orders were also compared using the Mann–Whitney *U* test after weighting by horizon thickness. Unless otherwise noted, all statistics were calculated in SPSS 15.0 (SPSS, 2006). Radiocarbon dates were converted to calendar years with Calib 5.0 (Stuiver et al., 2005).

RESULTS

Soil Morphology

The majority of soils near the summit of Mt. Mansfield have a fairly simple and consistent sequence of horizons (Table 1). Most profiles demonstrate a variation on the sequence Oi, Oa and/or A, Bw (rare), AC, and Cr horizons. Boundaries between horizons are generally smooth and clear. Fibric (Oi) horizons up to 15 cm thick are present in each of the 31 profiles sampled. Depending on the vegetation, these horizons are predominantly composed of *Carex* stems or *Vaccinium* leaves. Underlying the Oi horizon in 18 profiles (58%) is an Oa horizon, averaging 10 cm thick. These horizons are generally black or very dark brown in color. Together, the Oi and Oa horizons are thick enough to constitute a folistic epipedon in 13 (42%) of the profiles (Table 1). High organic matter contents complicated determination of the particle size distribution in the Oa horizons, but the one sample successfully analyzed had a texture of loam (11% clay).

Buried fibric layers (Oib horizons) dominated by *Sphagnum* were encountered in two of the profiles (Table 1). One in thr Mans-03 profile returned a radiocarbon age of 1020 ± 70 ^{14}C yr (Beta-197219), which calibrates to between 880 and 1175 CE. *Abies* needles retrieved from the bottom of the Mans-04 profile, beneath the Oib horizon, dated to 830 ± 40 ^{14}C yr (1155–1275 CE, Beta-197220).

Twenty of the profiles (65%) contain A horizons averaging 13 cm thick. Colors are slightly lighter than those of the Oa horizons (generally values of 3 or 4 and hues from 5YR to 10YR). Soil structure is poorly developed and most samples are massive with abundant fine roots, although root content is markedly less than in the overlying horizons. The A horizons have textures of sandy loam to silt loam and contain from 5 to 15% clay.

Horizons reflecting more advanced pedogenesis are rare. Weak B horizons are present in two profiles (∼7%), one of which contains two Bs horizons with dark grayish brown (10YR 4/2) to dark yellowish brown (10YR 4/4) colors (Mans-02, Table 1). This profile also contains an elluvial (E) horizon 2 cm thick, hinting at incipient podzolization. This horizon has a brown color (7.5YR 5/2) that contrasts with the dark gray (7.5YR 4/1) A horizon above it and the dark yellowish brown (10YR 4/4) Bs1 horizon beneath it.

Four of the profiles (13%) contain AC horizons ranging from 4 to 19 cm thick (mean of 9 cm). These are found at deeper depths, and have higher color values than overlying Oa, A, or B horizons. In three of the four instances, the AC horizon rests directly on bedrock. In one profile (LMK070, Table 1), the AC horizon is underlain by an 8-cm-thick Oab horizon containing 30% C.

Finally, seven of the profiles (23%) contain a saprolitic Cr horizon (sensu Birkeland, 1984) averaging 6 cm thick. Colors vary widely from dark gray (5YR 4/1) to white (10YR 8/1), although the higher values reflect an abundance of mica derived from the underlying schist, and are not diagnostic of leaching. Clasts, usually of bull quartz with occasional orange-red staining, are present in Cr horizons and total >20% of the horizon volume in some profiles. The one Cr horizon analyzed has a silt loam texture with 15% clay.

Soil Classification

The alpine soils studied near the summit of Mt. Mansfield classify into three soil orders (Table 1). Ten profiles (32%) classify as Histosols because organic materials make up more than two-thirds of the depth to the lithic contact and the total thickness of the mineral horizons is <10 cm. Because the mean annual temperature at 50 cm in these soils is <8°C, but summer and winter mean temperatures vary by >6°C (unpublished data, 2007), these soils have a frigid thermal regime and classify as Lithic Udifolists (all are <50 cm deep).

The remainder of the soil profiles (n = 22) contain greater amounts of mineral material than the Histosols. Nineteen of these profiles (61%) belong in the Entisol order as Lithic Udorthents because they lack a cambic horizon and have a lithic contact within 50 cm of the surface. The remaining two profiles (6%) classify as Inceptisols because they contain evidence of weak profile development (Bw and Bs horizons). Under a frigid thermal regime, one classifies as a Typic Dystrudept (55 cm thick), while the other is a Lithic Dystrudept (45 cm thick).

Physical and Chemical Properties

Soil C content and bulk density exhibit a strong inverse relationship (Fig. 2). All Oa horizons have bulk densities <0.70 g cm^{-3}, whereas bulk densities of mineral horizons range up to three times that value. There is a fairly clear division between the C content of organic and mineral horizons, even with the arbitrary threshold of 20%. Three of the mineral soils, however, have C contents between 17 and 20% and appear more related to the organic horizons. There is no apparent clustering, and it is not possible to distinguish a unique C content–bulk density range for individual horizons.

All of the horizons are quite acidic, with pH values ranging from 4.2 to 2.5 (mean of 3.3). The lowest pH values are found in the Oa horizons, and the difference in mean pH between Oa (3.1) and A horizons (3.5) is highly significant (P = 0.002, Table 2). No consistent trend in pH values with depth is apparent in the mineral horizons; both the highest and lowest values were attained in A horizons. Cation exchange capacity (CEC) ranges from 35 to 9 cmol kg^{-1} and base saturation (BS) ranges from 14 to 2%; mean values of both are significantly (P < 0.01) greater in Oa horizons (Table 2). Oa horizons contain significantly more C, Ca, K, and Mg, while A horizons have significantly more Al, Cu, Fe, and S (Table 2).

