

**PEDOGENESIS, WEATHERING PROCESSES, AND ELEMENTAL
DISTRIBUTION ALONG A SOIL CLIMOSEQUENCE IN THE
SOUTHERN PIEDMONT**

by

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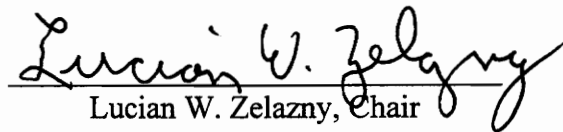
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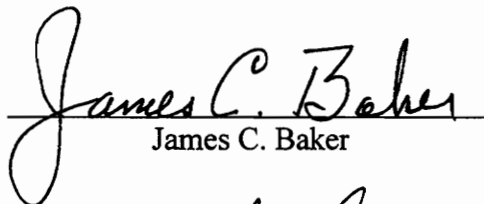
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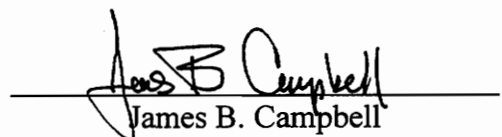
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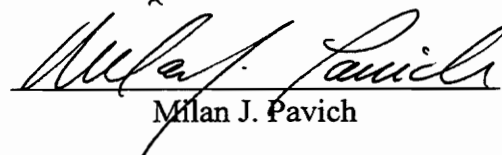
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(ABSTRACT)

Lack of age control for deposits >40,000 yr makes it impossible to differentiate between the complex effects of climate and geomorphic age on weathering intensity, soil profile development, mineral stability, and elemental flux within the pedoenvironment.

Upland soils formed in similar granitic parent materials along a climatic gradient in the nonglaciaded Southern Piedmont Province were analyzed in order to investigate whether soil response to variations in the weathering environment follows predictable, systematic patterns related to climate, and to quantify the nature and magnitude of these relationships, particularly with regard to processes of secondary mineral neoformation and transformation.

With few exceptions, all measures of weathering intensity increase dramatically from north to south on the Piedmont. The effective depth of pedogenesis in the Georgia and Alabama soils studied is nearly twice that of soils in Virginia and North Carolina, closely corresponding to calculated effective leaching indices developed for each site.

Kaolinite and gibbsite are poor indicators of soil age, weathering intensity, or paleoenvironment because of complex mechanisms of formation. Kaolinite genesis is shown to result from i) desilication of 2:1 phyllosilicates in surficial horizons ii) hydrolysis of K-feldspars, iii) resilication of gibbsite deep in the saprolite, and iv) recrystallization of halloysite in the soil/saprolite transition zone. Gibbsite forms from both desilication of kaolinite and by reprecipitation after Na-feldspar dissolution at depth. Halloysite is common to all soils as the product of both feldspar and biotite weathering, thus illustrating the overriding influence of microenvironment in secondary minerals formation.

Clay content is linearly related to Fe_d in Virginia, North Carolina, and Georgia soils, with regression slopes decreasing from north to south. Clay content levels off at higher Fe_d values for Alabama soils, indicating that i) steady-state conditions are approached as pedogenic clay formation reaches some intrinsic, self-limiting threshold value, and ii) the limiting factor in clay production over time is leaching intensity, and not the concentration of Fe-bearing primary minerals, in contrast to the soils studied in Virginia, North Carolina, and Georgia, appear to have reached a steady-state with regard to clay production. These data suggest that soils farther south on the Piedmont have experienced conditions of greater geomorphic stability and longer weathering.

Until we can obtain numerical dates for Piedmont soils, however, one can only speculate as to whether the observed differences in pedogenesis and weathering intensity are due to age effects, climatic effects — or both.

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Additionally, thanks go to all those individuals who assisted with the field work, usually during hot summer days. They are: Louie Frost, Larry LaPlante, Tom Macfie and Jody Palmer, in Georgia; and Lawrence McGhee in Alabama.

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Most of all, I thank God that I'm done.

¹ My apologies for omitting the mice — we never got along anyway.

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Chapter I

Introduction

Strong relationships between climate, primarily temperature and precipitation, and soil properties such as solum thickness, depth to carbonates, N and clay contents, pH, and cation exchange capacity (CEC) have long been known to exist (Jenny, 1941; Arkley, 1963, Ruhe, 1984). In the southeastern U.S., soil properties are dramatically influenced by different temperature and moisture regimes, largely through effects on solubilities of solid and gaseous phases, chemical reaction rates, and speciation of ions in soil solution (Buol et al., 1990). These factors, and additionally the rate at which drainage water leaches through the soil, dictate to a large extent the nature and products of primary and secondary mineral weathering, the degree of soil profile development, the magnitude and depth of clay illuviation, the extent of soil rubification, and the surface charge properties of the soil exchange complex.

In the nonglaciaded Southern Piedmont Province (Fig. 1-1), soil rubification, weathering intensity, and degree of profile development on similar parent materials increase noticeably from north to south (Kaster, 1960; Porter et al., 1963; Frost, 1979) as do reported increases in both hematite and gibbsite, and decreases in goethite in this same direction (Southern Regional Project S-14, 1959). A similar change from soils having dominantly mixed mineralogy to those having kaolinitic mineralogy (Soil Survey Staff,

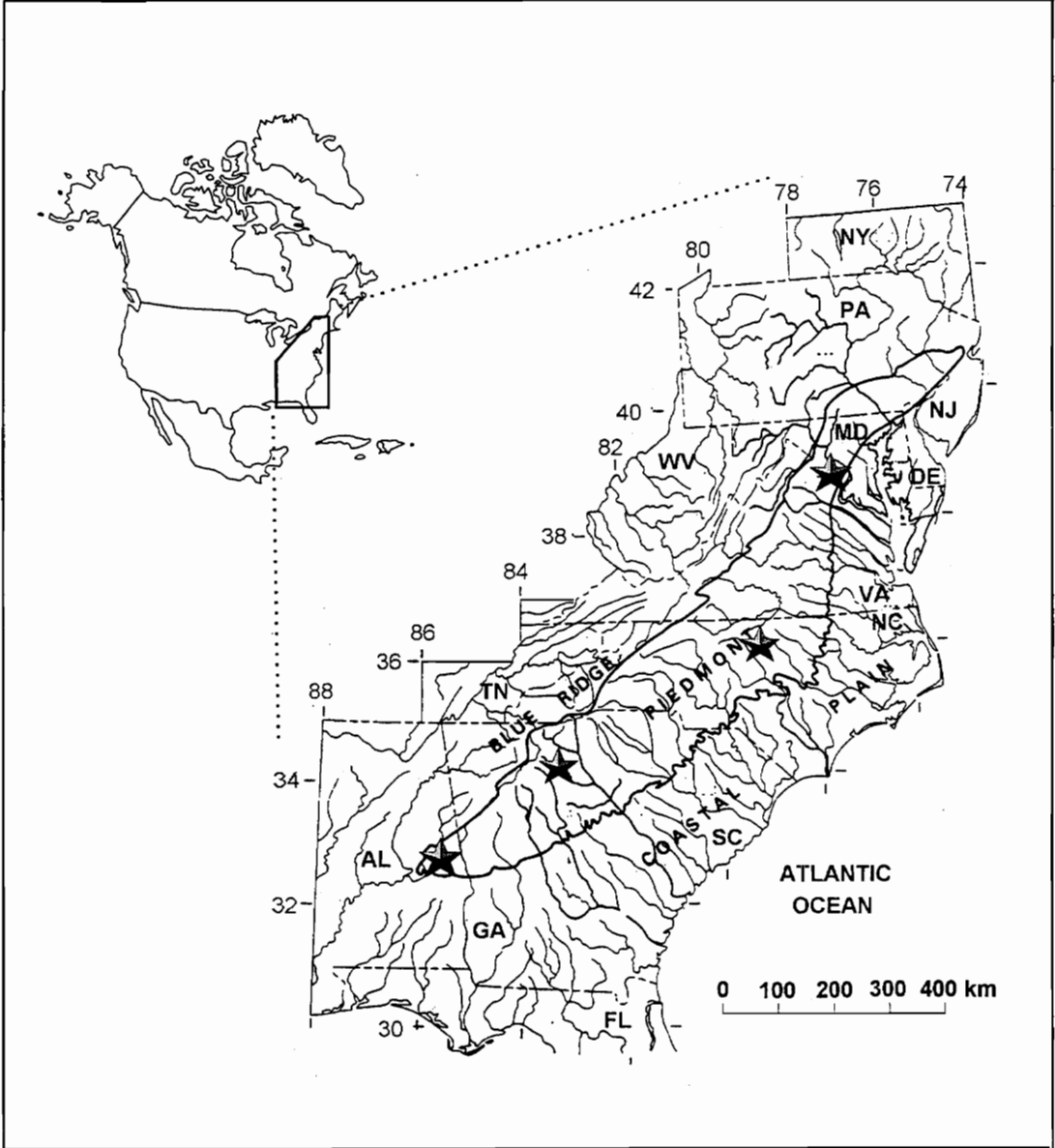


Fig. 1-1. Location of the study areas (★) showing the Piedmont, Blue Ridge, and Coastal Plain physiographic provinces with major drainages (modified after Hack, 1982).

1990) also closely coincides with a shift from mesic soils (MAST¹ 8° to 15°C) in the northern part of the Piedmont to thermic soils (MAST 15° to 22°C) in the southern part.

The Piedmont region may be particularly sensitive to potential changes in climate because of high population density and agricultural drought hazard. Properties of Piedmont soils that would be expected to change in response to changes in the weathering environment within 10 to 100 years are generally restricted to levels of soluble and exchangeable nutrients, amount and quality of organic C and N levels, water availability and soil hydrologic properties, and effects on microbial processes (Rounsevell and Loveland (1994). In the longer-term, soil properties that may be influenced by different temperature and moisture regimes include solubilities of solid and gaseous phases, chemical reaction rates, and speciation of ions in soil solution (Buol et al., 1990). Millennia-scale effects could include changes in Si solubility, alterations to both crystalline and noncrystalline Fe and Al oxides, and increases in the amount of hydroxy-Al or -Fe interlaying in vermiculite interlayers. These effects on Fe- and Al-oxides and vermiculite interlayering may be especially important because increases in these components would significantly modify soil physicochemical properties resulting, partly, in reduction of CEC, increases in anion retention characteristics, increases in K⁺ selectivity, and changes in flocculation/dispersion phenomena which influence soil structure, surface crusting, and rates of infiltration, runoff, and erosion. These factors may potentially cause a northward shift in the distribution and extent of kandic, oxidic,

¹ Mean annual soil temperature (at 50 cm depth)

and kaolinitic soils (Soil Survey Staff, 1990) in the southeastern US. Expanded areas of these low-fertility soils would result in greater potential for groundwater pollution through higher rates of fertilizers, herbicides, and insecticides required to sustain agricultural productivity.

Despite the acknowledged dominion of temperature, precipitation, and (or) potential evapotranspiration over many Piedmont soil properties, there is currently a scarcity of quantitative data regarding the specific effects of both climate and time on weathering processes, owing largely to a lack of age control for deposits >40,000 yr (Mills and Delcourt, 1991). We presently cannot differentiate between the combined effects of climate and geomorphic age on weathering intensity, soil profile development, mineral stability and transformation, and elemental flux within the pedoenvironment (Boardman, 1985; Feldman et al., 1995), nor do we understand how these parameters will be affected by future changes in the chemical and physical climate. Additionally, we know relatively little about rates of geochemical weathering and pedogenic processes under field conditions (Velbel, 1986), and even less about minimum climatic thresholds required to initiate changes in steady-state soil systems and processes. Without this information, we can neither identify the degree to which soil properties and secondary phases persist, alter, or become obliterated in the weathering environment as conditions change, nor can we predict with any certainty which soil features are reliable as indicators of retrospective or prospective environmental change.

In addition to acknowledged uncertainties regarding climatic effects on weathering, other aspects concerning the evolution and age of the Appalachian landscape have puzzled geomorphologists and pedologists for more than a century. Recent research on Piedmont morphogenesis has largely centered around documenting rates of geochemical denudation of crystalline bedrock in order to determine the nature and residence time of soil/saprolite and thus determine landscape age and geomorphic history of the region (Cleaves et al., 1970; Pavich, 1986, 1989a,b; Cleaves, 1989). These studies have commonly assumed that existing soils have formed in uniform parent material from the underlying saprolite. Little mineralogical evidence has been provided either to confirm or disprove this relationship, however, which leaves open the possibility that many Piedmont soils are not the same age as the geomorphic surfaces with which they coexist. The widespread relict colluvial deposits, stone lines, and other indicators of sediment transport that are commonly found on both steep slopes and nearly level, 'stable' upland interfluves in the Piedmont as far south as Georgia, for example (Parizak and Woodruff, 1957; Eargle, 1940; 1977; Overstreet et al., 1968; Whittecar, 1985; McCracken et al., 1989; Genthner, 1990; Whittecar and Ryter, 1992), strongly suggest that extensive areas of the nonglaciaded Piedmont may have been subject to mass-wasting by periglacial stripping. However, the climatic significance of these deposits south of the glacial border has only recently begun to be recognized. I would hope that this study would shed some additional light on what is already known about the Cenozoic weathering and erosional history of the Piedmont, and perhaps the broader Appalachian region.

