Genesis of Silty and Clayey Material in Some Alpine Soils in The Teton Mountains, Wyoming and Idaho

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Introduction

There is some difference of opinion as to whether fine-grained soil material in alpine areas of mountain ranges in the western United States is primarily the result of in situ weathering or of deposition of eolian material from other sources. Birkeland (1) and Williams (14) consider some silt and clay-rich material in the alpine areas of the Rocky Mountains to be loess. Retzer (11) and Mahaney (8) consider such fine-grained material to be primarily the result of in situ weathering. Nimlos and McConnel (10) found large percentages of silt in surface horizons of alpine soils in Montana, but considered this to be due to overestimation by the laboratory methods used. This investigation was undertaken to help provide additional data on this subject.

Most of the Snake River Plain is southcentral and southeastern Idaho is known to be covered by a thick mantle of loess or lacustrine material with characteristics similar to loess, so it is reasonable to expect that loess may be a significant part of the soil profile in the mountains east of the plain. Blackwelder (2) observed exposures of loess twenty feet deep on the forested portion of the western slope of the Teton Mountains in southeastern Idaho and northwestern Wyoming and found that its chemical composition was similar to loesses of the Missouri and Mississippi valleys. Lewis et al. (6) in a study of the distribution of loesses in Idaho concluded that the maximum elevation of loess in the Middle Rocky Mountains Province to the east of the Snake River Plain is 2135 m. More recently, Lund et al. (7) have concluded that most of the soils on the Snake River Plain of southcentral Idaho studied by Lewis et al. (6) are of lacustrine origin with only the surficial material being of possible loessial origin.

A soil survey has been published that covers the thicker loess soils of the Teton Basin (3), but there has been little detailed investigation of the characteristics of the soils formed on the westward-dipping Paleozoic sediments on the western slope of the Teton Mountains. Much of this area is above timberline and those areas above glacial valleys were affected by periglacial processes throughout much of the Pleistocene. Are the soil materials of these surfaces partly eolian in origin? If so, what is their relationship to the thick deposits of silty material in the intermountain basins west of the range? Field work during the summer of 1973 in the vicinity of Teton Canyon, Wyoming tested the hypothesis that surface material in soils on the west slope of the Teton Mountains is primarily eolian in origin. Specifically, we hypothesized that the particle size distribution and mineralogy of the mountain surface soils would show a strong similarity to loess soils in the basin to the southwest and contrast with the residuum of the diverse bedrock lithologies upon which these soils lie.

Methods and Description of the Study Area

Soil profiles along a 50 km SW-NE transect crossing four physiographic units (basin, foothills, subalpine and alpine) were sampled (Table 1). Sample sites ranged in elevation from 3200 m near the headwaters of Teton Creek in Wyoming to 1880 m in Swan Valley, a narrow loess-filled basin crossed by the Snake River before it

TABLE 1. *Description of sample sites of soils in Teton area, Wyoming and Idaho*[1].

Site No.[5]	Depth sampled (cm)	Sand %	Silt %	Clay %	Thickness of silty material (cm)	Underlying material	Altitude (m)	Physiographic unit & hillslope position
1(A)	0-10	6.6	643.3	29.1	900 +	?	1880	basin/shoulder
(A)	900	5.5	58.8	35.7				
2(A)	3-24	6.1	60.5	33.4	440	till or outwash	1960	basin/shoulder
(A)	360-400	8.7	65.6	25.7				
3(A)	3-10	6.0	68.8	25.2	76	Quaternary tuff	2200	foothills/backslope
(D)	76-118	9.8	56.8	33.4				
4(A)	0-4	6.3	60.2	33.5	46 +	Madison Ls	2485	foothills/summit
5(B)	0-11	8.5	51.6	36.9	44	Terra rossa/ Madison Ls	2430	foothills/backslope
(D)	44-296	22.7	22.8	54.5				
6(B)	0-45	9.6	44.7	45.7	45 +	Bighorn Dolomite	2790	subalpine/summit
7(B)	0-63	14.7	44.7	40.6	63 +	Bighorn Dolomite	2825	subalpine/toeslope (sinkhole)
8(D)	surface	12.3	47.9	39.8	—	Park Shale	2750	subalpine/backslope (outcrop)
9(B)	0-35	6.2	53.5	40.3	135 +	Death Canyon Limestone	2745	subalpine/toeslope (sinkhole)
(B)	70-100	4.6	52.9	42.5				
10(A)	0-30	6.5	65.0	28.5	63 +	Death Canyon Limestone	2800	subalpine/shoulder
(A)	30-63	10.1	60.8	29.1				
11(C)	0-10	32.3	38.5	29.2	10	Granite[2]	2800	alpine/footslope
(D)	25-54	71.2	13.8	15.0				
12(A)	surface	4.7	67.9	27.4	—	—[3]	2945	alpine
13(C)	10	7.8	62.5	29.7	20	Flathead[4] Quartzite	3100	alpine/shoulder
(D)	44	30.1	42.5	27.4				
14(D)	50	46.3	34.0	19.7	—	Flathead Quartzite	3200	alpine/summit (outcrop)