Grouping soil chemistry data by profile and weighting by horizon thickness allows statistical comparisons between Histosols and Entisols (Table 3). Histosols have significantly greater loss-on-ignition values than Entisols (P = 0.015) and a nearly significant difference in C content (P = 0.052). Histosols also have significantly more K. The mean bulk density of Histosols is less than that of Entisols (0.5 vs. 0.7 g cm^{-3}),

Table 1. Morphological properties for 31 alpine soil pedons sampled on Mt. Mansfield, Vermont.

Profile	Horizon	Depth	Thickness	Total thickness	Munsell color (moist)	Epipedon	Order	Vegetation
		——— cm ———						
Mans-01	Oi	0–5	5	19	--	ochric	Entisol	*Carex*
	Oa	5–12	7		7.5YR 2.5/1			
	A	12–19	7		7.5YR 2.5/1			
Mans-02	Oi	0–7	7	55	--	ochric	Inceptisol	*Carex*
	A	7–12	5		7.5YR 4/1			
	E	12–14	2		7.5YR 5/2			
	Bs1†	14–24	10		10YR 4/4			
	Bs2†	24–45	21		10YR 4/3			
	Ab/C	45–50	5		7.5YR 3/2			
	Cr	50–55	5		10YR 4/2			
Mans-03	Oi	0–7	7	50	--	folistic	Entisol	*Vaccinium*
	Oa	7–10	3		7.5YR 2.5/1			
	Oib	10–30	20		7.5YR 5/3			
	Ab	30–50	20		7.5YR 3/2			
Mans-04	Oi	0–6	6	48	--	folistic	Entisol	*Abies*
	Oa	6–26	20		10YR 2/2			
	Oib	26–33	7		10YR 3/3			
	Ab	33–48	15		10YR 6/6			
LMK010	Oi	0–3	3	25	--	ochric	Entisol	*Vaccinium*
	A	3–13	10		7.5YR 2.5/1			
	Cr	13–25	12		10YR 8/1			
LMK020	Oi	0–5	5	30	--	folistic	Histosol	*Ledum*
	Oa	5–25	20		10YR 3/1			
	AC	25–30	5		2.5YR 3/1			
LMK030	Oi	0–14	14	28	--	ochric	Entisol	*Vaccinium*
	A	14–28	14		7.5YR 3/1			
LMK040	Oi	0–8	8	15	--	ochric	Entisol	*Vaccinium*
	A	8–15	7		5YR 2.5/1			
LMK050	Oi	0–6	6	25	--	folistic	Histosol	*Vaccinium*
	Oa1	6–20	14		10YR 2/1			
	Oa2	20–25	5		10YR 2/1			
LMK060	Oi	0–8	8	19	--	ochric	Entisol	*Vaccinium*
	A	8–15	7		7.5YR 3/1			
	AC	15–19	4		10YR 4/2			
LMK070	Oi	0–7	7	40	--	ochric	Entisol	*Vaccinium*
	Oa	7–13	6		10YR 2/1			
	AC	13–32	19		10YR 5/1			
	Oab	32–40	8		7.5YR 3/1			
LMK080	Oi	0–18	18	35	--	folistic	Histosol	*Ledum*
	Oa1	18–26	8		7.5YR 2.5/1			
	Oa2	26–35	9		7.5YR 2.5/1			
LMK090	Oi	0–12	12	28	--	ochric	Entisol	*Carex*
	A	12–28	16		7.5YR 2.5/1			
LMK100	Oi	0–9	9	45	--	ochric	Inceptisol	*Carex*
	A	9–19	10		10YR 3/2			
	Bw	19–36	17		10YR 4/2			
	AC	36–45	9		10YR 4/2			
LMK110	Oa	0–7	7	20	--	ochric	Entisol	*Carex*
	A	7–20	13		10YR 2/1			
LMK120	Oi	0–6	6	20	--	folistic	Histosol	*Vaccinium*
	Oa	6–15	9		10YR 2/1			
	Cr	15–20	5		5YR 4/1			
LMK130	Oi	0–10	10	26	--	ochric	Entisol	*Vaccinium*
	A	10–26	16		7.5YR 3/1			
LMK140	Oi	0–8	8	20	--	folistic	Histosol	*Carex*
	Oa	8–20	12		10YR 2/1			
LMK150	Oi	0–6	6	23	--	folistic	Histosol	*Vaccinium*
	Oa	6–17	11		10YR 2/1			
	Cr	17–23	6		10YR 4/2			
LMK160	Oi	0–8	8	19	--	folistic	Histosol	*Carex*
	Oa	8–18	10		10YR 2/1			
	Cr	18–19	1		10YR 3/2			
LMK170	Oi	0–8	8	22	--	folistic	Histosol	*Vaccinium*
	Oa	8–20	12		7.5YR 2.5/1			
	Cr	20–22	2		7.5YR 3/1			
05-010	Oi	0–6	6	20	--	folistic	Histosol	*Vaccinium*
	Oa	6–20	14		10YR 2/2			

Table 1 Continued.