Toward that end, I initiated a detailed study of weathering trends, soils, and surficial deposits along a carefully controlled, well-defined regional climatic gradient in the Southern Piedmont. The principal objectives of the study were to:

- investigate whether the response of soils to variations in the weathering environment follows predictable, systematic patterns related to climate on the Southern Piedmont,
- quantify, for a range of pedogenic processes, the magnitude of soil physical, chemical, and mineralogical changes induced by the natural range of existing climatic regimes across the region,
- describe and document weathering processes and products of secondary mineral neoformation and transformation from north to south on the Piedmont,
- investigate mechanisms of biotite kaolinitization on the Piedmont.

This chapter provides an overview of the methods used, description of the geologic setting of the study areas, and a summary of bedrock geology at each site. A discussion of modern climatic parameters on the Piedmont is also provided along with information regarding model estimates of predicted climate change through the next century. Chapter II includes a detailed review of the literature concerning the evolution of the Piedmont landscape and pertinent pedological studies. Chapters III and IV deal with the deposition of a previously unidentified loess cap on the northern Virginia Piedmont, and the subsequent modification and alteration in both the cap and an underlying paleosol. Chapter V reports on the observed trends in mineral weathering, physicochemical properties, and nature of elemental flux from north to south on the Piedmont. Finally,

Chapter VI includes a preliminary investigation of a previously unidentified weathering mechanism involving the *in situ* alteration of intact biotite grains to halloysite in deep saprolites throughout the Piedmont. Results are summarized in Chapter VII.

Methods

I used a climosequence approach to study soil development from north to south on the Piedmont. While it is virtually impossible to exert rigorous experimental controls on open, natural soil systems, the similarity of soil parent materials and landscape features selected for study in this project should permit accurate analysis of soil properties as a function of climate. Duplicate soil pedons were sampled at each of four sites located in northern Virginia, north-central North Carolina, northeastern Georgia, and east-central Alabama, respectively (Fig. 1-1). These sites span the modern climatic extremes on the Piedmont (Fig. 1-2, 1-3; Table 1-1) and encompass the range of temperature and precipitation predicted by climate models under several global warming scenarios through the next century (Houghton et al., 1990).

Topographic and parent material effects were minimized by sampling exclusively on stable upland interfluvies underlain by lithologically and structurally similar granitic plutons that have been well-characterized (Adams, 1933; Gault, 1945; Deininger, 1975; Neathery et al., 1975; Seiders et al., 1975; Briggs et al., 1978; Stormer and Whitney, 1980; Drake 1986; Drake and Froelich, 1968; Drummond and Green, 1987). These lithostratigraphic units exhibit a relatively high degree of homogeneity both within and between sites, particularly with regard to bulk chemical composition and mineralogy

Table 1-1. Selected climatic averages for the study areas and calculated leaching indices. †

Site	Location	Elevation m	Temperature °C	Precipitation		PET‡		Leaching Index§	Deep Percolation >100 cm¶
				Monthly	Total Annual	Monthly	Total Annual		
VA	38°57'N, 77°27'W	88	12.2	85	1024	99	1190	238	80
NC	36°05'N, 79°57'W	269	14.4	90	1082	109	1311	209	85
GA	33°57'N, 83°19'W	243	16.3	106	1275	119	1423	281	124
AL	32°36'N, 85°30'W	198	17.2	119	1422	121	1452	357	201
	(r² vs. latitude)		(-0.98)	(-0.89)	(-0.88)	(-0.98)	(-0.98)	(-0.60)	(-0.74)

† U.S. Dept. of Commerce. Climatological Data Summaries, 1987-1988.

‡ Potential evapotranspiration calculated by the Blaney and Criddle method (Doorenbos and Pruitt, 1975).

§ After Arkley (1963).

¶ Assumes soil AWHC = 15.7 cm H₂O / 100 cm soil.

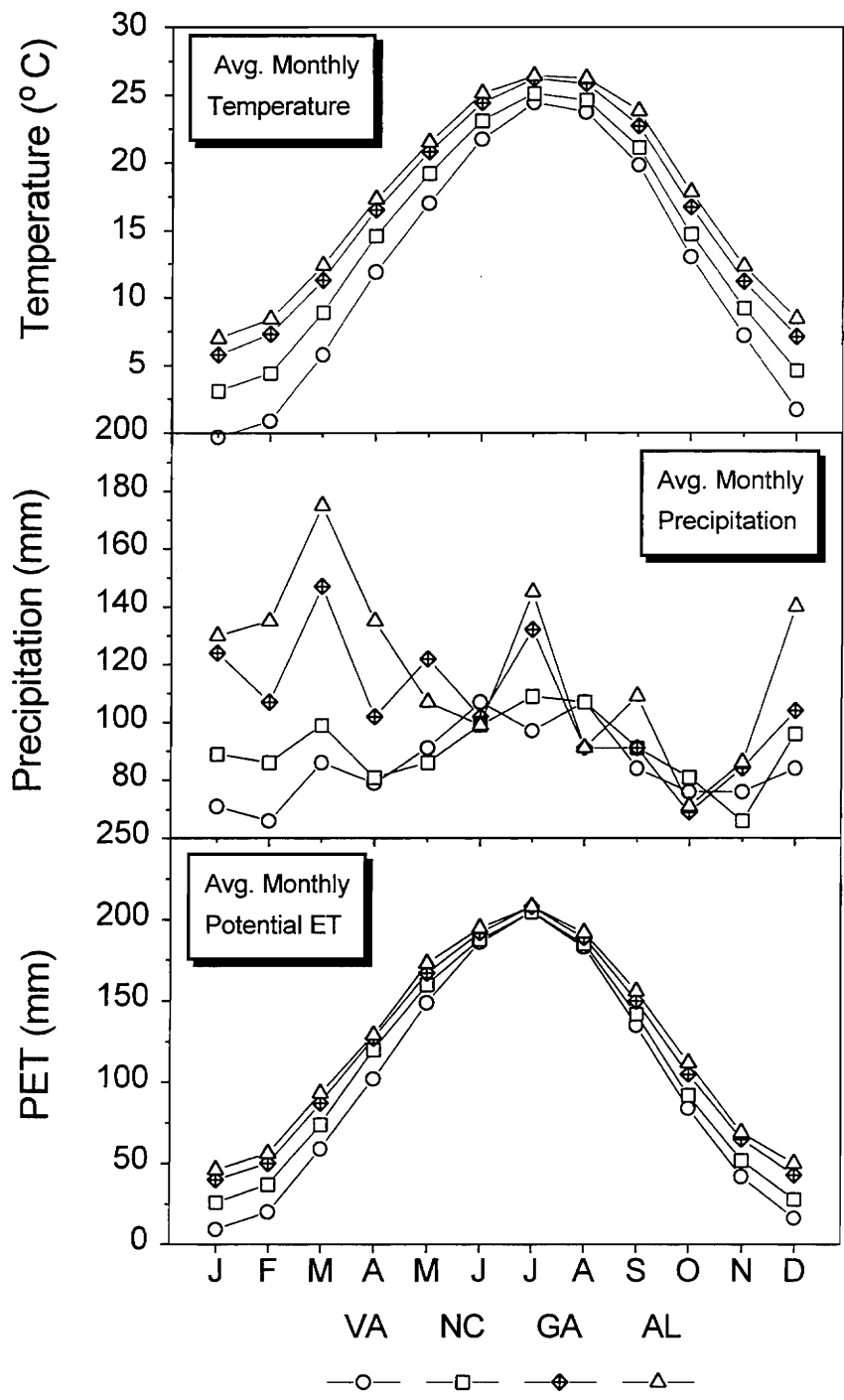


Fig. 1-2. Average monthly temperature, precipitation, and potential evapotranspiration (PET) at each study site.

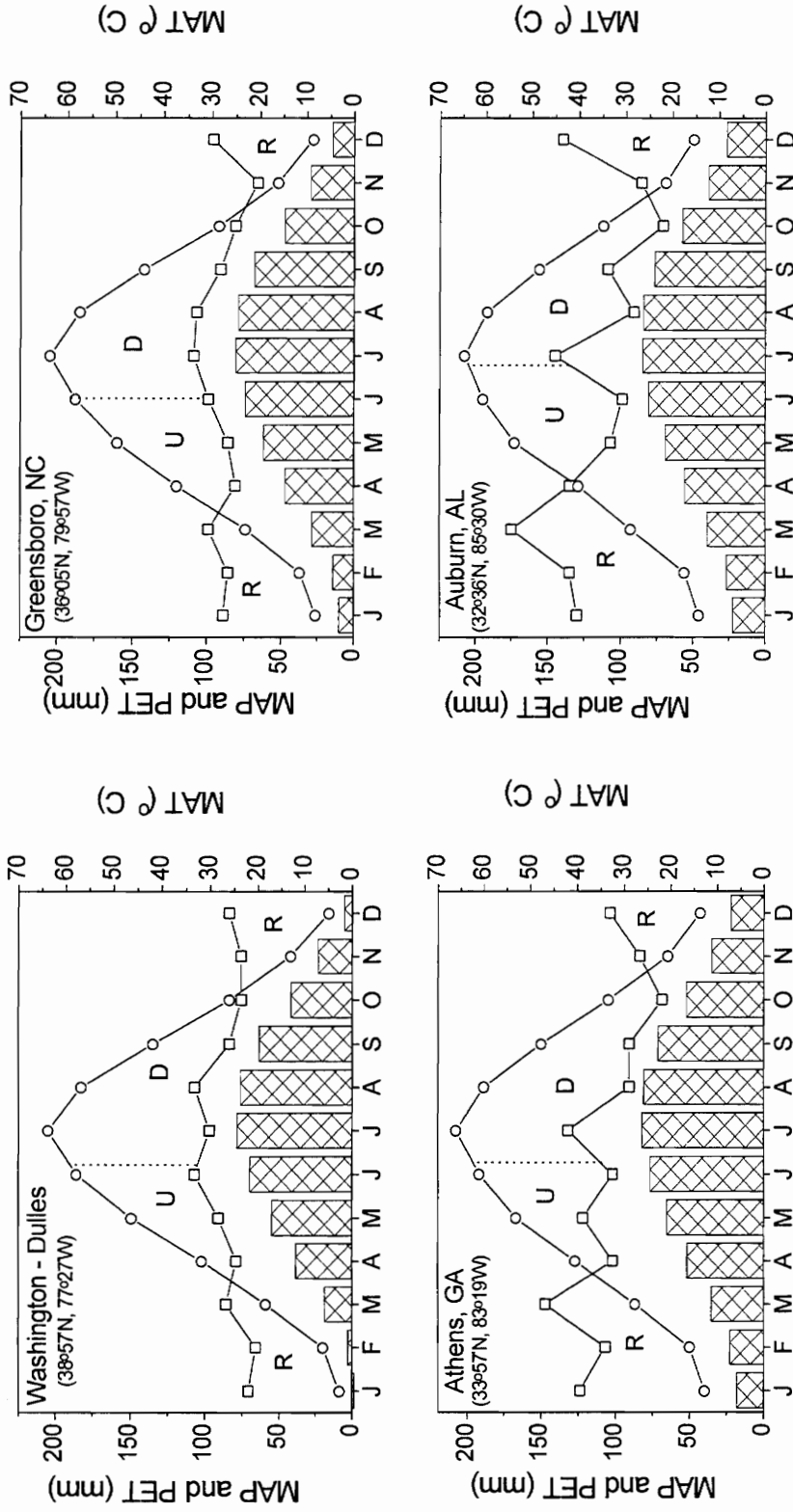


Fig. 1-3. Water balance data for each study site showing mean monthly temperature (MAT), mean annual precipitation (MAP), and potential evapotranspiration (PET). Periods of soil moisture recharge (R), utilization (U), and deficit (D) are shown for a soil with an assumed water holding capacity of $15.7 \text{ cm H}_2\text{O} \cdot 100 \text{ cm}^{-1}$ soil are shown.

(Tables 1-2 and 1-3). Total Fe content is very similar between all four granites, which permits independent determination of the effects of climate on the formation and transformation of secondary Fe-oxides.

Duplicate soil pedons separated by not less than 2 km were sampled at each of the four study areas (Fig. 1-1). All pedons were located on high, broad, nearly level or gently convex interfluvies to ensure that areas selected for detailed study have formed in residuum and are representative of soils exhibiting maximum pedogenic development in the landscape with minimal likelihood of lithologic discontinuity. These areas are neither highly erosive nor subject to overwash from other landscape positions, and thus the transport of recent sediments by slope processes, either to or from each site, is expected to have been minimal.

Geologic Setting of the Study Areas

The Piedmont physiographic province is an upland plateau of low to moderate relief, extending from southeastern New York State to central Alabama, with a regional dip to the east of about 3.8 m/km (Thornbury, 1965). It is bounded on the west by the Blue Ridge escarpment and on the east, at the Fall Zone, by Cretaceous and Tertiary rocks and sediments that comprise the Coastal Plain (Fig. 1-1). Large-scale features of the landscape are primarily controlled by lithology which consists of complexly-folded metamorphic, metaigneous, and metasedimentary rocks, and elongated strips of unmetamorphosed late

Table 1-2. Mean chemical composition of the Occoquan Granite (VA), Roxboro Formation (NC), Elberton Granite (GA), and Elkahatchee Quartz Diorite (AL). †

Site	n	FeO	Fe ₂ O ₃	Fe	Total		SiO ₂	Al ₂ O ₃	TiO ₂	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅
					SiO ₂	Al ₂ O ₃									
%															
VA	8	1.45a†	1.26a	2.71a	5026a	72.66a	13.84a	0.39a	0.6a	0.79a	2.44a	2.81a	2.84a	0.03a	
NC	7	1.11a	.94ab	2.04a	5.73b	74.76b	13.07b	0.20b	.010b	0.23b	1.02b	4.13b	3.77b	0.01a	
GA§	29	1.09a	0.79b	1.88a	4.75c	71.50a	15.06c	0.38a	0.03c	0.58ab	1.65b	3.98b	4.82c	0.10b	
AL	7	3.06b	2.05c	5.12b	3.67d	60.66c	16.55d	0.79c	0.07a	3.05c	4.17c	4.31b	2.58a	0.33c	

† After Seiders et al. (1975), Briggs et al. (1978), Storer and Whitney (1980), and Drummond and Green (1987).