[1] The following additional data is available from the authors, on request: a) detailed pedon descriptions for most sample sites, b) tables containing partiile size data for all USDA and silt fractions, percent moisture, organic matter and carbonate for 59 samples from 18 samples sites, and c) a location map for all sample sites on a USGS topographic map base scale 1: 24000

[2] Turf covered lobe of a rock glacier of Pinedale age.

[3] Silt accumulation on a snow bank.

[4] Stripe of fine material between stone stripes.

[5] Letter in parenthess indicates sample group as follows: (A) basin loess and other surface horizon poresumed to be primarily eolian in origin, (B) surface horizons reflecting mixing of colian material and residual clays, (C) surface horizons reflecting mixing of colian material and coarse-grained residuum, (D) subsurface horizon where bedrock residuum is the predominant material.

opens onto the broader Snake River Plain of southeastern Idaho. Prevailing winds in the area are from the southwest and the Swan Valley site lies upwind from the alpine study area.

The physiographic units were separated on the basis of topography and vegetation (Figure 1). The basin unit (1830-1980 m) is characterized by rolling topography and a thick mantle of loess. Sagebrush is the dominant natural vegetation. In the drier parts of the basin unit soils formed in the loess are Cryoborolls with cambic horizons. The foothills unit (1980-2440 m) lies on the broad dipslope of the Mississippian-age Madison Limestone. Valleys in this unit are generally filled with glacial deposits of Bull Lake and Pinedale age. Small areas of Quaternary tuff also are present. The natural vegetation is fir and pine on north-facing slopes and sagebrush and mountain mahogany on south-facing slopes. Precipitation is greater on the foothills unit and the soils usually have argillic horizons with Cryoborolls under sagebrush-grass vegetation and Cryoboralfs under forest vegetation. Soils on glacial moraines are Cryochrepts formed under coniferous forests. The subalpine unit (2440-2800 m) lies mainly on karst benches of the ordovician Bighorn Dolomite and Cambrian Death Canyon Limestone. Stands of spruce, fir and pine exist in protected parts of the benches, but sparsely vegetated areas of limestone rubble and sinkholes predominate. The alpine unit (2800-3365 m) lies on the dip slope of the Cambrian Flathead Quartzite. Pre-Cambrian gneisses and granites are exposed in cirques and major valley bottoms. Alpine meadows, bogs and tundra predominate. Frost features such as stone stripes are common. Soils with thin cambic horizons can be found in the alpine and subalpine areas, but usually there is no evidence of pedogenic weathering. A horizons are also commonly absent in these areas.