Profile	Horizon	Depth	Thickness	Total thickness	Munsell color (moist)	Epipedon	Order	Vegetation
05-020	Oi	0–8	8	25	--	ochric	Entisol	*Vaccinium*
	A	8–25	17		10YR 4/2			
05-030	Oi	0–6	6	42	--	ochric	Entisol	*Carex*
	A1	6–14	8		10YR 3/1			
	A2	14–40	26		10YR 4/2			
	Cr	40–42	2					
05-040	Oi	0–5	5	20	--	ochric	Entisol	*Carex*
	A	5–20	15		10YR 4/2			
05-050	Oi	0–6	6	20	--	folistic	Histosol	*Vaccinium*
	Oa	6–20	14		10YR 2/2			
05-060	Oi	0–10	10	34	--	folistic	Entisol	Carex
	Oa	10–20	10		10YR 3/1			
	A	20–34	14		10YR 4/2			
05-070	Oi	0–4	4	23	--	ochric	Entisol	Vaccinium
	A	4–11	7		10YR 3/1			
	C	11–23	12		10YR 6/3			
05-082	Oi	0–4	4	18	--	ochric	Entisol	*Carex*
	A	4–18	14		10YR 3/2			
05-099	Oi	0–3	3	20	--	ochric	Entisol	*Carex*
	Oa	3–9	6		10YR 2/2			
	A	9–20	11		10YR 4/2			
05-100	Oi	0–5	5	23	--	ochric	Entisol	*Carex*
	Oa	5–13	8		10YR 3/2			
	A	13–23	12		10YR 3/2			

† Bs designation intended to acknowledge incipient podzolization; oxalate extraction not performed.

and Histosols tend to have greater CEC and BS than Entisols. All of these differences are notable, with *P* values near 0.1 (Table 3). Similarly, the greater available P content of Histosols (2.1 vs. 1.5 kg m^{-3}) is nearly significant ($P = 0.052$).

Soil and Vegetation Distribution

The network of soil probing reveals considerable detail about the range and distribution of soil thickness in the study area. Of the 303 sites, 258 had measurable soil, indicating that ~15% of the study area is bare rock. Excluding these areas, the mean soil thickness is 18 cm, with a range of 2 to 58 cm, a median of 16 cm, and a skewness of 1.41 (Fig. 3). The total distribution of soil probings is clustered ($z = -5.78$, $P = 0.000$), reflecting the presence of cliffs and krummholz patches that focused probings into soil-covered areas, and the overlap of two separate series of measurements in the western half of the study area. Within this distribution, deep soils are strongly clustered (Getis-Ord G = 334.16, $z = 3.79$, $P = 0.000$) and exhibit significant spatial autocorrelation (Moran's $I = 2.24$, $z = 3.18$, $P = 0.001$). These statistics confirm the visual impression that the deepest soils are located in isolated pockets (Fig. 4). Areas of shallow soils tend to be more spatially contiguous, especially in the northwestern quadrant of the mapping area, which contains the actual summit of the West Chin.

Vaccinium heath and *Carex* meadows are the most common vegetation communities, covering more than half of the study area (53%). *Vaccinium* is twice as abundant as *Carex* (37 vs. 16%), yet there is no significant difference in soil thickness under these two vegetation types ($t = 0.60$, $P = 0.550$, Fig. 5). The thickest soils are found under *Ledum* (mean of 56.8 cm, $n = 15$). Vegetation communities are randomly distributed with no high–low clustering (Getis-Ord G = 35.98, $z = 0.66$, $P = 0.509$) and no spatial autocorrelation (Moran's $I = -0.78$, $z = -1.1$, $P = 0.271$).

Soil profiles exhibit no significant clustering by order (Getis-Ord G = 0.03, $z = -0.07$, $P = 0.994$; Moran's $I = 0.03$, $z = 0.88$, $P = 0.379$). There is also no correlation between soil order and vegetation community (Pearson's $\chi^2 = 8.971$, $P = 0.175$).

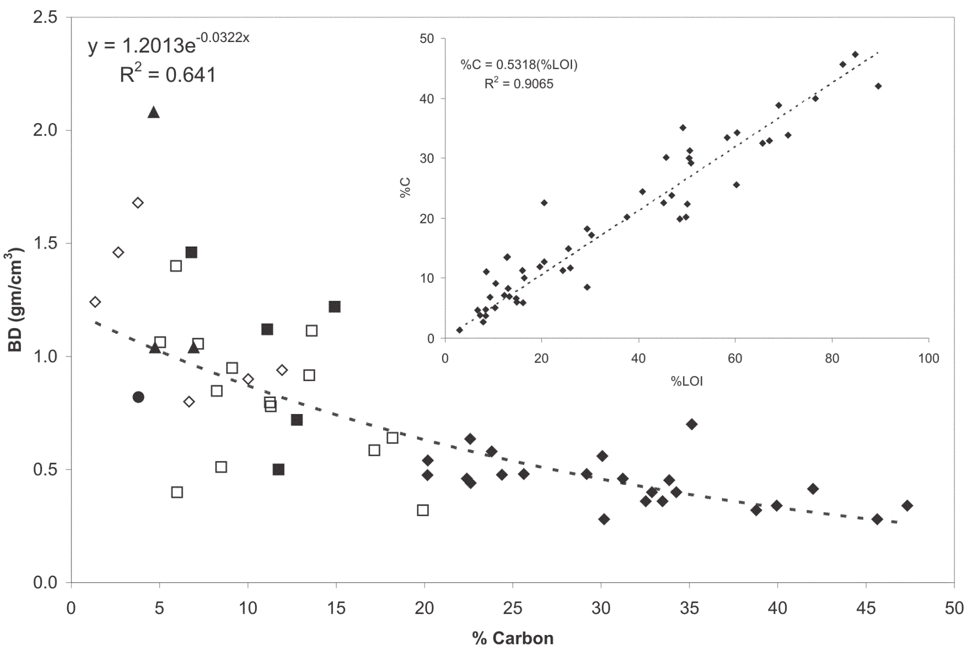

Fig. 2. Relationship between soil organic C content and bulk density (BD). Oa horizons are filled diamonds; A horizons, open squares; B horizons, filled triangles; E horizon, filled circle; AC horizons, filled squares; Cr horizons, open diamonds. Inset illustrates the strong linear relationship between organic matter content determined via loss-on-ignition and total C determined by a CN analyzer.