‡ Means followed by different letters within a column are significantly different (P=0.05).

§ Numbers represent average of 29 observations.

Table 1-3. Mean ranges of mineralogical composition of the Occoquan Granite (VA), Roxboro Formation (NC), Elberton Granite (GA), and Elkahatchee Quartz Diorite (AL).†

Mineral	Occoquan	Roxboro	Elberton	Elkahatchee
			%	
Quartz	29-42	26-34	~30	25-33
K-Feldspar	24-30	21-32	30-35	a ‡
Plagioclase	17-22	28-41	30-35	38-49
Biotite	1-4	0-4	4-7	14-18
Muscovite	8-15	0-2	0-5	0-6
Chlorite	a	a	--	a
Epidote	2-5	2-9	--	1-5
Sphene	a	a	a	a
Apatite	a	--	--	a
Garnet	a	--	--	a §
Opagues ¶	a	a	--	--
Allanite	--	a	a	a
Pyrite	--	--	--	a
Carbonates	a	--	--	a
Zircon	a	a	a	a

† After Briggs et al. (1978), Deininger (1975), Drummond and Green (1987), Seiders et al. (1975), and Stormer et al. (1980).

‡ Common accessory mineral.

§ Local areas high in spessartine and almadine garnets have been observed.

¶ Ilmenite-leucoxene, hematite, magnetite, etc.

Triassic age sandstones, shales, and conglomerates. Triassic dikes are also common in the metamorphic rocks of the Piedmont.

Local relief is moderate throughout much of the Piedmont, and even low-order stream channels are deeply incised into saprolite or bedrock. Relief in the Triassic Lowlands is considerably less than in most other areas, with less stream incision and broad floodplains. Drainage patterns throughout the Piedmont are closely controlled by rock joints and foliation (Hack, 1982). Inselbergs/monadnocks, or locally high areas capped with resistant rocks, are also commonly observed (Kesel, 1974).

Summary of Bedrock Geology

The granitic plutons selected for this study were initially recommended to me because of their high degree of homogeneity, both within and across sites (A.K. Sinha, Dept. of Geological Sciences, Virginia Polytech. Inst. and State Univ., Blacksburg, VA 24061, personal communication). Similarity in bulk chemical composition is of considerable interest in this study, particularly with regard to total Fe content which is needed to quantitatively assess the formation and transformation of secondary Fe oxides in relation to climate. Mean chemical composition of representative rock samples from each study site shows a high degree of similarity with respect to both total Fe and FeO contents for all but the Alabama rocks, which are significantly higher in these parameters (Table 1-2). Silica and Al levels are also nearly equal for rocks of each of the three northernmost sites,

but the Alabama rocks are slightly more mafic, with less Si and more alkaline earth metals.

Occoquan Granite, Virginia

The Occoquan Batholith in Fairfax and Prince William Counties, Virginia, consists of medium-grained, light gray monzogranite with moderate to strong metamorphic foliation and a well-developed mineral lineation. Seiders et al. (1975) date the crystallization of magma at 560 Ma B.P. based on U-Pb dating of zircons. Independent Rb-Sr whole-rock analysis by Mose and Nagel (1982) sets the emplacement of the Occoquan at 494 Ma B.P., but Drake (1986) cautions that this method tends to underestimate age. The granite underwent a mild metamorphism near the end of the Paleozoic (Seiders et al., 1975).

Major minerals include quartz, plagioclase (commonly with inclusions of sericite), and microcline with subordinate muscovite, biotite, and epidote (Seiders et al., 1975) (Table 1-3). Plagioclase varies between albite (Ab₉₄₋₉₅) and a mixture of albite and oligoclase (Pavich, 1986). Microcline is generally free of inclusions and is commonly untwinned. Accessory minerals include apatite, sphene, garnet, chlorite, allanite, calcite, and zircon.

Roxboro Formation, North Carolina

The Roxboro Metagranite is located in an area centered around Person and Caswell Counties, North Carolina. The rock intruded into thick sequences of Precambrian and

Early Cambrian volcanic and epiclastic rocks about 575 Ma B.P. and later underwent regional metamorphism about 200 Ma after its emplacement (Briggs et al., 1978). A foliation and preferred orientation of biotite, muscovite, chlorite, and epidote was apparently developed in this pluton during this deformational event, which also resulted in a significant modification of the original igneous mineralogy. Quartz and nearly equal amounts of both alkali and plagioclase feldspars comprise 84 to 95% of the total mineral suite with biotite, muscovite, chlorite, and epidote making up the bulk of the remainder (Briggs et al., 1978) (Table 1-2). All K-feldspars are nearly pure microcline (Or_{97}), and all plagioclase is nearly pure low albite (Ab_{97-99}) (Briggs et al., 1978).

Elberton Granite, Georgia

The Elberton Batholith intrudes into the high-grade metamorphic terrain of the Inner Piedmont region of eastern Georgia in an area approximately 60 km long and 10 km wide, extending from south of Lexington, Georgia to north of Elberton, Georgia. The area of exposure is generally parallel to northeast-striking Inner Piedmont rock units (Ross and Bickford, 1980) and is placed to the southeast side of the belt of maximum regional metamorphic intensity (Ramspott, 1964). The Elberton Granite is a fine- to medium-grained light-gray granite that shows no evidence of metamorphic deformation or metamorphic fabric (Stormer et al., 1980). There is very poorly defined foliation associated with biotite alignment, usually variable in direction. The entire pluton is reported to be the product of a single intrusive event that occurred ~330 Ma B.P., during the peak of tectonic and metamorphic events in the area (Stormer et al., 1980).

The Elberton Granite exhibits extreme mineralogical and chemical homogeneity despite minor variations in color and grain-size. Darker colors of the rock are attributed to impurities in the feldspars (Ramspott, 1964). Variations in grain-size across the entire 500 sq. km exposure are slight and gradual. Compositionally, the rock contains ~30% quartz, 30-35% plagioclase, and 30-35% potassium feldspar (Ramspott, 1964; Stormer et al., 1980) (Table 1-3). Biotite content ranges from about 4-7%. Common accessory minerals include sphene, zircon, allanite, magnetite, apatite, and ilmenite-hematite. Muscovite, epidote, and chlorite may be locally important, but these minerals are reported to be the secondary products of post-emplacement alteration (Whitney et al., 1980). Plagioclase is predominantly unzoned oligoclase (Ab₇₄₋₈₆) and K-feldspar is mainly microcline with a composition of ~Or₉₅.

Elkahatchee Quartz Diorite, Alabama

Igneous and metamorphic rocks that dip under the Alabama Coastal Plain sediments to the south are at the southern limit of the crystalline Appalachian system. The Elkahatchee Quartz Diorite comprises the largest part of the Pickneyville Quartz Diorite Complex, occupying an area about 65 km long and up to 20 km wide that strikes northeast in eastern Coosa and northwestern Tallapoosa Counties, Alabama (Gault, 1945; Deninger, 1975). The Elkahatchee is a medium- to dark-gray, coarse-grained mesocratic rock with well-defined foliation, lineation, and local textural or light-dark banding ranging from millimeters to meters. Essential minerals, comprising >90% of the mineral suite, include plagioclase, quartz, and biotite that are recognizable in hand specimen.

Smaller amounts of muscovite and epidote make up the remainder (Gault, 1945) (Table 1-3). Accessory minerals include apatite, sphene, zircon, pyrite, allanite, garnet, and K-feldspar. Plagioclase ranges from labradorite (Ab_{45}) to albite (Ab_{95}), but sodic andesine and oligoclase are most abundant. Light-colored quartzitic, granitic dikes up to about 30 cm thick are widespread in the Elkahatchee. Local igneous intrusions of the Hissop Granite that are mineralogically similar to these granitic dikes also occur within the Elkahatchee, occurring in elongate bands, parallel to the regional strike. These bodies range from 2 to 10 km in length and from a few hundred to several thousand meters in width (Deninger, 1975).

Emplacement of the Elkahatchee is reported to have occurred during the Late Cambrian period (~520 Ma B.P.), and a regional metamorphic event is thought to have occurred at ~300 Ma B.P. (Neathery et al., 1975).

Modern Climate and Predicted Change

Modern climate of the Southern Piedmont is humid, temperate to subtropical, spanning a considerable temperature, precipitation, and evapotranspiration gradient from about 32° to 41°N latitude (Thornbury, 1965). Mean annual temperature across the region ranges from about 12° to 17°C and is highly correlated with latitude (Table 1-1). Southern sites experience temperatures about 7°C higher than northern areas during winter months but only about 3°C higher during the summer (Fig. 1-2). Mean annual precipitation ranges from about 1000 mm in the north to about 1500 mm in the south,

showing considerably more variability than temperature trends. Precipitation is relatively uniformly distributed throughout the year at each site, with maximum values occurring during the late Fall and relative increases occurring in early Spring and mid-Summer. Calculated potential evapotranspiration trends closely parallel those of temperature, but these values essentially converge for all sites during periods of maximum summer temperature. Water balance data for the study areas indicates that soils in lower latitudes experience less summer moisture deficit and have higher capacity for deep percolation below the solum than soils farther north (Fig. 1-3, Table 1-1). Comparison of these values with global changes predicted by climatic models for the end of the next century suggests that the proposed study area covers the range of temperature increase expected for the Piedmont through the year 2100.

U.S. Department of Energy models project the equilibrium climate response to a doubling of atmospheric CO₂ (or its radiative equivalent from all greenhouse gases) to range from 1.9° to 5.2°C (Schlesinger and Mitchell, 1985; Mitchell et al., 1990). Scharpenseel et al. (1990) reports that average temperatures are expected to increase 2° to 3°C in the tropics and 5°C in sub-polar regions within 50 years. The World Meteorological Organization, through the Intergovernmental Panel on Climate Change (IPCC), estimates that under the scenerio of current emissions of greenhouse gases, global mean temperatures will increase at a rate of about 0.3° to 0.4°C per decade (Houghton et al., 1990). This will result in an increase of about 1°C above the present value by the year 2025 and a projected temperature rise of about 4.5°C by the end of the next century. If

emissions are subject to controls, the most optimistic IPCC estimates indicate that average rates of temperature increase can be limited to about 0.1°C per decade, or an increase of about 2.0°C by the year 2100. The average surface temperature of the Earth during the last ice age, to put these values into perspective, was only about 5°C cooler than it is today (Ausubel, 1991).

Among the many possible ramifications of global warming, climatological models predict a northward shift in the sea-ice boundary, eustatic sea level rise of 0.7 to 3 m, up to 10% increase in average annual precipitation and river runoff in Northern Hemisphere high latitudes, and increasing dryness in northern mid-latitude interior continental regions during summer (Mahlman, 1989; Scharpenseel et al., 1990). Drier conditions are expected to be caused primarily by an earlier termination of snowmelt and rainy periods in these areas and thus an earlier onset of the normal spring-to-summer reduction of soil moisture through evapotranspiration. Any minor increases in rainfall that some continental interior regions may experience are expected to be taken up by increased evapotranspiration of vegetation or crops at the expected higher temperatures (Brinkman, 1990).

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Chapter II

Piedmont Morphogenesis and Pedologic Studies

The study of soil genesis and the study of landscape evolution is inextricably connected. Soils form an integral part of all landscape elements and geomorphic surfaces and thus the history of the Piedmont landscape is intimately tied to the history of soil development across the region. There is a strong relation between degree of soil development and time, particularly on stable surfaces, but critical questions involving kinetics of mineral dissolution, erosional history, and response to climatic change must be resolved before a complete understanding of the relationships between pedogenesis and the evolution of the Piedmont regolith can be obtained. In order to correctly interpret these interactions, it is necessary to review the theoretical framework from which our current understanding of humid-temperate-region geomorphic processes arises.

Theoretical Models

Evolution of the Appalachian landscape has puzzled geomorphologists for more than a century. That many unresolved problems remain is evidenced by the considerable polarization among geomorphologists recorded in the literature during the past two

decades. Aspects of Appalachian landscapes that have remained somewhat obscure include: (1) the noticeable accordance of summits, (2) the reversal of the original northwestward-flowing drainage, (3) alignment of wind and water gaps, and (4) the transverse course of major streams with respect to structural-lithologic trends where other courses that circumvented resistant rock ridges were assumed to have been available (Clark, 1989).

The first general model of landscape development was proposed by W. M. Davis (1889) whose concepts went virtually unchallenged for more than fifty years. Following increasing criticism of the classical Davisian system of landform evolution (Davis, 1889, 1899), the “Appalachian problem” became codified into several subsets, many of which are embellishments of statements in the 1889 publication (King, 1953; Penck, 1953).