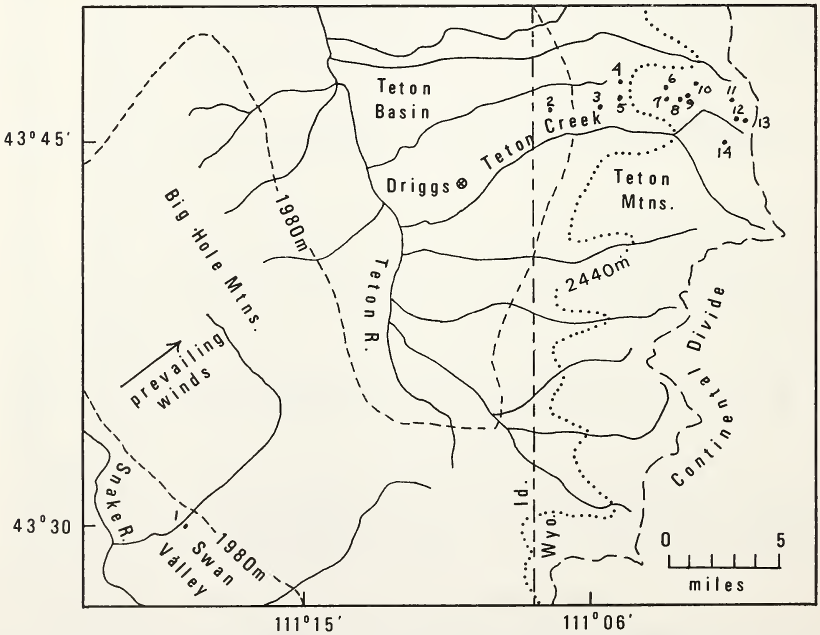

FIGURE 1. *Physiographic units and sample locations, Teton area, Wyoming and Idaho. Dashed line indicates boundary between basin and foothills units; dotted line is the boundary between the foothills and subalpine units.*

A complete profile description was made at most sample sites (see footnote 1, Table 1). The A horizon (or upper 10 cm when no A horizon was present), B horizon (generally present only in the basin and foothills units) and, wherever possible, C horizons formed in material weathered from the underlying bedrock were sampled. In the subalpine and alpine areas often no horizonization was evident in the silty material (sites 6,7,9,10 and 13), so arbitrary sampling depths were chosen.

Particle size distribution was determined by sedimentation methods developed by Kilmer and Alexander (5). Heavy mineral separates of very fine sand (100 to 250 microns) from selected samples were mounted on slides using balsalm and were studied with the aid of a petrographic microscope. Identification and tabulation of heavy minerals followed the techniques and mineral descriptions outlined by Milner (9). Calculation of mineral percentages was based on counts of 200 to 400 grains (Table 2).

Clay mineralogy was determined on selected samples by x-ray diffraction with Cu target at the Department of Geology, University of Wisconsin-Madison (Table 2). Samples were prepared by mounting a slurry of the less than 2 micron fraction on warm porcelain tiles using an eyedropper. Quantitative analysis of the relative amounts of different minerals in the clay fraction was not attempted.

Results and Discussion

The surface and C horizons of the soils sampled along the transect are dominated by silt and clay (Figure 2). For comparison, Retzer's (11) data for A and C horizons over similar parent materials (limestone, shale, quartzite and granite) in the Colorado alpine are also plotted. The contrast is striking. The soil surface material seems to fall into two distinct classes in terms of particle size distribution. The first group (Figure 3a) has curves that compare well in terms of shape with the midwestern Roxana loess but contain significantly higher percentages of clay.

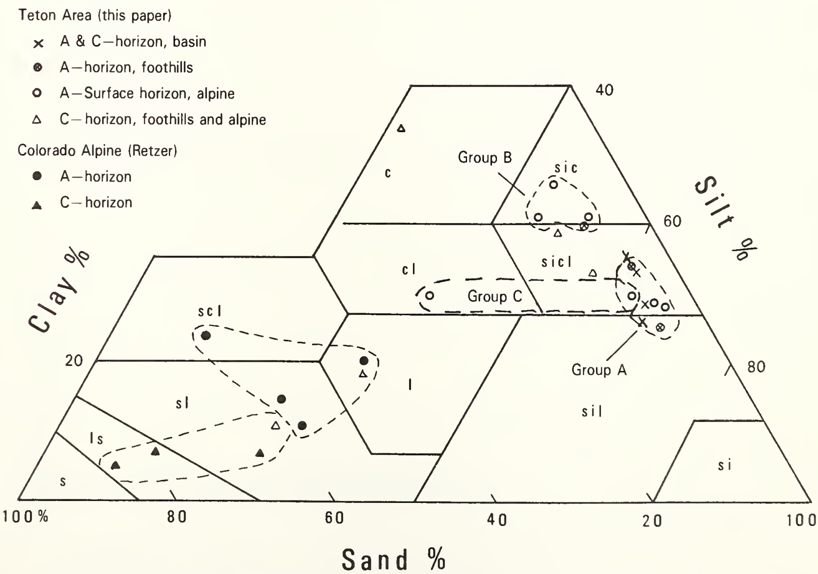

FIGURE 2. *USDA texture classes of surface and C horizons of soils in the Teton area, Wyoming and Idaho, and the Colorado alpine. Colorado data from Retzer (9).*