DISCUSSION

Controls on Pedogenesis

The main variation between the alpine soil profiles on Mt. Mansfield is the relative thickness and presence or absence of O and A horizons. All of the soils contain a high proportion of organic materials (up to 47% C), but the amounts vary considerably between and within profiles, and relatively minor differences determined the Histosol vs. Entisol classification. Overall, the major soil-forming mechanism in the Mt. Mansfield alpine zone appears to be paludification (the accumulation of organic matter), encouraged by cool temperatures and abundant moisture that combine to reduce rates of decay.

The presence of saprolite (Cr horizons) at the base of many soil profiles indicates that bedrock

Table 2. Properties of alpine soils on Mt. Mansfield arranged by horizon.

Horizon†	Statistic‡	Bulk density	pH	Available P	Al	B	Ca	Cu	Fe	K	Mg	Mn	Na	S	Zn	C	Loss-on-ignition	N	C/N	Cation exchange capacity	Base saturation
		g cm^{-3}		kg m^{-3}						kg m^{-3}							%	kg m^{-3}		cmol kg^{-1}	%
Oa (22)	mean	0.4	3.1	2.2	54.6	0.3	58.5	0.1	46.6	50.5	19.9	0.6	8.0	5.2	0.3	134.5	57.5	6.8	20.0	23.4	6.5
	σ	0.1	0.2	1.3	23.2	0.2	37.2	0.1	41.4	29.7	7.5	0.3	3.6	8.2	0.3	33.3	16.5	1.5	2.9	5.1	2.5
A (15)	mean	0.8	3.5	1.6	123.8	0.6	39.1	0.2	180.0	27.5	15.6	0.9	9.1	11.2	0.3	83.5	20.0	4.6	18.7	13.5	4.4
	σ	0.3	0.4	1.2	65.5	0.7	29.1	0.1	162.5	18.4	10.4	1.1	3.9	9.5	0.2	33.2	11.0	1.9	4.1	4.0	2.5
	U	44	65	108	48	124	91	91	79	79	96	131	148	88	100	39	10	59	116	22	70
	z	-3.7	-3.0	-1.7	-3.5	-1.2	-2.2	-2.2	-2.6	-2.6	-2.0	-0.9	-0.4	-2.3	-1.9	-3.8	-4.7	-3.2	-1.4	-4.4	-2.9
	P(2)	0.000***	0.002**	0.101	0.000***	0.257	0.028*	0.028*	0.009**	0.009**	0.042*	0.360	0.699	0.022*	0.057	0.000***	0.000***	0.001***	0.165	0.000***	0.004**
B (3)	mean	1.4	3.7	2.4	350.8	2.6	82.7	0.5	577.4	19.1	16.8	2.7	19.0	1.4	0.3	72.7	9.4	4.0	18.2	13.2	4.0
	σ	0.6	0.2	1.1	84.1	2.2	64.8	0.2	507.4	8.4	7.8	2.3	12.3	1.0	0.2	23.8	3.4	0.6	4.4	3.6	2.0
E (1)	mean	0.8	3.1	1.2	75.4	0.7	57.7	0.2	133.6	21.4	20.7	1.0	12.1	1.5	0.2	31.2	7.2	1.8	17.3	11.2	6.6
AC (5)	mean	1.0	3.3	1.7	163.4	1.3	62.3	0.2	295.2	18.4	14.9	2.4	9.9	0.8	0.1	111.2	17.9	4.9	22.7	13.6	4.4
	σ	0.4	0.4	0.4	157.6	1.2	24.7	0.2	291.7	5.7	2.0	2.8	7.3	0.2	0.1	46.0	8.5	1.9	2.2	3.0	1.1
Cr (7)	mean	1.2	3.5	1.3	144.5	0.4	57.9	0.2	75.2	22.9	16.2	1.3	12.7	1.9	0.2	53.5	11.6	2.9	18.0	11.6	4.6
	σ	0.3	0.4	1.0	102.2	0.4	33.6	0.2	79.7	16.5	8.5	1.5	6.7	2.8	0.2	39.4	6.3	1.9	3.3	2.2	1.4

* Significant at the 0.05 probability level.

** Significant at the 0.01 probability level.

*** Significant at the 0.001 probability level.

† Number of individual horizons in parentheses.

‡ Statistics from the Mann–Whitney U test provided for the Oa and A horizons only; P values are two-tailed.

Table 3. Comparison of weighted mean values between soil orders on Mt. Mansfield, Vermont.

Order†	Statistic	Thickness	Bulk density	pH	Available P	Al	B	Ca	Cu	Fe	K	Mg	Mn	Na	S	Zn	C	Loss-on-ignition	N	C/N	Cation exchange capacity	Base saturation
		cm	g cm^{-3}		kg m^{-3}						kg m^{-3}							%	kg m^{-3}		cmol kg^{-1}	%
Histosols (10)	mean	23.4	0.5	3.2	2.1	67.5	0.3	49.5	0.1	45.1	56.0	21.2	0.6	8.9	3.2	0.2	130.3	54.6	6.6	19.7	22.1	5.9
	σ	5.3	0.1	0.1	1.2	30.8	0.2	13.9	0.1	35.0	35.1	8.6	0.2	4.3	5.7	0.4	34.3	17.6	1.5	2.4	3.0	1.8
Entisols (19)‡	mean	27.5	0.7	3.3	1.5	82.1	0.4	37.5	0.1	104.1	29.4	15.6	0.5	7.6	7.9	0.3	92.6	33.8	4.9	19.2	17.5	4.8
	σ	10.4	0.3	0.3	1.1	46.9	0.4	22.7	0.1	107.9	21.9	7.5	0.3	3.0	9.7	0.2	45.7	22.0	2.4	3.8	7.1	1.9
	U	79	54	72	47	74	79	49	60	67	39	54	63	67	53	75	46	37	53	77	52	54
	z	0.71	-1.6	-0.7	-1.9	-0.6	-0.3	-1.8	-1.5	-0.9	-2.3	-1.6	-1.1	-0.9	-1.6	-0.6	-2.0	-2.4	-1.6	-0.4	-1.7	-1.6
	P(2)	0.239	0.127	0.537	0.052	0.570	0.786	0.074	0.204	0.386	0.018*	0.127	0.264	0.359	0.103	0.604	0.052	0.015*	0.103	0.711	0.103	0.115
Inceptisols (2)	mean	50.0	1.4	3.6	0.8	477.5	0.6	20.8	0.1	641.5	8.4	8.1	0.5	11.8	0.3	0.0	24.9	12.6	24.9	19.8	13.0	5.1
	σ	7.1	0.5	0.1	0.1	354.0	0.2	18.4	0.1	330.3	1.0	0.3	0.6	8.5	0.4	0.1	34.1	3.8	34.1	3.3	2.7	1.5