In sympathy with the popular Darwinian evolution theory of the time, Davis (1889) envisioned a “Geographic Cycle” or a “Cycle of Erosion” in anthropomorphic terms where, after an initial period of uplift, landscapes proceeded in time through an orderly, sequential evolution, beginning with an initial stage of active downcutting by streams (youth), a mature stage, and terminating with old age in a penultimate form termed a ‘peneplain’. Over time, the entire landscape was thought to be reduced to a base level toward which the region erodes and to which streams become ‘graded’, a condition wherein stream gradients are so reduced that they can no longer cut vertically downward, but are said to be in equilibrium, depositing on the floodplain an amount of material equal to what they erode by lateral cutting. In other words, a graded stream is able to transport

its load of sediment eroded from upstream and no more. Once streams attain grade, the rate of river incision is thought to be reduced to negligible proportions thus initiating a stage of maturity where the streams, unable to erode their beds, shift laterally forming floodplains and meanders. Hillslopes are subsequently reduced by weathering and surface creep to flatter and flatter angles until they meet upon the interfluves, destroying the initial surface. This process of 'lateral planation' resulted in the lowering of drainage divides and lateral expansion of debris-covered areas of the valley floor produced by stream meandering and deposition. As slopes continue to flatten through time, interfluves are lowered more rapidly than the river beds until only a convex-upward surface of transported materials resting unconformably on residual material would remain. Old age was identified by landscapes covered with broad sheets of alluvium/colluvium and floodplains several times wider than stream meander belts. This surface of regional extent produced by long-continued fluvial erosion, having low local relief and little structural control of topographic elements characterized Davis' peneplain.

Renewed periods of uplift were thought to initiate the cycle once again with streams now downcutting into relict planation surfaces. Davis assumed that these periods of rapid uplift were spasmodic in nature and separated by much longer periods of stability during which the cycle ran its course. To allow for interruptions which were viewed as temporary disturbances, Davis developed the concept of rejuvenation which occurred when a landscape was caught in mid-cycle and forced to revert to downcutting conditions because of renewed uplift and adjustment of baselevel.

Multiple residual peneplains formed by successive incomplete erosion cycles have been invoked to explain much of the present Appalachian landscape. The accordant sandstone ridgetops in the Valley and Ridge Province of Pennsylvania have been described as remnants of the Schooley peneplain, and the uplands of the Valley floor as fragments of the Harrisburg peneplain (Sevon, 1983). Similarly, the Piedmont was described by Johnson (1931) as remnants of the younger Somerville peneplain. The Piedmont was also viewed by Davis (1909) as representing a relict planation surface formed by synchronous processes and multiple erosion cycles no longer operating. Recent study, however, has cast some doubt on the existence of these surfaces (Hack, 1965; Poag and Sevon, 1989; Sevon, 1989a), owing largely to the lack of ancient deposits on ridgetops, lack of general accordance of summits when examined in detail, and lack of any direct dating of surfaces.

Critics of Davisian geomorphology argue that many previous geomorphological studies were largely historical, directed toward elucidating denudation chronology rather than to determining solid Earth processes that are currently operating (Flemal, 1971). Moreover, no large areas having the characteristics of a peneplains are presently known to exist with the exception of a postulated peneplain in the Ozark region (Bretz, 1962).

If indeed the concept of the peneplain has any application in the Appalachian Highlands, it seems most likely to apply in the Piedmont Plateau. Holmes (1964) suggested that the Piedmont Plateau represents positive evidence of former peneplanation, long-continued crustal stability, and baselevel control of erosion. He

interpreted the accordance of low-relief divides underlain by varying bedrock types as evidence of baselevel-controlled mass downwasting rather than the effects of geologic controls on weathering. Major structural trends on the Piedmont, he asserted, are predominantly north-northeast on the Piedmont while drainage trends are mainly south-southeast.

All peneplains that have been postulated thus far have been at least partially explained by structural, lithologic, or tectonic controls on landform development. Adherents of Davis contend that insufficient time has elapsed for the Cycle to run its course following disruption by Pleistocene climatic and eustatic changes. It is now generally conceded, however, that tectonic movements (Gilluly, 1949) and isostatic uplift in response to erosion (Anhert, 1970) have occurred sufficiently often in the Appalachians to preclude the long periods of quiescence required for the Davisian Cycle to come to completion.

Davis' concepts held sway among geomorphologists until the English translation of a 1924 work by Walter Penck (1953). Penck accepted the idea that landscapes could be reduced to endforms of low relief by a change of position of baselevel, but rejected the concept of cycles, maintaining instead that these endforms never became the initial forms for new episodes of dissection, primarily because uplift was assumed to be continuous rather than periodic. Because each landform was related to a certain period of uplift, a major goal of Penck's work was to determine regional tectonic history.

Penck challenged Davis' concept of downwasting of hillslopes (with slope angles decreasing or flattening through time) and suggested that slopes actually backwasted, retreating parallel to themselves and persisting until they eventually intersected. A condition of constant uplift was thought to result in a uniform rate of stream downcutting where slope profiles would retreat at a straight angle from the valley axis. Periods of accelerated uplift were expected to be marked by convex-skyward landforms resulting from enhanced stream downcutting and steeper slope gradients near the stream. Concave landforms would indicate a period of slower stream entrenchment and minimal uplift. Penck's ideas gained only nominal acceptance outside of Europe, but it is worth noting that the gently sloping surface between the main drainage line and the base of a retreating upland described by him is a characteristic feature of pediments which abound in arid and semiarid regions today.

The 1940's and 1950's realized a shift away from the historical approach to geomorphology toward a study of the empirical processes and relations between a landscape and its component landforms. King (1953) visualized the cyclic development of landscapes similar to Davis' model, but differed in emphasizing hillslope rather than fluvial processes in the evolution of landforms. He placed limits on the amount of hillslope flattening that would occur after initial stream incision, opting instead for a model similar to Penck's (1953) where scarps retreat parallel to themselves leaving pediments which would eventually coalesce, forming his ultimate cyclic landform - the pediplain. Pediplains were distinguished from Davis' peneplains by their multi-concave

instead of multi-convex topography, and while Davis would have suggested that this surface is important only in arid regions, King believed that the pediplain would be the principal landform in all climates. Problems of time, uplift, and the attainment of the penultimate stage that plagued Davis' system also make it unlikely that King's model will gain more than limited acceptance.

A time-independent, steady-state model of landform evolution was advanced and developed by Hack (1960; 1965) as an outgrowth of both Gilbert's (1877) work in the Henry Mountains of Utah and Nikiforoff's (1949) concept of soil evolution. Hack (1960) challenged the belief that lateral planation occurs in any drainage basin, irrespective of its geology, as a result of multiple cycles of erosion. Instead, he applied the name 'dynamic equilibrium' to the condition where differences and characteristics of landforms are interpreted as being the result of structural and lithologic, rather than temporal, controls on geomorphic processes that are presently acting.

Whereas Davis (1899) believed that the effect of structure was almost nullified as the evolution of topography proceeded beyond the mature stage, Hack viewed landforms as remaining adjusted through time to the geologic structures on which they occur and to processes which are active on them in a manner analogous to the concept of the graded stream, but on a larger scale. Hack maintained that total relief, not Davis' idea of base level, determines the rate of erosion and thus the steepness of slopes. Under ideal conditions in which the forms of the landscape remained in a perfect steady-state, erosion and isostatic uplift would be balanced such that downwasting of all parts of the landscape

would occur at the same rate causing topographic forms to remain static. If, for example, uplift was rapid, relief would be high because of the greater potential energy required for erosion to balance the uplift. Alternatively, if uplift rates declined substantially, relief would be expected to gradually decrease. Because bedrock geology dictated erosion rates, maintenance of equilibrium conditions required that rocks of similar resistance be exposed at the surface.

Tectonic movement and, to a lesser extent, climatic fluctuation were thought to restrict ideal conditions from ever being achieved leading Hack (1965) to later stipulate that every area has some “features that are in equilibrium and some that are not.” If relative rates of erosion changed in response to uplift or climate change, the state of equilibrium would also be expected to change, requiring modification or obliteration of the previous form. Factors such as uplift or climate that may indeed change with time, however, were believed to operate gradually enough so that a reasonable balance could be maintained between erosion and rock weathering such that topographic form could be maintained even though the system may be adjusting to new conditions. The adjustment of form to process and the return of equilibrium conditions, at least for some if not all, parts of the landscape, was expected to occur relatively quickly, without involving any particular cycle or succession of changes through which landforms inevitably evolve.

Rather than a representing the highly dissected remnant of a peneplain, the Piedmont Plateau represented to Hack a ‘ridge and ravine topography’ which existed, to use Davisian terms, in a state of perpetual maturity. Relict landforms were expected to be

preserved in the landscape only if unlikely sudden changes occurred, until a new steady-state is achieved. He acknowledged, however, that a condition of disequilibrium did indeed exist between the Blue Ridge escarpment and the Piedmont, indicating that the escarpment is fixed by geologic/structural contacts that are in some places identical.

An attempt to accommodate both the cyclic and non-cyclic schools of thought was tendered by Howard (1965) and Schumm and Lichty (1965) who viewed the distinction between dependence and independence of geomorphic forces as relevant only after considering the span of time involved and the size of the geomorphic system under investigation. This approach partially vindicated previous supporters of baselevel control of erosion by suggesting that it may be “applicable within the broad context of geologic time” (Holmes, 1964). In their model, Schumm and Lichty (1965) viewed geomorphic processes operating within relatively long geologic time spans as being ‘cyclic’ in nature and encompassing an entire Davisian erosion cycle within areas the size of entire drainage systems. They viewed ‘graded’ time spans as being short spans of cyclic time wherein a graded condition or dynamic equilibrium exists with regard to processes acting on landforms. Cyclic or progressive changes affecting entire landscapes, when viewed from this shorter span of time, were interpreted as a series of adjustments to the equilibrium condition between specific landform elements. A graded condition was thus restricted to components of drainage systems. Geomorphic factors operating within graded time were considered time-independent so that any relict features of the landscape were the result of processes occurring during longer, cyclic time. Although increasing

amounts of the landscape may approach dynamic equilibrium as the erosion cycle progresses, entire drainage systems were not expected to reach a graded state because progressive reduction of relief and changes to baselevel would ultimately preclude long-term maintenance of topographic form. A 'steady' time span represented yet another shorter time interval where, when viewed on an even smaller scale, the cause-effect relationships between geomorphic forms and processes were at true steady-state (e.g., the balance of sediment deposition to the upstream side of a point bar and the equal removal of material from the downstream side).

Origin and Age of the Piedmont Landscape

Cleaves and Costa (1979) and Costa and Cleaves (1984) applied the concepts of Schumm and Lichty (1965) to the problem of Piedmont morphogenesis in Maryland. Within short periods of geologic time (10^0 - 10^5 yr) they allowed that hillslopes, stream reaches, and smaller-scale features may be in equilibrium with processes acting upon them; however, they stipulated that changes in baselevel, climate, and erodibility of surficial materials would preclude entire drainage basins or geomorphic provinces from attaining steady-state conditions.

Evidence cited for landscape equilibrium included (1) control of first- and second-order stream alignment by bedrock jointing and foliation, (2) control of third- and fourth-order streams by regional structural trends, (3) adjustment of local relief to lithology, particularly in areas of highly contrasting bedrock, and (4) correlation of saprolite

thickness with landscape position and rock weatherability (e.g., mineralogy, texture, and degree of foliation). Complex secondary mineral suites in both this study and the study by Bricker et al. (1968) in a nearby area reflected apparent equilibrium with respect to microenvironmental differences in factors such as topography and internal drainage, suggesting rapid interaction between groundwater and primary minerals throughout the weathering zone.

However, Costa and Cleaves (1984) also cite the multiple record of Late Miocene, Pliocene, Pleistocene, and Holocene fluvial sediments on the Coastal Plain as evidence of periodic vigorous erosion of the Piedmont. In the Rappahannock River basin of Virginia, these terrace deposits are separated by scarps and show progressive weathering with increasing age, suggesting periods of stability between erosion events. Costa and Cleaves (1984) also argued that the Piedmont itself contains a record of long-term regional erosion and weathering. They viewed the uniform, low-relief Piedmont surface as a highly dissected peneplain that truncates rocks of varying lithologies, which suggested to them minimal structural control on landforms. They suggested, as did Holmes (1964), that the thick mantle of saprolite on the Piedmont was the result of a long period of crustal stability and minimal erosion, concluding that the present upland Piedmont surface in Maryland is a relict feature resulting from valley incision by fluvial erosion processes initiated during the Late Miocene (11-5 Ma B.P.), and periglacial mass-wasting processes during the Pleistocene. They attributed the presence of alluvial fans and other clastic stream deposits that are unrelated to any erosional processes presently active to

these processes. Cleaves (1989) later stipulated, however, that conclusive evidence of periglacial features on the Maryland Piedmont is currently unavailable.