TABLE 2. *Heavy mineral percentages of surface horizons (silty cover) and C horizons (residuum) of very fine sand fraction (100-250 microns) from soils in the Teton area, Wyoming and Idaho.*

Site number		Mineral percentage (100-250 micron)														clay minerals (x = present)			
Silty material	Province	Muscovite	Zircon	Tourmaline	Garnet	Biotite	Apatite	Epidote	Hornblende	Augite	Hypersthene	Enstatite	Tremolite	Unidentified weathering products and colloidal material	Others	No. grains counted	Smectite	Vermiculite	Illite	Kaolinite
1	basin	5	3	8	—	6	—	—	21	22	10	3	—	11	11	316	—	—	—	—
2	basin	4	6	12	4	2	—	—	10	12	25	—	3	10	12	358	x	x	x	x
3	foothills	2	6	12	4	1	—	2	13	18	21	—	6	5	10	468	—	—	—	—
4	foothills	5	4	6	—	2	—	—	16	29	14	—	—	13	11	300	x	x	x	x
5	foothills	5	5	7	3	2	—	3	21	18	11	5	—	10	9	301	x	x	x	x
6	subalpine	6	5	9	—	2	—	—	19	21	13	—	2	18	5	187	x		x	x
10	alpine	2	5	12	1	—	4	—	17	22	18	—	3	5	11	317	x		x	x
12	alpine	12	2	5	3	16	—	—	13	6	21	1	—	14	7	240	x	x	x	x
13	alpine	12	7	7	—	9	—	—	19	9	11	1	—	19	6	299	x	x	x	x
Residuum	Lithology																			
3	tuff	26	2	7	—	33	—	—	15	—	—	—	8	—	9	749	x		x	
5	limestone	4	14	5	2	—	—	—	—	—	—	7	—	61	7	257	x		x	x
8	shale	43	—	3	—	23	—	—	2	2	4	—	—	22	1	205			x	
14	quartzite	29	4	4	—	21	—	—	7	4	3	—	—	27	1	345	x		x	x

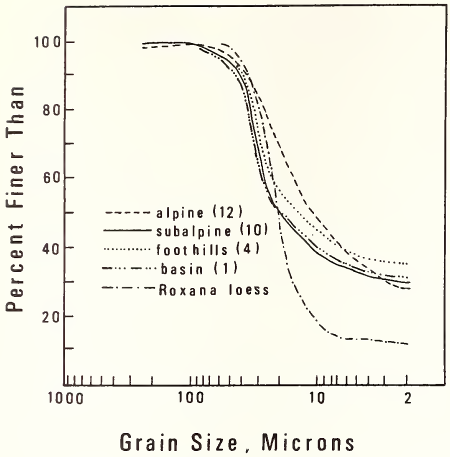

FIGURE 3a. *Cumulative particle size distribution curves of some surface horizons in the Teton area, Wyoming and Idaho compared with the Roxana loess in the Midwest.*

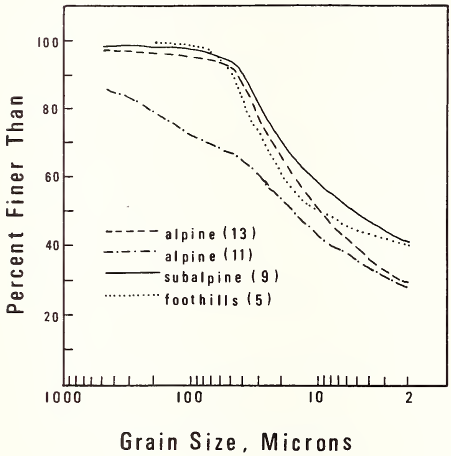

FIGURE 3b. *Cumulative particle size distribution curves of surface horizons of soils in the Teton area, Wyoming and Idaho, where loess has been mixed with residual material.*