* Significant at the 0.05 probability level.

† Number of individual profiles in parentheses.

‡ Statistics from the Mann–Whitney U test provided for Histosols and Entisols only; P values are two-tailed.

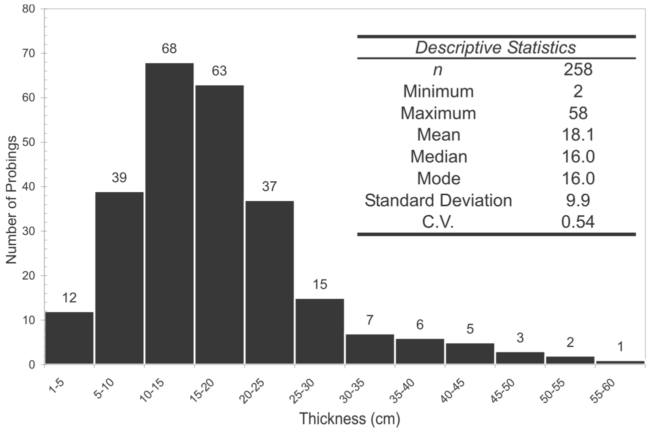

Fig. 3. Histogram of soil thickness in the study area. Soil profiles are generally thin over a lithic contact, but thicker soils are present locally.

weathering is occurring. Some of this weathering is probably driven by acidic litter produced by the *Vaccinium* and *Abies* krummholz. Furthermore, the cool temperatures common to the alpine zone increase the solubility of CO_2 in percolating soil water, increasing the concentration of H_2CO_3. Together, these processes enhance the rate and efficiency of hydrolysis reactions responsible for breaking down silicate minerals and generating clays (Schaetzl and Anderson, 2005). The acidity of the soils, therefore, may counteract the slow rates of chemical weathering conditioned by the cool temperatures.

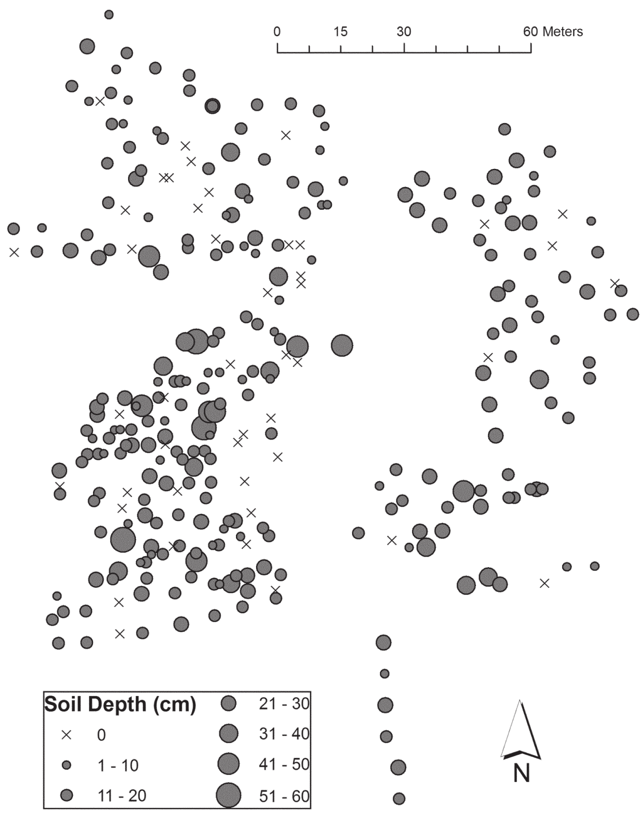

Fig. 4. Distribution of measured soil thickness in the study area based on 303 individual probings with a graduated steel rod. Deeper soils tend to found in isolated pockets, while shallower soils cluster in more contiguous areas. The central swath was intentionally skipped to avoid measuring soils potentially impacted by foot traffic along a major hiking trail.

Considering the five factors that influence soil development, climate (including local soil moisture conditions) and relief appear to exert the greatest influence in this setting (Jenny, 1994). Parent material and time (age) should be constant for all of the studied soils. The main vegetation communities form a complicated mosaic in the alpine zone, as evidenced by the lack of statistical clustering or autocorrelation; however, the lack of correlation between soil thickness or soil order and vegetation suggests that the vegetation assemblage is not a major control on soil development, perhaps because the vegetation above a given soil profile is changing in response to disturbance or succession at a rate faster than pedogenesis. Soil profiles may, therefore, reflect the integrated history of the overlying vegetation.