Pavich and co-workers (Pavich, 1986, 1989a,b; Pavich et al., 1985) have disagreed with the approach of Costa and Cleaves (1984), suggesting that a synchronous erosion surface, or its remnants, cannot exist across large areas ($>10^4$ km²) unaltered over time periods $>10^7$ yr, arguing instead that the rapid adjustment of form to process occurs on all scales. Pavich (1986; 1989a) applied Hack's (1960) concept of dynamic equilibrium to the problem of Piedmont morphogenesis, suggesting that regolith production, pedogenesis, chemical and mechanical erosion, and uplift operate concurrently to limit time-dependent development of penultimate landforms and to produce a steady-state landscape. Long-term maintenance of topographic form on the Piedmont is attributed to the following three active processes which are essentially driven by isostatic uplift in response to erosion: (1) thinning of the regolith profile and lowering of the geomorphic surface by volume reduction during the transformation of saprolite into soil, (2) mass loss from the surface by particulate erosion, and (3) loss of mass in dissolved solids that leach through the solum. Because only about 80% of the surface lowering produced by these denudation processes is compensated for by isostatic rebound, the overall Piedmont surface is viewed as being continually lowered while topographic form is maintained more or less constant.

In effect, the lowering of the soil surface by erosion is seen to occur at a rate equal to the lowering of the soil/saprolite boundary and the descent of the weathered rock zone

into the underlying bedrock. Rock type and structure is viewed as controlling both denudation rates and regolith production primarily by the influencing the rate of water flux through the weathering zone. Regolith or saprolite formation is thus considered to be a self-limiting process that is greater in areas underlain by foliated, weatherable rocks than in areas of resistant, fine-grained, massive rocks that restrict hydrologic movement.

No permanent record of previous (relict) pedological conditions are expected to be preserved in the soil because this self-limiting equilibrium between erosion and soil/ saprolite production limits the age of the surface and prevents soil development from coming to some 'endpoint'. This model would explain why thick E horizons or lateritic soils do not develop on the Piedmont, and in much of the tropics, despite favorable climatic and tectonic conditions (Pavich, 1986). These theoretical considerations suggest that age of the Piedmont surface is considerably younger than 10^7 yr - the age of the oldest (Late Miocene) terrace soils on the Coastal Plain.

Whether or not apparent equilibrium conditions occur at all, let alone equally on all parts of the entire Piedmont remains open to question, at least within relatively recent time frames. Hack (1982) reported possible long-term differences in stability for northern vs. southern areas of the Piedmont based on average valley width and the fact that Piedmont streams south of the Roanoke River entered the Coastal Plain at grade whereas those in the north were considerably above grade. He invoked differential uplift along the Fall Zone to explain these effects, implying that equilibrium between saprolite development, erosion, and uplift has been occurring longer in the southern part of the

Piedmont than in the Mid-Atlantic region. This hypothesis, in addition to proximity to the glacial border, may also partially explain why northern areas on the Piedmont have appreciably more transported surficial deposits than areas farther south.

The controversy regarding the stability and residence time of the upland Piedmont regolith has largely been fueled by the widely disparate rates of saprolitization and erosion reported in the literature. Costa and Cleaves (1984), for example, estimated the age of the upland surface of the Maryland Piedmont to be as old as Late Miocene based on sedimentological studies of fluvial Coastal Plain sediments. Pavich and co-workers (Pavich, 1986, 1989a,b; Pavich et al., 1985, 1989; Markewich et al., 1987) invoked a Quaternary age or an average residence time of ~1-5 Ma for a typical 15 m-thick regolith profile. They used several methods of age estimation including: (1) evaluation of the rate of rock weathering and saprolite formation using base flow chemistry in a watershed study, (2) inventory of cosmogenic ^{10}Be in the residual soil and saprolite, and (3) comparison of morphological properties of Piedmont soils with those developed on fluvial terraces. Despite considerable variability in the assumptions and approaches implied with these methods, all estimates agreed within a factor of three, suggesting a saprolitization rate of 4 to 20 m Ma^{-1} . These values agree with Genthner's (1990) estimate of erosion in the Appomattox River basin of central Virginia. Cleaves et al. (1970) also reported similar values, but later recalculated the original data to model minimum and maximum saprolitization rates based on both modern environmental conditions and those of the Late Wisconsin glacial environment (Cleaves, 1989). He

reported in the latter study that 80 to 254 m of saprolite (2 to 4 times the amount that is currently present) may have formed on the Piedmont since the Late Miocene, suggesting that at least three major tectonic/erosional episodes, where fluvial erosion greatly exceeded chemical denudation, must have occurred in order to strip this volume of saprolite from the upland surface. The data of Pavich and others (Pavich, 1985; 1986; Pavich et al., 1985; Brown et al., 1988), however, indicate that only about 5 m Ma^{-1} have been stripped from the Piedmont surface, which indicates that it is unlikely that 80+ m of saprolite could have formed since the Late Miocene as Cleaves (1989) suggests.

Recently, Poag and Sevon (1989) examined the record of Atlantic continental shelf sediments and hypothesized that three irregularly spaced tectonic/quiescent cycles must have occurred since the Late Triassic (230-208 Ma B.P.), but the authors indicated that this series of tectonic cycles is incompatible with the traditional concept of four post-Triassic, Davisian geomorphic cycles postulated for the central Appalachians. In view of the fact that the ocean basins are no older than ~200 Ma, and the oldest plains on Earth have developed over much the same time span (Selby, 1985), it seems unlikely that the ocean record of chemical and clastic deposition studied by Poag and Sevon (1989) is related to any currently-existing landform or physiographic province.

Overstreet et al. (1968) recognized that the age of the Piedmont saprolite is obscure, but they speculated that it is probably chiefly Late Cretaceous and Tertiary in age based on the record of Coastal Plain formations that rest unconformably on saprolite. However, a reasonable argument against this hypothesis is that saprolite formation is an on-going

process that continues today even below more recent overburden. By this reasoning, saprolite production may post-date deposition of these ancient sediments in spite of landscape inversion. The interested reader is referred to Sevon (1989a,b) for more thorough reviews of erosion and regional landscape lowering that may bear on these concepts of Piedmont morphogenesis.

Transported Surficial Deposits

While there is clearly no consensus as to the age of formation of the Piedmont saprolite, numerous reports in the literature of colluvial or otherwise reworked deposits of Quaternary Age occupying both lower, depositional slope positions and higher-lying erosional surfaces of the nonglaciated Piedmont upland landscape (Parizek and Woodruff, 1957; Eargle, 1940, 1977; Overstreet et al., 1968; Whittecar, 1985; McCracken et al., 1989; Genthner, 1990) cast considerable uncertainty on the long-term stability of the upland Piedmont surface as well as aspects of both equilibrium and cyclic landscape models previously discussed.

Parizek and Woodruff (1957) reported extensive areas of soils with stone lines (carpedoliths) commonly occurring on upper portions of interfluves on the northern Georgia Piedmont. They interpreted carpedoliths as being a residual veneer of lag gravels after the winnowing of finer materials from the soil surface. Gravels were subsequently buried principally by sheet wash and colluvial action, presumably in response to late Cenozoic climatic conditions conducive to enhanced surface runoff. Eargle (1977)

documented colluvial soils of 3 to 6 m in depth which typically covered poorly developed paleosols or organic deposits on the Piedmont near Spartanburg, South Carolina. These colluvial soils had previously been mapped as residual soils of the Cecil and Appling series. Eargle (1977) and others (Ciolkosz et al., 1979; Mills, 1977, 1982) have attributed buried paleosols in colluvium to periods of geomorphic instability and rapid erosion alternating with periods of Pleistocene slope stability and pedogenesis.

Overstreet et al. (1968) provided evidence and reviewed the literature documenting that saprolite was commonly overlain by colluvium of Quaternary age throughout the Inner Piedmont belt in Georgia, South Carolina and North Carolina, particularly on double concave slopes. These nonfluviatile deposits were reported to have a thin, discontinuous layer of angular quartz fragments, to range from 10-30 feet in thickness and to thin upslope and merge imperceptibly with residual deposits on interfluves. They indicated that the upper surface of the saprolite is generally sharply divided from the overlying colluvium and does not conform to the present topography. McCracken et al. (1989) used similar criteria to infer that three previous cut-and-fill cycles had resulted in the present topography of their study site in North Carolina.

Whittecar (1985) studied alluvial and colluvial deposits along a 22 km exposure of trench dug across the Piedmont of north-central Virginia. Stony colluvium of uniform thickness (0.5 to 1 m thick) covered about 60% of the landscape cut by the trench, commonly truncating saprolite, bedrock, or alluvium with an abrupt boundary. No indication of buried soils such as those that were reported in South Carolina (Eargle,

1977) were preserved in the Virginia colluvium, although some indication of possible time discontinuities and episodic deposition was suggested by abrupt changes in coarse fragment content through the profile. The cause of episodic colluviation was speculated to be slope instability induced by Pleistocene periglacial climates although no conclusive evidence for landscape cycles involving glacial/interglacial fluctuation has been found in the Piedmont as has been the case for areas closer to the glacial margin (Ciolkosz et al., 1979).

Genthner (1990) studied soils on gently rolling (5 - 12%) sideslopes in three central Virginia Piedmont counties and found that over 50% of delineations mapped in each area as residual Appling, Cecil, and Davidson soils were actually formed in colluvium of up to 2 m in thickness. Slope deposits extended well up the backslope and onto convex upper shoulder slopes. Stones lines, ranging from one stones's thickness up to 49% (by volume) coarse fragment content and consisting of either angular vein quartz or greenstone, were ubiquitous throughout the study areas. The author attributed these features to widespread stripping of the landscape induced by regional slope instability following Pleistocene climatic fluctuations.

Previous Research on Piedmont Soil Genesis and Mineralogy

Early research on Red-Yellow Podzolic soils of Piedmont involved documentation of clay mineral suites, weathering transformations, and surface charge properties of soils derived from both felsic and mafic crystalline rocks (Alexander et al., 1941; Coleman et

al., 1949, 1959; Cady, 1950; Rich and Obenshain, 1954; Rich, 1958, Bryant and Dixon, 1964), studies of soil genesis and classification (Simonson, 1949; McCaleb, 1959; Gibbs and Perkins, 1966), and characterization of map unit composition (Powell and Springer, 1965).

Because the Piedmont regolith has been stable longer than comparable uplands in the glaciated areas of the U.S., much of our current knowledge of mineral transformations and weathering in residual soils is the result of many year's research on Piedmont soils. Rich (Rich and Obenshain,1954; Rich,1958) was among the first researchers to demonstrate that hydroxy-Al polymers provided a steric hindrance to the collapse of vermiculite interlayers in a soil formed from muscovite schist in the Virginia Piedmont. This hydroxy-interlayered vermiculite (HIV) was later shown to be nearly ubiquitous in highly weathered soils of the southern U.S. and elsewhere (Rich, 1968; Vincente et al., 1977; Karathanasis et al., 1983). Harris et al. (1984) showed that Fe-oxide coatings produced during weathering of primary minerals could effectively armour highly weatherable (talc) minerals in an intensely leached Piedmont Ultisol formed over deep saprolite, thus preventing their depletion. Harris et al. (1985a,b) also documented mechanisms of the pseudomorphic replacement of biotite by kaolinite, demonstrating that this weathering pathway was the dominant form of biotite transformation on the Piedmont. Recent work involving the pedogenesis of boreal soils at high elevations in the southern Blue Ridge has implied that biotite weathering is temperature-dependent,

with vermiculitization being the dominant weathering mechanism in cold climates (Feldman et al., 1991).

Other studies of the genesis and mineralogy of Piedmont soils have largely involved consideration of deep weathering profiles comprising both soils and saprolite. Calvert et al. (1980a,b) conducted a detailed study of mineral alteration pathways in a Pacolet soil/saprolite profile formed from granitic gneiss in the North Carolina Piedmont. Results documented decreases in bulk density and relative increases in Si and Al associated with the transformation of saprolite to soil. The most rapid changes in bulk chemistry occurred in the deepest saprolite horizons near the altering rock, thus confirming the earlier work by Alexander et al. (1941) which showed that gibbsite is primarily formed at the expense of feldspar dissolution and that kaolinite is commonly the result of resilication of this material, often through a halloysite intermediary.

Iron oxide mineralogy of soils formed from similar granitic parent material on the North Carolina Piedmont was studied by Bigham et al. (1978a,b) who demonstrated that equal amounts of both hematite and goethite were present in red (2.5YR) Cecil soils, whereas only an Al-rich goethite was present in yellow (7.5YR) Appling soils. Surface areas and consequently anion adsorption capacities of the yellow, goethitic clays were consistently higher than those of the redder soils. Results suggested that soil goethites exist in a more finely divided state than do soil hematites, and that soil color is a meaningful criterion for taxonomic and management separations of Piedmont and other highly weathered soils.

Effects of Climate Change on Pedogenesis and Weathering

Soil properties are dramatically influenced by different temperature and moisture regimes in the southeastern U.S., largely through effects on solubilities of solid and gaseous phases, chemical reaction rates, and speciation of ions in soil solution (Buol et al., 1990). These factors, and additionally the rate at which drainage water leaches through the soil, dictate to a large extent the nature of primary and secondary mineral weathering, degree of soil profile development, magnitude and depth of clay illuviation, soil color, and surface charge properties of the soil exchange complex.

Abrupt changes in the weathering environment within the time frame of global change concerns (e.g., 10 to 10^0 yr) could affect levels of soluble and exchangeable nutrients, rates of organic decomposition, and thus the amount and quality of organic C and N in soils. In the longer-term (e.g., 10^2 to 10^4+ yr), predicted increases in global climate would affect these and other parameters including the amount and nature of hydroxy-Al or -Fe polymers incorporated into vermiculite interlayers, and the crystallographic properties and stability of authigenic kaolinite and both crystalline and noncrystalline Fe- and Al-oxides. This would significantly modify soil physicochemical properties resulting, partly, in reduction of CEC, increases in anion retention characteristics, increases in K^+ selectivity, and changes in flocculation/dispersion phenomena which influence soil structure, surface crusting, and rates of infiltration, runoff, and erosion.