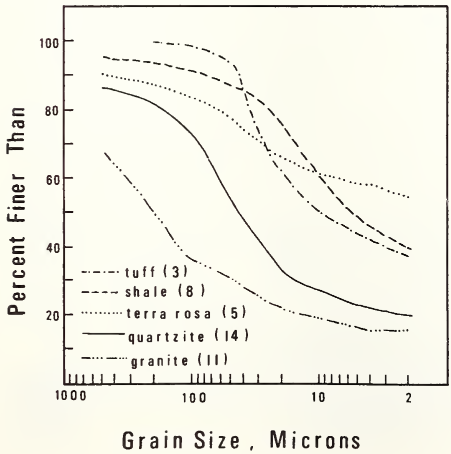

FIGURE 3c. *Cumulative particle size distribution curves of C horizons formed in residual material in the Teton area, Wyoming and Idaho.*

Samples of surface horizons from all physiographic units fall in this group and this material is presumed to be eolian in origin (Group A samples in Table 1). The second group (Figure 3b) generally has a higher percentage of clay than the first group, but also has more coarse material (Group B and C samples in Table 1). Samples from this group do not show pure loess curves and are presumed to be the result of mixing of loess with local bedrock material. This is particularly evident in sample 11, which was collected near the base of a rock glacier of granitic lithology. Only the basin soils do not have samples with this characteristic probably because loess is too thick for any mixing with bedrock material to occur. The C horizons formed in the bedrock residuum have a wide range of particle size distributions (Figure 3c), reflecting the diversity of lithology sampled.

Composition of the heavy mineral separates is quite uniform between surface horizons in the four physiographic units and contrasts strongly with the mineralogy of the residuum samples (Figure 4 and Table 2). Combined percentages of the less resistant minerals hornblende, augite and hypersthene in the loess range from 39 to 59 percent, a strong contrast to the 0 to 15 percent found in the residuum samples. In the residuum there is no significant enrichment of the resistant minerals zircon and tourmaline (3-19% in residuum vs. 7-18% in the loess), but percentages of muscovite and weathering products are much higher. Slightly lower percentages of less resistant and slightly higher percentages of more resistant minerals in the alpine samples (12 and 13) indicate that some mixing with local material has occurred.

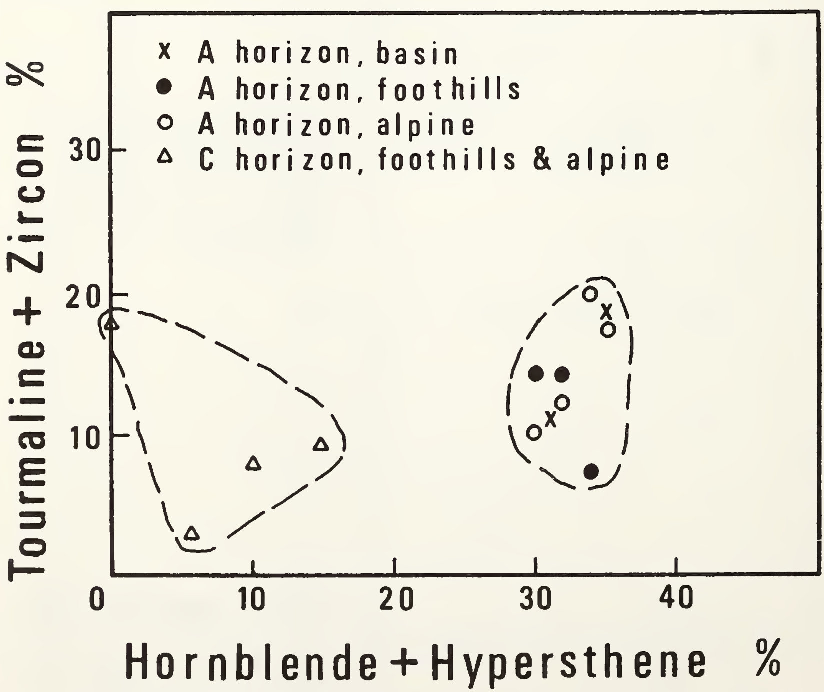

FIGURE 4. *Relationship of selected resistant to nonresistant minerals in the very fine sand fraction in surface and C horizons of soils in the Teton area, Wyoming and Idaho.*