The role played by the combination of climate and relief is revealed by the spatial distribution of soils in the study area. The network of deeper soil pockets, revealed by the probings (Fig. 4) and spatial statistics, is far more intricate than the relief of the soil surface. Thus, the mantle of soil is smoothing a topographically complicated bedrock substrate. This inference is validated by observation of exposed bedrock ledges elsewhere along the summit ridge, which exhibit pronounced relief at meter and submeter scales, reflecting well-developed schistosity and complex folding. Because many of these ledges were covered by soil until the escalation of recreational impacts during the past few decades, this microrelief probably continues under the modern soil mantle. Soils filling closed depressions or swales in the bedrock should retain moisture longer than those located on bedrock highs, and increased soil moisture reduces rates of organic matter decomposition, encouraging paludification and the accumulation of thicker organic deposits. Viewed this way, the map of soil thickness may reveal less about soil formation and more about the underlying bedrock topography.

Relevant results were presented by Certini et al. (2002) from a study of bedrock weathering within closed cavities generated by removal of xenoliths in north-central Italy. Water trapped within these cavities becomes acidified by decaying organic matter, driving cavity enlargement. In the larger cavities, clay minerals have formed in addition to thick accumulations of organic matter, demonstrating a direct linkage between bedrock weathering and soil formation. A similar process may be occurring in bedrock depressions on Mt. Mansfield.

It is worth noting that bedrock relief and its control on local soil moisture cannot be the only process responsible for soil development on Mt. Mansfield. If pedogenesis proceeds more rapidly, or more effectively, in soils overlying bedrock depressions, then the deepest soils should exhibit the greatest development, particularly since the position of the studied soil profiles on the mountain summit precludes the accumulation of colluvium. The two profiles classified as Inceptisols, reflecting the presence of an incipient cambic horizon, were among the thickest encountered in the study area. If the presence or absence of saprolite is considered a benchmark of soil development (as an indication of advanced weathering of the underlying schist), however, then the mean thickness (29 cm) of soils containing saprolite is not significantly greater than the thickness of soils lacking Cr horizons (27 cm, $P = 0.417$). Thus, while bedrock irregularities may facilitate the formation of deeper soils, deeper cavities don't necessarily lead to more rapid pedogenesis. This situation may reflect more frequent water saturation in deeper bedrock depressions that could inhibit weathering reactions requiring aerobic conditions. Deeper, wetter soils may also remain frozen for longer in the spring, reducing both the time available for chemical weathering and the rates at which these reactions proceed.

Soil Age and History

The two radiocarbon dates, which overlap at approximately 1175 CE, indicate that alpine soils on Mt. Mansfield have been stable through environmental changes such as the Medieval Warm Period (ca. 900–1300 CE) and the Little Ice Age (ca. 1350–1850 CE) (e.g., Bradley and Jones, 1993; Hughes and Diaz, 1994), and at no time in the past millennium were the locations of these two profiles reduced to bare bedrock by erosion. This result raises the possibility that these soils could record past environmental changes in the alpine zone.

The most obvious indication that conditions have varied with time comes from the Oib horizons encountered in two of the profiles (Mans-03 and -04). These horizons were composed almost entirely of *Sphagnum* fibers (probably *S. magellanicum* Brid.), and both were found within deep profiles located ~5 m apart. *Sphagnum* is not present at the surface of these sites today, so the presence of *Sphagnum* in the subsurface indicates that approximately 1000 yr ago the site of these profiles resembled the alpine bogs along the summit ridge described by Bowley (1970), or the modern blankets of *Sphagnum* and *Ledum* present elsewhere within the study area. Existence of a bog at this site would require a higher than modern water table,

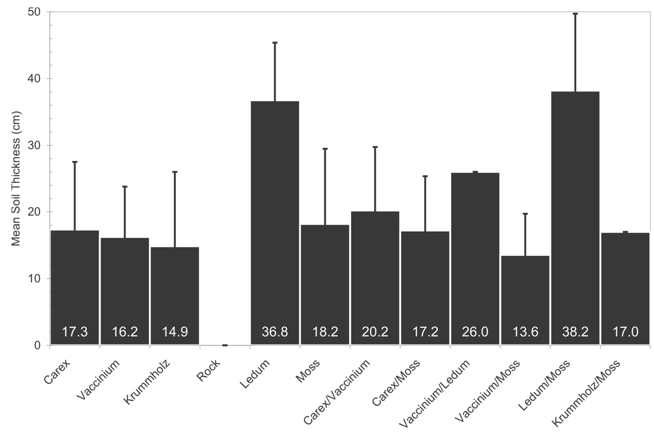

Fig. 5. Mean soil thickness (with 1 σ error bars) arranged by vegetation community. *Carex* and *Vaccinium* are the two most dominant vegetation types, making up nearly 50% of the study area; however, there is no significant difference in the mean thickness of their underlying soils.

and because there is no geomorphic mechanism that could have allowed the site to become better drained during the past 1000 yr, the climate approximately 1000 yr ago must have been wetter. An alternative explanation is that this site was occupied by a bog until approximately 1000 yr ago when the accumulation of organics became thick enough to raise the soil surface above the water table, facilitating a conversion from bog vegetation to *Vaccinium*, *Carex*, and *Abies*. Either way, the presence of A horizons resembling those found in other profiles beneath the

Table 4. Comparison of Mt. Mansfield alpine soils with others described from the northeastern U.S.