Kaolinite formation in southeastern soils is related to: i) pseudomorphic replacement after biotite or feldspars, ii) resilication of gibbsite, iii) dissolution of 2:1 phyllosilicates, or iv) reprecipitation (or coprecipitation with Fe) from solution. Using the method of Mattigod and Kittrick (1980), it can be shown that a temperature increase of 4.5°C would result in a ~10% increase in the concentration of aqueous Si governing gibbsite-kaolinite equilibria in the system $K_2O-Al_2O_3-SiO_2-H_2O$. This means that at a temperature of 25°C, kaolinite would weather to gibbsite if Si levels in solution drop to <0.177 ppm. However, at a temperature of 29.5°C, kaolinite becomes unstable at aqueous Si levels as high as 0.195 ppm or less. Similarly, K-feldspars, muscovite, and more surface-reactive clay minerals also become destabilized relative to both gibbsite and kaolinite as temperature increases. Calculations also indicate that kaolinite becomes more stable relative to gibbsite as ambient moisture and the activity of water decreases.

Although temperature effects on the thermodynamic stability of kaolinite can be accurately predicted, we know very little about its effect on the pseudomorphic alteration of primary silicates. Harris et al. (1985a,b) has shown that the dominant alteration product of biotite in mesic (MAST¹ 8° to 15°C) and warmer southeastern soils is kaolinite. This pseudomorph retains the optical properties of biotite (e.g., $2V\alpha=5^\circ-15^\circ$, brown/green pleochroism, etc.) but displays a 7.2Å x-ray diffraction spacing characteristic of kaolinite. Conversely, Feldman et al. (1991) found that vermiculitization through an interstratified intermediary, and not kaolinization of biotite,

¹ Mean annual soil temperature (at 50 cm depth)

was the dominant weathering mechanism in the frigid (MAST <8°C) soils at the higher elevations of the southern Blue Ridge. The results of Ghabru et al. (1989) for Canadian Spodosols corroborates this temperature-dependence of biotite weathering, but quantitative limits for this effect and the influence of high levels of soil organic acids are not known.

Although thermodynamic considerations dictate that hydroxy-interlayered vermiculite (HIV) will ultimately weather to kaolinite (Karathanasis et al., 1983; Barnhisel and Bertsch, 1989), the observation that HIV predominates in surface horizons while kaolinite dominates in subsurface horizons where weathering intensity is more moderate is also difficult to explain. The problem has been suggested as being kinetically based (Barnhisel and Bertsch, 1989). Harris et al. (1980) found that the amount of HIV increased with age relative to kaolinite in soils along a chronosequence formed in New River alluvium. Work by Markewich et al. (1987) on a chronosequence of eight soils formed in fluviomarine sediments confirmed this relationship on all but the oldest (>10⁶ yr) soil of their study. The decrease in HIV in this soil was associated with increases in both Al₂O₃ and Fe₂O₃ relative to SiO₂ suggesting that desilication of 2:1 phyllosilicates is age-dependent.

In addition to promoting higher rates of Al and Fe hydrolysis and polymerization and enhancing hydroxy-Al or -Fe interlayering in vermiculite (Carstea, 1968; Zelazny, 1970), increasing soil temperature has been shown to result in accelerated weathering of primary and secondary soil minerals (Jackson and Sherman, 1953) which affects the type and

distribution of residual oxides in soils. These factors may potentially cause a northward shift in the distribution and extent of kandic, oxidic, and kaolinitic soils (Soil Survey Staff, 1990) in the southeastern U.S. Expanded areas of these low-fertility soils would result in greater potential for groundwater pollution through higher rates of fertilizers, herbicides, and insecticides required to sustain agricultural productivity.

Because the nature and distribution of secondary Fe-oxides in soils reflects the immediate environment of pedogenesis, these components are expected to adjust rapidly to climatic changes and thus are useful criteria for estimating the magnitude of pedogenic weathering of soil parent materials. Higher temperatures and increasing water deficit systematically favor hematite formation over that of goethite (Kampf and Schwertmann, 1982a; Schwertmann and Kampf, 1985). Although moisture deficit can decrease the chemical activity of water in the weathering zone, variation in temperature was found by Torrent et al. (1982) to be more important than relative humidity in laboratory experiments on the crystallization of goethite and hematite from ferrihydrite.

The influence of temperature and leaching intensity also explain the observed and reported distribution of hematitic (reddish), goethitic (yellowish-brown), and gibbsitic soils (Bigham et al., 1978b). Soil rubification, weathering intensity, and degree of profile development on similar parent materials increase noticeably from north to south on the Piedmont (Kaster, 1960; Porter et al., 1963; Frost, 1979) in conjunction with both reported and observed increases in hematite and gibbsite, and decreases in goethite in this same direction (Southern Regional Project S-14, 1959). This transition also closely

coincides with a shift from mesic soils (MAST 8° to 15°C) in the northern part of the Piedmont to thermic soils (MAST 15° to 22°C) in the southern part, and a similar change from soils having dominantly mixed mineralogy to those having kaolinitic mineralogy (Soil Survey Staff, 1990).

Goethite in soils is frequently associated with ferrihydrite, a fine-grained paracrystalline precursor of hematite and goethite which is a characteristic component of young Fe-oxide accumulations or transformations (Schwertmann, 1989). Ferrihydrite's relative instability in the weathering environment suggests that it occurs in soil systems that are not in thermodynamic equilibrium (Schwertmann, 1985). Its tendency to form in cooler, more temperate climates suggests that its presence may also be a sensitive indicator of changing pedoenvironmental conditions (McFadden, 1988; Schwertmann, 1989). High solution concentrations of Si (within a range commonly occurring in natural waters) interfere with formation of goethite and favor the formation of ferrihydrite in a given weathering environment (Schwertmann, 1989). Consequently, ferrihydrite is less likely to form in the southern range of the study area where warmer, moister conditions have resulted in greater leaching intensity and desilication over time. Decreases in ferrihydrite and increases in both goethite and hematite contents in warmer climates are also associated with a systematic increase in the crystallinity of these secondary Fe-oxides.

The degree of Al substitution in soil goethite and hematite increases at higher temperatures and has been correlated with both age and paleoenvironment (Fitzpatrick

and Schwertmann, 1982). The range of Al substitution in goethites lies between 0 and 33 mole % and reaches a maximum in highly weathered soils of subtropical and tropical areas (Fitzpatrick and Schwertmann, 1982; Schwertmann, 1985). Though goethites and hematites are thermodynamically and kinetically more stable in the weathering environment with increasing Al substitution (Torrent et al., 1987; Macedo and Bryant, 1989; Bryant and Macedo, 1990), a significant negative relationship between Al substitution and degree of crystallinity was found by Fitzpatrick and Schwertmann (1982), suggesting that differences in the weathering environment affecting Al availability are reflected in crystal growth and morphology. Recent study has also shown that the mean crystallite dimension (MCD_{hkl}) of hematites synthesized in the presence of Al are greater along x-y crystallographic axes (platy particle morphology) than hematites synthesized in Al-free systems where homogenous growth (spherical particles) was observed (Fontes and Weed, 1991).

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Chapter III

Late Pleistocene Eolian Activity and Post-Depositional Alteration on the Piedmont of Northern Virginia

ABSTRACT

A ~1m thick loess cap overlying a stone line and buried paleosol formed in granitic residuum records a period of changing geomorphic conditions in the mid-Atlantic area during the close of the Pleistocene. Thermoluminescence dating and palynological evidence indicates that the loess was emplaced in a cool, dry, high wind environment at 13.8 ± 1.0 ka BP when open tundra was <175 km to the north. Glacial outwash from the Susquehanna River basin and nonglacial alluvium from the Potomac basin are plausible sediment source areas for the loess.

A weakly-developed argillic horizon formed in the loess contrasts sharply with the paleoargillic horizon, which exhibits dark illuvial clay flows and relict organic coatings on coarse prism faces. Clay-free silt content decreases abruptly from ~55-60% in the loess cap to <25% in the underlying paleosol, concurrent with similar abrupt increases in CEC, reductant-soluble Fe, and exchangeable Mg below the stone line. Mineralogical analysis identified the presence of a high-charge smectite phase in the paleosol, which is

inconsistent with long-term weathering under the modern conditions of intense, acid leaching. The formation of this mineral phase may be related to drier climatic conditions at or before the time of loess emplacement.

Results also show that most of the loess was subsequently eroded from all but the most stable, level ridgetops in the area. Much of the loess was not entirely lost through erosion, however, but remains on the landscape and is incorporated into the existing soil through frost-driven colluvial processes. The majority of soils in the area are thus bisequal, having formed in an admixture of loess and colluvium over residuum with a basal mixing zone of up to 20 cm. Based on the record of widespread post-depositional periglacial activity in the area, we conclude that the transition from a colluvial-dominated to a fluvial-dominated geomorphic regime did not occur in the mid-Atlantic area until the end of the early Holocene.

Introduction

The majority of upland soils in the Piedmont Province south of the glacial border exhibit evidence of long-term surface stability, strong pedogenic development, and desilication as indicated by the dominance of clay minerals in advanced stages of weathering. Yet, for many years, reports in the literature have documented a considerable amount of reworked colluvium and occasionally eolian material on upper backslopes and 'stable' Piedmont interfluves as far south as Georgia (Darmody and Foss, 1982; Eargle, 1940, 1977; McCracken, 1989; Overstreet et al., 1968; Parizek and Woodruff, 1957;

Whittecar, 1985). These surficial deposits commonly truncate saprolite or a buried paleosol, marking an erosional surface that is frequently protected by a lag gravel veneer of angular vein quartz.

The widespread regional occurrence and morphological diversity of these transported deposits strongly suggests that major changes in the erosional stability of the Piedmont surface, most likely due to either climatic or tectonic fluctuations, have occurred throughout the Quaternary. However, the causes and timing of such events on the Piedmont, and indeed the wider Appalachian region, are complex and remain poorly understood, largely because of the limited number of datable deposits (Mills and Delcourt, 1991), uncertainty regarding weathering rates and soil profile development, and our inability to quantitatively differentiate between the complex effects of climate vs. geomorphic age on elemental flux in the weathering environment (Feldman et al., 1992).

In this paper, we document the existence of a dated eolian sediment in the mid-Atlantic area that marks the transition from a glacial to an interglacial climate during the late Quaternary. We characterize the genesis, nature, and significance of secondary products formed in the ~1 m thick remnant of a dated, discontinuous loess cap that overlies a residual paleosol in an area of the northern Virginia Piedmont. Finally, we provide evidence that post-depositional cryoturbation has caused reworking of contrasting parent materials throughout the local landscape, resulting in a preponderance of polygenetic soils that otherwise appear to be residual based on morphological evidence alone.

Site Description

We located a previously unidentified ~0.5-1.0 m thick exposure of loess on a convex interfluvium at an elevation of 97 m near Woodbridge, VA. This area lies in the outer Piedmont Crystalline Province (Thornbury, 1965) and is drained by the Occoquan River, which forms the boundary between Prince William and Fairfax Counties, VA (Fig. 1-1). Upland interfluviums in the area are typically convex and narrow (<100-300 m across), ranging in elevation from ~90-110 m and commonly underlain by saprolite that is greater than 15 m thick (Pavich, 1989). Major topographic and drainage features are primarily controlled by the structure and foliation of the underlying lithology. The area is highly dissected by third- and fourth-order streams, which are commonly incised into bedrock. Interfluviums are flanked by short, steep slopes comprised of convex shoulders, concave head slopes, linear backslopes, and convex nose slopes. Maximum local relief is ~75 m. The majority of the area is covered by deciduous forest, except for areas that are developed.

Present climate of the Piedmont is humid-temperate to subtropical (Thornbury, 1965). Mean annual temperature (MAT) of the area is 12.2°C (Fig. 1-2, 1-3). Mean annual precipitation is 1024 mm yr⁻¹ that is relatively uniformly distributed throughout the year (U.S. Dept. of Commerce, 1987). Calculated water balance data for the study area, based on MAT and calculated potential evapotranspiration (PET) values, show that soils undergo a summer moisture deficit and have only minimal capacity for deep percolation below the solum (Table 1-1, Fig. 1-3).

Materials and Methods

Fieldwork

A bisequal soil has formed in the loess where it overlies a buried paleosol derived from granitic saprolite (Table 3-1). A continuous stone line of angular vein quartz marks an erosional surface at the contact between these two parent materials. This soil (pedon VA 1) was described and sampled by genetic horizon at a representative location along a 2 m-deep vertical, recently-excavated cut-bank with an exposure of ~300 m along contour. Deeper samples (>5.5 m) were obtained at ~50 cm intervals using a modified bucket auger method (Stolt et al., 1991), which permitted bulk sampling and also sampling for bulk density determinations.

Because very few exposures of well-preserved loess remain on this eroded landscape, we sampled a satellite pedon (VA 2), located on a similar landscape position ~2 km away, in order to assess the effects of loess redistribution on pedogenesis in other area soils formed in the same parent material (Table 3-2).