The surface horizons of soils in all four physiographic units contain the clay minerals smectite, vermiculite, illite and kaolinite, except sites 6 and 10, which lack vermiculite (Table 2). In contrast, none of the four residual samples contained vermiculite. The sample of tuff also contained no kaolinite and the shale contained only illite. The clay mineral assemblage of the A horizons of sites 6 and 10 are similar to the terra rossa. The sample of material weathered from the Flathead Quartzite contains a much larger percentage of clay (19.7%) than would be expected. The clay minerals in this fraction are also similar to the terra rossa, suggesting that locally derived residual clays from the Madison Limestone are an important source of clays in soils over other geologic units. The large amount of fine material in the sample that was considered to be primarily weathered quartzite indicates that eolian material may be present even when it is not obvious in the field.

It is interesting to note that there is a rapid decrease in the maximum thickness of silty material between the basin and foothills units, but no significant decrease between the foothills and subalpine units (Table 1). Large rock fragments are common at the surface in the alpine tundra and on the subalpine benches underlain by limestone, but when excavated, they are found to lie in a matrix of silty material that may extend deeper than 63 cm (maximum reach of the longest armed investigator). These thicknesses of silty material do not support the conclusion of Lewis et al. (6) that the maximum elevation reached by loesses from the Snake River Plan is 2135 m. Site 12, which is silty material collected from a snowbank is at an elevation of 2945 m. Maximum thicknesses of silty material (in excess of 135 cm) on the subalpine shelves are in sinkholes which have apparently been filled by alluvial action. Soils formed in material that has been transported by water (sites 7 and 9) should be considered alluvial rather than loessial, but most of the alluvial material was probably blown to the area from a more distant source, and then moved short distances by alluvial processes.

The silty soils cluster into two groups based on clay content and a third group based on sand content and particle size distribution (Figure 2). In group A, clay content ranges from 26 to 34 percent. This groups includes samples from all physiographic units. In group B clay content ranges from 40 to 46 percent clay, and all lie on limestone. In this group, material from sites 7 and 9, located in sinkholes had been transported by water. It is likely that residual clays from local shales and terra rosa had been washed in as well. The sample from site 5 was on the east-facing backslope of the Madison Limestone and residual clay could have been worked into the profile by overwash and/or creep. Site 6 is in a summit position on the shelf formed by the Bighorn Dolomite and the mix of clay minerals is simlar to the terra rosa (Table 2) suggesting that it may have received clay washed or blown down from the Madison Limestone which rises above the Bighorn shelf in a sheer cliff several hundred feet high. The group C soils (sites 11 and 13) show mixing of eolian material with coarse-grained bedrock lithologies (granite and quartzite). Site 13 has by far the lowest silt content of the surface soil samples (38.5 percent) but this is still much higher than the residual material in the same profile (13.8 percent). Site 12 is interesting in that it appears to fall within group A based on relative percentages of silt, sand and clay (see Figure 2), but the particle size distribution curve (Figure 3b) shows that the silts tend to be coarser and do not form a loess-shaped curve.

The percentage of fine-grained material derived from eolian sources at the soil surface may exceed 80 percent, if we assume that the clay enrichment in group B (Figure 1) is solely from residual clays. Generally the fine-grained material that is eolian in origin decreases with depth. However, in some areas underlain by limestone, mixing is so thorough that there is no discernable change throughout the profile.

If a significant portion of the silts and clays in the soils on the western slope of the Teton Mountains are eolian in origin, as the preceding data suggests, the question remains, what is the source and age of this material? The clay content of the fine-grained material in the study area ranges from 25 to 46 percent, which is much higher than is typical for midwestern loesses. Deep less soils in Indiana, for example, commonly have from 12 to 16 percent clay in the A2 horizon (12). Glenn et al. (Agronomy Abstracts, 1976, p.160), in a study of loess deposits and soils of the Snake and tributary river valleys in western Wyoming and eastern Idaho concluded that local floodplains are the major sources of loess in the area. The high clay content of the local alluvium probably comes from two sources: 1) terra rossa eroded from the dip-slope of the Madison Limestone and 2) ash from volcanic activity, pervalant in the area throughout the Pleistocene.