Study†	Profile	State	Modern horizonation	Thickness	Loss-on-ignition	pH	Bulk density	C/N	Original classification
				cm	%		g cm-3		
A	His	VT	Oi/Oa/Cr	30–19	90–35	3.5–2.9	1.46–0.32	26.0–15.1	Histosol
	Ent	VT	Oi/Oa/A/Cr	50–15	85–12	4.2–2.5	1.24–0.28	26.7–11.2	Entisol
B	1T-1	ME	Oi-Oa-A-Bhs-R	23	28–14	4.2–3.8	1.09–1.08	--	Alpine Turf
	2T-1	ME	Oi-Oa-A-Bhs1-Bhs2f-R	61	25–17	4.5–3.5	1.31–0.88	--	Alpine Turf
C	1	NY	Oi-A-E-Bh1-Bh2-Bs1-Bs2-C-R	89	37–2	4.9–4.2	1.66–0.29	30.6–7.6	Typic Haplorthod
	2	NY	Oi-A1-A2-A3-Bh1-Bh2-C-R	79	30–4	4.7–3.9	1.38–0.32	--	Typic Haplorthod
	21	NY	Oi1-Oi2-Oi3-Oa1-Oa2-Bh-R	5	93–15	4.7–4.2	0.82–0.09	--	Lithic Dysleptist
	22	NY	Oi1-Oi2-Oi3-Oa1-Oa2-Oa3-Oa4-Bh-R	8	92–20	5.0–4.1	0.70–0.07	41.7–19.3	Humodic Dyssaprist
D	1	NY	Oi1-Oi2-Oa1-Oa2	38	93–53	4.1–3.7	0.47–0.10	31.2–24.7	Lithic Borofolist
E	1	NH	A-Cr	90	39–1	5.0–3.9	--	22.1–9.6	Ranker with dust humus
	2	NH	A-Cr1-Cr2	90	22–1	5.1–4.3	--	21.7–10.7	Ranker with dust humus
	4	NH	A-Cr	90	21–1	5.2–4.5	--	19.7–11.2	Ranker with dust humus
	5	NH	A1-A2-B-Cr	100	22–1	5.1–4.5	--	19.6–0.7	Ranker with dust humus
	7	NH	A-ACr	64	20–1	5.0–4.5	--	18.7–1.0	Ranker with dust humus
	8	NH	A-Cr	100	4–1	5.1–4.4	--	18.6–0.8	Ranker with dust humus
	9	NH	A-Cr	75	33–1	4.9–4.1	--	17.5–0.8	Ranker with dust humus
	11	NH	Oe-Oa-A1-A2-Cr	90	78–1	5.0–3.7	--	23.5–5.0	Ranker raw over dust
	12	NH	Oe-A-B-Cr	90	74–1	4.8–3.6	--	18.6–11.2	Podzol-Ranker
	13	NH	Oe-Oa-A1-A2-Cr1-Cr2	90	85–0	4.8–3.4	--	23.5–12.6	Podzol-Ranker
	14	NH	Oe-A-E-Bh-Bs-BsCr-Cr1-Cr2	75	36–0	5.1–3.7	--	18.1–10.2	Podzol
F	SM	NH	A-B-C	91	49–2	4.9–4.1	--	24–16	--
	SRDSH	NH	A-B-C	61	46–2	4.7–4.1	--	29–19	--
	DSHR	NH	A-B-C	40	26–2	4.3–4.1	--	25–20	--
	DSH	NH	A-E-B-C	70	49–6	4.6–4.0	--	28–24	--
G	SH	ME	Oi-Oa-A-Bs1-Bs2-Bs3-C	64	8–3	4.5–3.9	--	--	Podzol

† A, this study–range of values in profiles classified as Histosols (His) and Entisols (Ent); B, Bockheim & Struchtemeyer (1969); C, Witty (1968); D, Witty and Arnold (1970); E, Harries (1965); F, Bliss (1963); G, Bliss and Woodwell (1965).

Oib horizons argues that water-saturated conditions at this site approximately 1000 yr ago represent a transient departure from a normal, somewhat drier, late-Holocene state.

Other evidence for past climate and vegetation changes is more difficult to discern in the Mt. Mansfield alpine soils. For instance, as noted above, the study site is only marginally above the modern tree line, and scattered *Abies* and *Picea* krummholz are present almost to the summit. If the tree line were higher during the Medieval Warm Period, then coniferous vegetation should have left evidence of leaching, especially considering the dominance of Spodosols beneath lower elevation forests on Mt. Mansfield (Soil Conservation Service, 1974). Only one soil (Mans-02) contains an E horizon, however, and only one other profile (LMK100) displayed obvious horizonation reflecting pedogenic translocation. This observation may indicate that the tree line has been stable during the past millennium, perhaps because exposure to winter ice and wind would limit upward growth even with warmer summer temperatures. Alternatively, the tree line may have been higher in the past but the morphologic evidence (i.e., E horizons) in the soils has been obscured or overprinted by the large amounts of organic matter. Attempts to investigate the composition of past vegetation communities through analysis of $\delta^{13}C$ values of soil organic matter (e.g., Tieszen et al., 1997) were hampered by the similar $\delta^{13}C$ values of the C_3-type plants common to this environment (unpublished data, 2006). As a result, evidence for past incursions of conifers into the modern alpine zone remains elusive.

Overall, the best evidence constraining the magnitude of past vegetation changes on Mt. Mansfield may be found in the vegetation itself. Intermingled with the more pedestrian *Carex* and *Vaccinium* species are true Arctic–alpine plants like bearberry willow (*Salix uva-ursi* Pursh) and diapensia (*Diapensia lapponica* L.). If trees reached the summit of the mountain at some point in the post-glacial period and extirpated these exotic species, it seems unlikely that they would have returned given that the nearest higher elevation tundra environments are located $\sim$100 km away (see Fig. 1a). More work is necessary to fully investigate the utility of soil archives as recorders of past climate changes in this environment.