Laboratory Analyses

Soil samples were air-dried, sieved to remove coarse fragments (>2 mm), and thoroughly mixed. Standard physical and chemical characterization included: i) particle-size analysis by the pipette method, ii) organic carbon by wet oxidation with dichromate and back-titration with 0.5 M $\text{Fe}(\text{NH}_4)_2(\text{SO}_4)_2$, iii) CEC by the sum of neutral M NH_4OAc -extractable bases (pH 7.0) plus total soil acidity by BaCl_2 -TEA at pH 8.2,

TABLE 1. MORPHOLOGY OF THE VA1 PEDON DEVELOPED IN LOESS OVERLYING A TRUNCATED PALEOSOL (*Site*: Upland ridgetop, slightly convex; *Elevation*: 97 m; *Aspect*: 82°SW; *Slope*: 2%; *Vegetation*: second-year fallow)

Horizon	Depth, cm	Description	Horizon	Depth, cm	Description
A	0-8	Dark yellowish brown (10YR 4/4) silt loam; moderate fine and very fine granular structure; very friable; many fine and very fine roots; abrupt smooth boundary.	2Bt4b (cont.)		few fine roots in cracks between prisms; gradual smooth boundary.
E	8-15	Grayish brown (10YR 5/2) loam; weak medium platy parting to very fine granular structure; firm, slightly brittle; many fine and medium roots; clear smooth boundary.	2CBb	109-178	Mottled yellowish red (5YR 5/6), strong brown (7.5YR 5/6), and yellowish brown (10YR 5/4) sandy clay loam; massive rock structure in place breaking to weak medium subangular blocky structure; friable; gradual smooth boundary.
BE	15-30	Yellowish brown (10YR 5/4) loam with common fine prominent strong brown (10YR 5/6) mottles in vesicular pores and cracks; few fine distinct grayish brown (10YR 5/2) weak medium subangular blocky structure; firm; common fine roots; clear smooth boundary.	2C1b	178-213	Mottled yellowish red (5YR 5/6), strong brown (7.5YR 5/6), and yellowish brown (10YR 5/4) sandy loam; structureless massive (oxidized granitic saprolite), friable, gradual smooth boundary.
Bt1	30-38	Strong brown (7.5 YR 5/6) heavy loam; common medium distinct yellowish brown (10YR 5/4) mottles; weak medium subangular blocky structure; friable; common fine roots; few patchy clay films on ped faces; gradual smooth boundary.	2C2b*	213-244	Mottled yellowish red (5YR 5/6), strong brown (7.5YR 5/6), and yellowish brown (10YR 5/4) sandy loam; structureless massive saprolite, friable.
Bt2	38-51	Strong brown (7.5 YR 5/6) heavy loam; weak medium subangular blocky structure; friable; few patchy clay films; ~8 cm-thick continuous stone line of unweathered, angular vein quartz present in basal zone; clear smooth boundary.	2C3b	244-264	Mottled yellowish brown (10YR 5/8), light gray (10YR 7/2), dark yellowish brown (10YR 4/4), and strong brown (7.5YR 5/6) sandy loam, massive, friable saprolite.
2Bt3b	51-74	Strong brown (7.5 YR 5/8) clay with common medium distinct yellowish brown (10YR 5/4) mottles; strong medium subangular blocky structure; firm; thick continuous strong clay films; few fine roots; gradual smooth boundary.	2C4b	284-295	Mottled yellowish brown (10YR 5/8), light gray (10YR 7/2), dark yellowish brown (10YR 4/4), and strong brown (7.5YR 5/6) loamy sand, massive, friable saprolite.
2Bt4b	74-109	Strong brown (7.5YR 5/8) sandy clay loam with common medium distinct yellowish brown (10YR 5/4) mottles; moderate very coarse and coarse prismatic parting to strong medium subangular blocky structure; firm; many thick continuous strong brown (7.5YR 5/6) clay films on blocky peds, common large, continuous brown (7.5YR 4/4) clay flows on prism faces;	2C5b	401-411	Dark and light gray weathered sandy loam granitic saprolite with dark yellowish brown (10YR 4/4) and yellowish brown (10YR 5/8) oxidized zones; rock-controlled structure; many mica flakes.
		(continued)	2C6b	498-508	Dark gray (10YR 4/1) and light gray (10 YR 7/1) weathered rippable granite; loamy sand.
			2C7b	508-528	Dark gray (10YR 4/1) and light gray (10 YR 7/1) weathered rippable granite; loamy sand.
			2C8b	528-564	Dark gray (10YR 4/1) and light gray (10 YR 7/1) weathered rippable granite; loamy sand.
			2C9b	564-594+	Dark (10YR 4/1) and light gray (10 YR 7/1) weathered rippable granite; loamy sand; pink feldspar grains evident.

* This horizon and underlying ones were sampled with an auger. Depths represent sampling intervals, not horizon thickness.

TABLE 2. MORPHOLOGY OF THE VA2 PEDON DEVELOPED IN COLLUVIUM OVERLYING RESIDUUM (*Site*: Upland ridgetop, slightly convex; *Elevation*: 97 m; *Aspect*: 90°E; *Slope*: 3%; *Vegetation*: second-year fallow)

Horizon	Depth, cm	Description	Horizon	Depth, cm	Description
Ap	0-13	Dark brown (10YR 4/3) loam; weak fine and very fine granular structure; very friable; many fine and medium fine roots; about 10% unweathered quartz gravels (<1 cm); abrupt smooth boundary.	2BC1b (cont.)		brown (10YR 5/8) loam; massive in place; friable; high mica content in unoxidized zones; gradual wavy boundary.
BA	13-20	Dark yellowish brown (10YR 5/4) loam; weak medium subangular blocky structure; friable, many fine and medium roots; about 10% unweathered quartz gravels (<1 cm); clear smooth boundary.	2BC2b*	234-244	Mottled light gray (10YR 7/2), yellowish brown (10YR 5/6, 5/4), strong brown (7.5YR 5/6), and yellowish red (5YR 4/6) sandy loam saprolite; continuous strong brown (7.5YR 5/6) clay films on blocky peds, common large, continuous brown (7.5YR 4/4) clay flows on prism faces; few fine roots in cracks between prisms; gradual smooth boundary.
Bt1	20-30	Yellowish brown (10YR 5/6) sandy clay loam; weak medium subangular blocky structure; firm; common fine roots; about 10% unweathered quartz gravels (<1 cm); clear smooth boundary.	2C1b	284-295	Mottled yellowish brown (10YR 5/8), strong brown (7.5YR 5/6), dark yellowish brown (10YR 4/4), and light gray (10YR 7/2) sandy loam; massive saprolite with common dark nodules; friable.
Bt2	30-46	Yellowish brown (10.5 YR 5/8) sandy clay loam; moderate medium subangular blocky structure; firm; common thin strong brown (7.5YR 5/6) clay films on ped faces; common fine roots; about 10% unweathered quartz gravels (<1 cm) throughout matrix; faint stone line of ~20% quartz gravels up to 7 cm across in lower 10 cm of horizon; clear smooth boundary.	2C2b	340-350	Mottled yellowish brown (10YR 5/8), strong brown (7.5YR 5/6), dark yellowish brown (10YR 4/4), and light gray (10YR 7/2) sandy loam; massive saprolite with common dark nodules; friable.
2Bt3b	46-91	Strong brown (7.5 YR 5/6) clay with common fine yellowish red (5YR 4/6) mottles; strong medium subangular blocky structure; firm; common thick continuous strong brown (7.5YR 4/6) clay films on ped faces; few fine roots; about 10% unweathered quartz gravels (<1 cm); clear wavy boundary.	2C3b	411-421	Mottled yellowish brown (10YR 5/8), strong brown (7.5YR 5/6), dark yellowish brown (10YR 4/4), and light gray (10YR 7/2) massive saprolite; sandy loam; common dark nodules; friable.
2Bt4b	91-147	Yellowish red (5 YR 5/6) clay loam with common medium distinct strong brown (7.5YR 5/8) and yellowish brown 10YR 5/4 mottles; massive in place parting to weak fine subangular blocky structure; firm; few thin clay films; gradual wavy boundary.	2C4b	498-508	Mottled yellowish brown (10YR 5/8), strong brown (7.5YR 5/6), dark yellowish brown (10YR 4/4), and light gray (10YR 7/2) massive saprolite; loam; common dark nodules; friable.
2BC1b	147-182	Mottled saprolite with banded zones of oxidized yellowish red (5YR 5/6) and strong brown (7.5YR 5/8) sandy loam material, and unoxidized zones of light yellowish brown (10YR 6/4), brownish yellow (10YR 6/6), and yellowish	2C5b	615-625	Mottled yellowish brown (10YR 5/8), strong brown (7.5YR 5/6), dark yellowish brown (10YR 4/4), and light gray (10YR 7/2) sandy loam; massive saprolite with common dark nodules; friable.
		(continued)	2C6b	767-777+	Mottled yellowish brown (10YR 5/8), strong brown (7.5YR 5/6), dark yellowish brown (10YR 4/4), and light gray (10YR 7/2) sandy loam; massive saprolite with common dark nodules; friable.

* This horizon and underlying ones were sampled with an auger. Depths represent sampling intervals, not horizon thickness.

and iv) free Fe by Na-dithionite-citrate-bicarbonate (DCB) analysis (Soil Survey Staff, 1984). Soil pH was measured in the supernatant portion of a 1:1 soil/deionized water suspension after a 1 h equilibration period. Samples were prepared for mineralogical analysis by pretreating with 30% H₂O₂, adjusted to pH 5 with *M* NaOAc to remove organic matter where necessary (Kunze, 1965), and with DCB to remove Fe-oxide coatings (Mehra and Jackson, 1960). Samples were then adjusted to pH 9.5 with *M* Na₂CO₃ to effect particle dispersion, and sand (>50μ) was separated by wet sieving. The remaining clay (<2μ) and silt (2-50μ) fractions were separated by repeated decantation after centrifugation. Very fine (0.05-0.1 mm) sand fractions were examined petrographically to quantify mineral suites by the line count method (Galehouse, 1971). X-ray diffractograms of clay fractions were obtained from Mg-saturated, glycerol-solvated aliquots with no heat treatment (25°C) and after 4 hours of heating at 110°C, and also from K-saturated samples with no heat treatment, and after 4 hours heating at 110°, 300°, and 550°C, respectively. Samples were scanned with CuKα radiation for 4.0s at 0.075° step⁻¹. Selected clay samples were also treated with formamide to differentiate kaolinite from halloysite by intercalation (Churchman et al., 1984). Kaolinite and gibbsite were quantified by mass-equivalent calibration of differential scanning calorimetry (DSC) endothermic peak areas using poorly crystalline Georgia kaolinite and Reynolds¹ synthetic gibbsite (RH-31F) as standards.

¹ Reynolds Metals Co. , Alumina Research Division, Bauxite, AR 72011

Polished thin sections of oriented clods taken from the loess horizons of pedon VA 1 were prepared and examined with a petrographic microscope. Radiometric age of the loess was determined by thermoluminescence (TL) dating using the methods outlined in Millard and Maat (1994).

Results

Soil Morphology and Age of the Loess Cap

Pedon VA 1 formed in 51 cm of a yellowish brown silt cap over a strong brown, clayey paleosol derived from the underlying granitic saprolite (Table 3-1). The silt cap was TL dated at 13.8 ± 1.0 ka BP making it contemporaneous with late Pleistocene loess deposits found on the Piedmont and eastern shore of Maryland (Foss et al., 1978; Darmody and Foss, 1982). Pedogenic alterations evident in the loess mantle include i) the formation of an incipient argillic horizon with a near doubling of clay content in lower horizons, and illuvial argillans evident in thin section, ii) development of weak subangular blocky structure, and iii) redistribution of free Fe from 0.4% in the surface to 1.4% Fe in the lower loessal horizons.

The continuous stone line of one stone's thickness at the base of the loess is derived from strongly dipping, truncated quartz veins in the underlying saprolite. Below this erosional surface, the highly developed paleoargillic horizon has primary and secondary soil structure, high clay content, and thick argillans. Dark, illuvial clay flows and relict organic materials coat coarse, vertical prism faces in this horizon. The transition from

soil to saprolite, marked by an increase in rock-controlled structure and mottling, begins at a depth of ~2 m.

The surface horizons of pedon VA 2 are morphologically more well developed than the loess cap of VA 1, with subsurface horizons having higher clay content and moderate degree of structural development (Table 3-2). Although sharp differences in parent material composition were not observed in the field, there is a faint stone line at a depth of ~46 cm that separates an upper colluvial mantle from the residuum below. The stone line is discontinuous in the sampled pit, but it is traceable in several nearby roadcuts, and has been found at other locations throughout the area at approximately this same depth.