Studies in other areas have found that the very fine sand and fine silt fraction of loess decreases with distance from source (see, for example, Souster et al. (13) for the Swift Current area of Saskatchewan). Lund et al. (7) point to the lack of consistent changes in texture with distance from the Snake River as evidence that fine-grained material in the Kimberly-Stricker Butte area, Twin Falls County, Idaho, is primarily lacustrine rather than loessial in origin. This area is 200 miles southwest of the Teton study area and a review of the basin soil profile descriptions and the Teton area soil survey (3) indicates that a reinterpretation of these soils in light of the findings of Lund et al. (7) is probably unnecessary. The faint bedding on soil profiles observed in Twin Falls County was not evident in the nine meter road cut at site number 1 in the Teton study area, and the generally steeper slopes of soils described in the soil survey as loessial (mostly greater than 4 percent) compared to the Twin Falls County study area (mostly less than 4 percent) also indicates that a lacustrine origin is unlikely. Unfortunately, not enough samples of the basin loesses were taken to determine whether there were trends in particle size distribution with distance from the Teton River.

Eolian silts and clays seem to be the major source of fine-grained material found in soils above timberline, but the only "pure" loess (site 12) found in the alpine was deposited on a snowbank near the Continental Divide. This is because frost churning, rock creep and transport by water are important proccesses affecting soil development in the subalpine and alpine areas of the Teton Mountains. After loess is deposited on the surface it is continually reworked and mixed with bedrock material. Loess deposited on limestone is often tranported to sinkholes with some mixing with residual clays. Frost churning also moves loess vertically in the soil profile filling in spaces between limestone fragments to depths deeper than 63 cm. In alpine areas on the Flathead Quartzite the loess is segregated into bands between stone stripes and mixed with coarser material formed by periglacial processes. Loess deposited on rock glacier debris becomes mixed with coarser material by frost churning and rock creep.

The genesis of the alpine soils is further complicated by evidence that fine-grained material is transported by wind within the study area. As discussed earlier, the clays in the sample of quartzite residuum are similar mineralogically to the terra rossa and wind is the only likely method of transport, since the site is located on the summit of a ridge. As a consequence, the term "loessial" should probably be used with caution when describing alpine soils in the Tetons, even though most of the silty material has probably been transported by wind from outside the area.

The presence of dirty snow banks in the alpine areas provides clear evidence that deposition of loess is an active process today. The thickness of silty material over glacial tills of Pinedale age (late Wisconsin) was observed to range between 16 and 22 cm in the study area. At a site that lies above the maximum height of Pinedale glaciation (not listed in Table 1) 62 cm of silty material with an argillic horizon covers a well developed paleosol formed in an additional 122 cm of silty material over Bull Lake Glacial till (early Wisconsin). At site 4 silty material 76 cm thick with an argillic horizon lies on top of a paleosol formed in additional silty material and Quaternary tuff. The paleosol was interpreted to be the same age as found in the Bull Lake till.

The Bull Lake till and Quaternary tuff are above the part of the valley affected by Pinedale glaciation, and it appears the accumulation of eolian material in the foothills since the soil-forming period of the Bull Lake-Pinedale interglacial has been on the order of 60 to 80 cm. These thicknesses are similar to the maximum thickness of loess observed in the alpine area. These observations are consistent with the conclusions of Fryxell (4) that deposition of loess around Jackson Hole Wyoming, east of the study area, probably began immediately following the withdrawal of Bull Lake glaciers.

Summary

Fine-grained materials in the surface horizons of soils in the alpine areas of the Teton Mountains are predominantly eolian in origin. Two lines of evidence support this conclusion: 1) particle size distribution and mineral composition of the very fine sand and clay fractions of the alpine soils sampled are very similar to those of soils sampled upwind in areas which are known to be loess, and 2) particle size and mineralogy of the surface material is distinctly different from that of the material weathered from the bedrock over which the soils lie. There has been some mixing of loess with residual materials by water and frost action, but the contribution of residuum to the overlying fine-grained material of the soil profile is relatively minor, except in the case of residual clays from limestone. No definite conclusion can be made from the available data regarding the source of the loess in the alpine areas, but most of the material has been deposited since the end of the Bull Lake (late Wisconsin) glaciation.

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