Regional Comparison

The data set developed in this study allows alpine soils on Mt. Mansfield to be compared with those described from other alpine zones in the northeastern United States. Table 4 presents abbreviated descriptions of profiles from the published and unpublished literature along with summary descriptions for the Histosols and Entisols on Mt. Mansfield. Because the U.S. Soil Taxonomy has evolved since completion of the older studies, horizon designations have been converted to their modern equivalents for ease of comparison. The original classification of each of the soils was retained, however, for clarity and historical continuity.

The soils studied on Mt. Mansfield are most similar to the "alpine turf soils" of Bockheim and Struchtemeyer (1969), the "Ranker with dust humus" of Harries (1965), and the "sedge meadow" of Bliss (1963). All of these profiles resemble the classical "Ranker" soils described in some European soil taxonomic systems (e.g., Avery, 1973) as thin organic layers over silicate bedrock. Ranker was also included in the 1974 version of the FAO world soil classification (FAO-UNESCO, 1974), although these soils are now classified as Leptosols (FAO, 1988).

One commonality between all of the individual study areas is the lack of evidence for till or other erratic-laden glacial deposits as a soil parent material; all of the mineral soils appear to be forming in weathered bedrock. This result is somewhat surprising given the abundant evidence for ice passing over these mountains during the last glaciation (Christman, 1959). If glacial sediments were originally deposited on these summits, mass wasting and slopewash during the period immediately following deglaciation must have removed the glacial cover from the higher elevations, leaving nothing but bare bedrock (Miller and Spear, 1999).

Further study of the data in Table 4 reveals that the thickest Mt. Mansfield soils are shallower than those reported from Maine and New Hampshire, where soil profiles were excavated to depths of >50 cm. On the other hand, the Mt. Mansfield soils fall between the extremes reported from Whiteface Mountain where thick (>70-cm) organic soils were noted (Witty, 1968). The import of these discrepancies is unclear, but given the similarity in the vegetation and climate between the two locations, it is possible that the difference reflects a contrast in bedrock weatherability.

All of the reported bulk density measurements for alpine soils from the region are consistent and uniformly low, except in the deepest, most mineral-dominated horizons. The lowest bulk density values were found in the Folists described from Whiteface Mountain (Witty, 1968; Witty and Arnold, 1970).

The Mt. Mansfield soils contain more organic matter than other alpine soils in the region except for the Folists from Whiteface. This difference is particularly notable when considering the Rankers from Mt. Washington (Harries, 1965) and the Podzol on Mt. Katahdin (Bliss and Woodwell, 1965), which are thicker than the profiles measured on Mt. Mansfield. The maximum C/N ratios reported from Mt. Mansfield are also greater than those measured in the Mt. Washington soils, perhaps reflecting a difference in the vegetation communities. The minimum C/N values on Mt. Mansfield are less than the minima for Folists on Whiteface Mountain, indicating that the organic soils in the latter study site are less decomposed.

Intraregional comparison also reveals intriguing differences that warrant further investigation. For instance, Bockheim (1968) and Bockheim and Struchtemeyer (1969) described "incipient Podzols" under *Abies balsamea* krummholz on Saddleback Mountain, and attributed the greater eluviation to the acidity of litter derived from this vegetation. Similarly, Bliss and Woodwell (1965) reported a Podzol under *Vaccinium* on Mt. Katahdin, and Harries (1965) noted Podzols and Podzol–Rankers under heath on Mt. Washington. This study did not specifically target soils developed under krummholz, but 37% of the Mt. Mansfield study area is vegetated by *Vaccinium*, thus it surprising that more evidence of podzolization was not noted. The difference may lie in the climate of these environments, as the higher summits in New Hampshire and Maine are wetter and cooler. Alternatively, perhaps the schist parent material for soils on Mt. Mansfield differs from the schist on Mt. Washington—it obviously differs from the granite bedrock underlying Saddleback Mountain and Mt. Katahdin—producing a soil less prone to podzolization.

Other evidence for the influence of bedrock lithology on alpine pedogenesis is noted in the southern Presidential Range of New Hampshire, where Harries (1965) described several profiles he classified as "Braunerde." Compared with the Rankers, these profiles contain more evidence for Fe translocation, brighter horizon colors, and more highly weathered clasts. Harries (1965) concluded that the gneissic bedrock in this region was more easily weathered than the schist underlying Mt. Washington farther north. Together these intraregional differences raise the possibility that carefully selected soils from these different alpine areas could be used to construct climo-, bio-, or lithosequences that would be useful in illuminating controls on pedogenesis in these environments.

CONCLUSIONS

Field and laboratory analysis reveals considerable information about the properties and history of the highest soils in Vermont. Soils located above the limit of continuous trees on Mt. Mansfield are thin, acidic, and rich in organic matter. Most profiles classify as either Histosols or Entisols depending on the thickness of the organic horizons. Histosols contain significantly more K and C than Entisols. They also have lower bulk densities and greater CEC, BS, and available P. All of the soil profiles terminate at a lithic contact, and the total solum thickness is strongly controlled by the submeter-scale topography of the underlying bedrock. There is no significant difference in the mean thickness of soil profiles under the two major vegetation types, *Vaccinium* and *Carex*, or clustering in the distribution of soil orders and vegetation communities. Saprolitic Cr horizons at the base of one-quarter of the profiles indicate that bedrock weathering is occurring, probably due to the low pH values. Radiocarbon dating indicates that the soils have been stable across time scales from 100s to at least 1000 yr, and buried organic horizons reveal that climate changes have impacted the balance between precipitation and evaporation in the alpine environment. While alpine soils on Mt. Mansfield resemble those described from the highest mountains in the other northeastern states, intraregional differences hint at the interplay between bedrock, climate, and vegetation in controlling alpine pedogenesis. Further research will be directed at elucidating these relationships.

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