Physical and Chemical Characteristics

The intact loess cap of pedon VA 1 is dominated by ~45-50% silt, which abruptly decreases to <25% below the stone line and remains uniform with depth in the underlying paleosol and saprolite (Fig. 3-1a,b,c). The loess is largely comprised of both medium (5-20 μ) and coarse (20-50 μ) silt particles (Fig. 3-1d,e,f), with evidence of mechanical abrasion, including subangular morphology, concoidal fracturing, and breakage features on grain surfaces (Fig. 3-2a). Silt grains from the VA 1 paleoargillic horizon, in contrast, have irregular edges and dissolution features characteristic of chemical weathering in place (Fig. 3-2b). Much of the fine silt (2-5 μ) fraction has been cemented after deposition into rosette-like forms of about 50 μ across (Fig. 3-2c). Because the loess is carbonate-free, as indicated by pH values <5 and low levels of exchangeable Ca (data not shown),

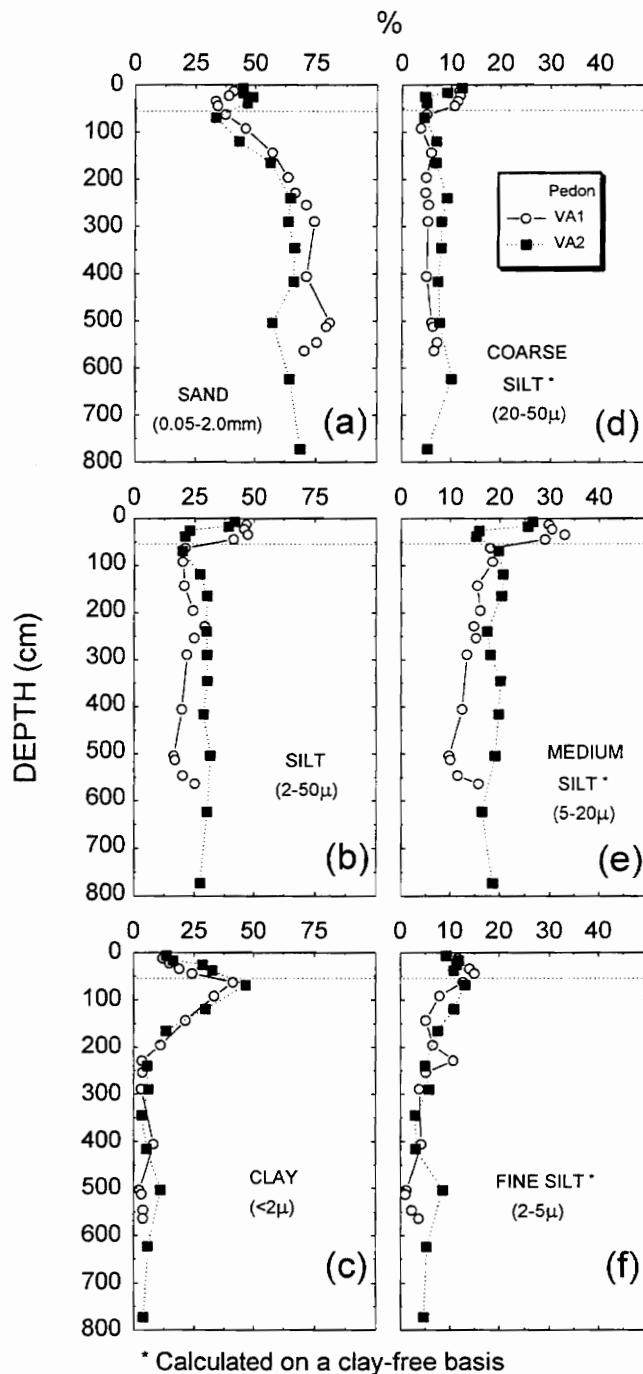


Fig. 3-1. Particle size distributions for the VA 1 and VA 2 pedons (a,b,c) and distribution of the silt sub-fractions, expressed on a clay-free basis (d,e,f). Thickness of the VA 1 loess cap is shown by the horizontal stippled line at ~50 cm.

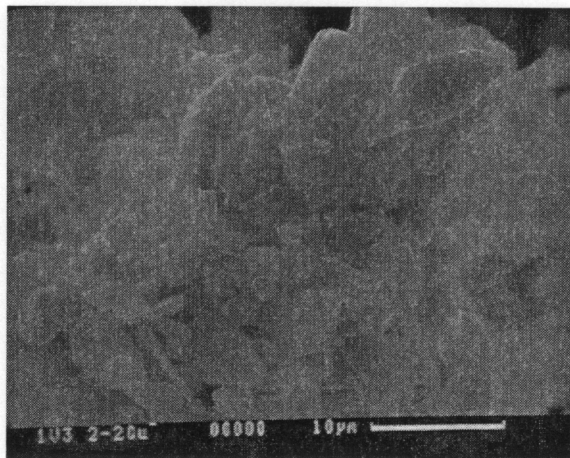
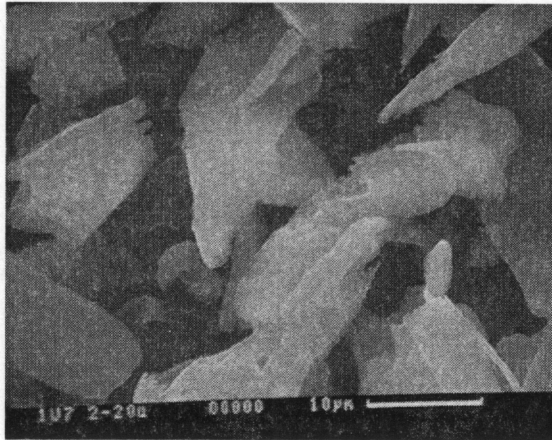
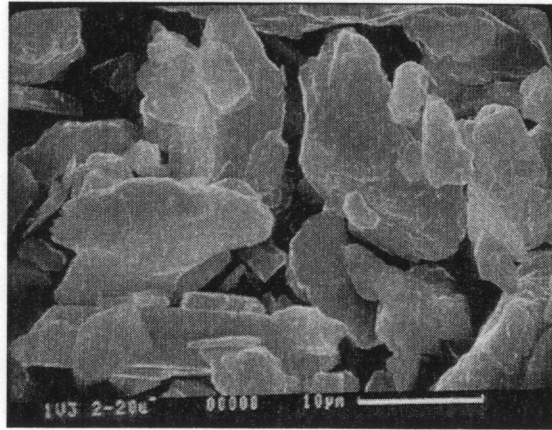


Fig. 3-2. Scanning electron micrographs of medium (5-20 μ) silt fractions from selected horizons of pedon VA 1. (a,c) the loess cap (8-15 cm) and (b) the paleoargillic horizon (74-109 cm).

Si is the probable cementing agent.

Both soils have nearly identical amounts of pedogenic clay and secondary Fe-oxides (Fe_d) (Fig. 3-3c), with maxima occurring immediately below the stone line in the VA 1 pedon at ~50 cm. Pedon VA 2, without a clear lithological discontinuity, shows an enrichment in coarse and medium silt similar to pedon VA1 in the surface and subsurface horizons only.

Exchangeable Mg and CEC increase abruptly in the paleosol, concurrent with decreasing silt content below the stone line of pedon VA 1 (Fig. 3-3b,c). Exchangeable Mg also increases noticeably in pedon VA 2 immediately below the colluvium/residuum boundary at 46 cm. The lower limit of clay, Fe_d , and exchangeable Mg accumulation at ~250-300 cm in both profiles indicates that this depth essentially defines the zone of long-term leaching, or the depth of the wetting front under which the paleoargillic horizon has formed in this part of the northern Virginia Piedmont.

Mineralogy of Contrasting Parent Materials

The very fine sand (0.05-0.10 mm) fraction of the undisturbed surface loess mantle is enriched in quartz and depleted in muscovite mica relative to the underlying paleosol (Fig. 3-4). K-feldspars range to ~25% of this fraction in the cap and persist in small amounts below the stone line as a result of downward infilling through desiccation cracks. Plagioclase feldspars are also present in amounts less than ~10% in the loess cap, and increase with depth below ~400 cm in the saprolite.

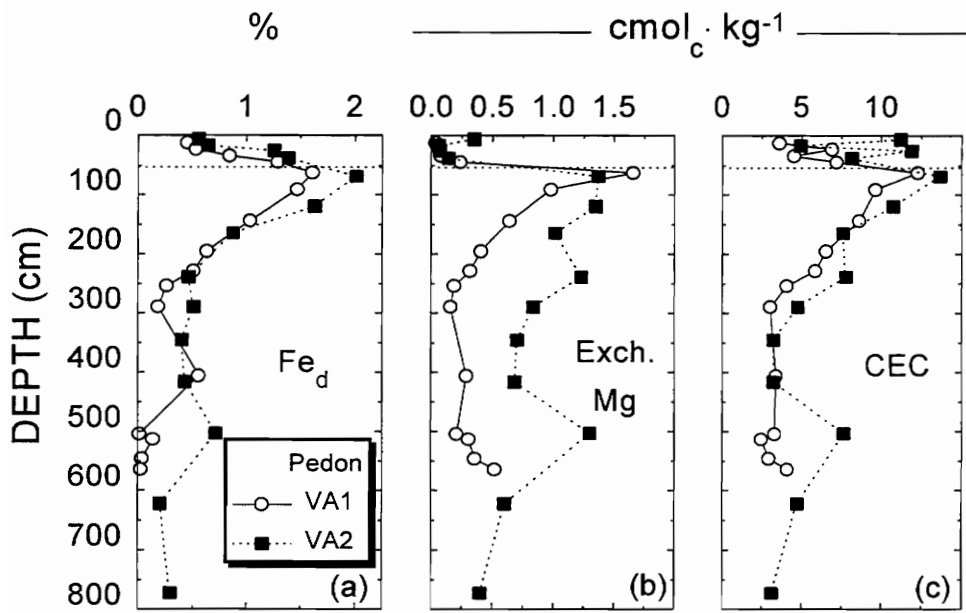


Fig. 3-3. Reductant soluble Fe (Fe_d), exchangeable Mg, and CEC for pedons VA 1 and VA 2. The VA 1 stone line is shown by the horizontal stippled line at ~50 cm.

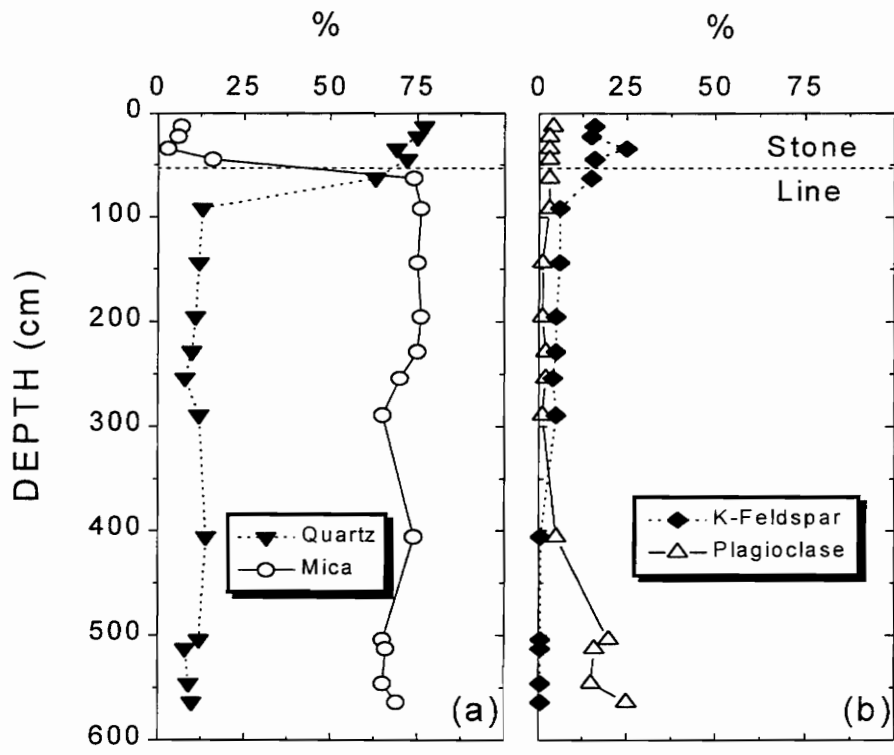


Fig. 3-4. Mineralogy of the very fine sand (0.05-0.10 mm) fraction for pedons VA 1 and VA 2. The stone line is indicated at ~50 cm.

The coarse silt (20-50 μ) fraction, which dominates the loess, is similarly comprised of quartz > muscovite > K-feldspar > plagioclase with only scant evidence of secondary clay mineral formation (data not presented). The clay fraction of the loess cap, however, is dominated by secondary phases in the early stages of weathering (Fig. 3-5). The minimal variation in clay mineralogy with depth implies rapid deposition of the loess, and indicates that relatively little pedogenic alteration has taken place since the time of deposition. Major minerals of the loess cap clay fraction include kaolinite (7.2Å + 3.45Å) \cong hydroxy-interlayered vermiculite (HIV) (14.6Å) > quartz (4.24Å + 3.33Å) > mica (10.0Å + 3.33Å) > regularly interstratified mica/vermiculite (RMV) (24.6Å + 12.3Å).

The clay mineralogy of the underlying paleosol is distinctly different from that of the surface cap in that it reflects a much more advanced stage of weathering, suggesting that the geomorphic surface prior to truncation was stable for a relatively long period. Kaolinite content, quantified by DSC, ranges from 24-30% in the loess where it formed both as an initial product of feldspar hydrolysis and from desilication of 2:1 phyllosilicates. Kaolinite content of the paleoargillitic horizons, in contrast, exceeds 75% of the total clay fraction. Its formation in these buried horizons is related to several complex mechanisms including desilication of 2:1 clays in near-surface horizons, resilication of gibbsite lower in the profile, and recrystallization of halloysite produced deep in the saprolite as a pseudomorphic transformation product of sodic plagioclase. The halloysite component of the 2C6b horizon (498-508 cm depth) clay fraction is

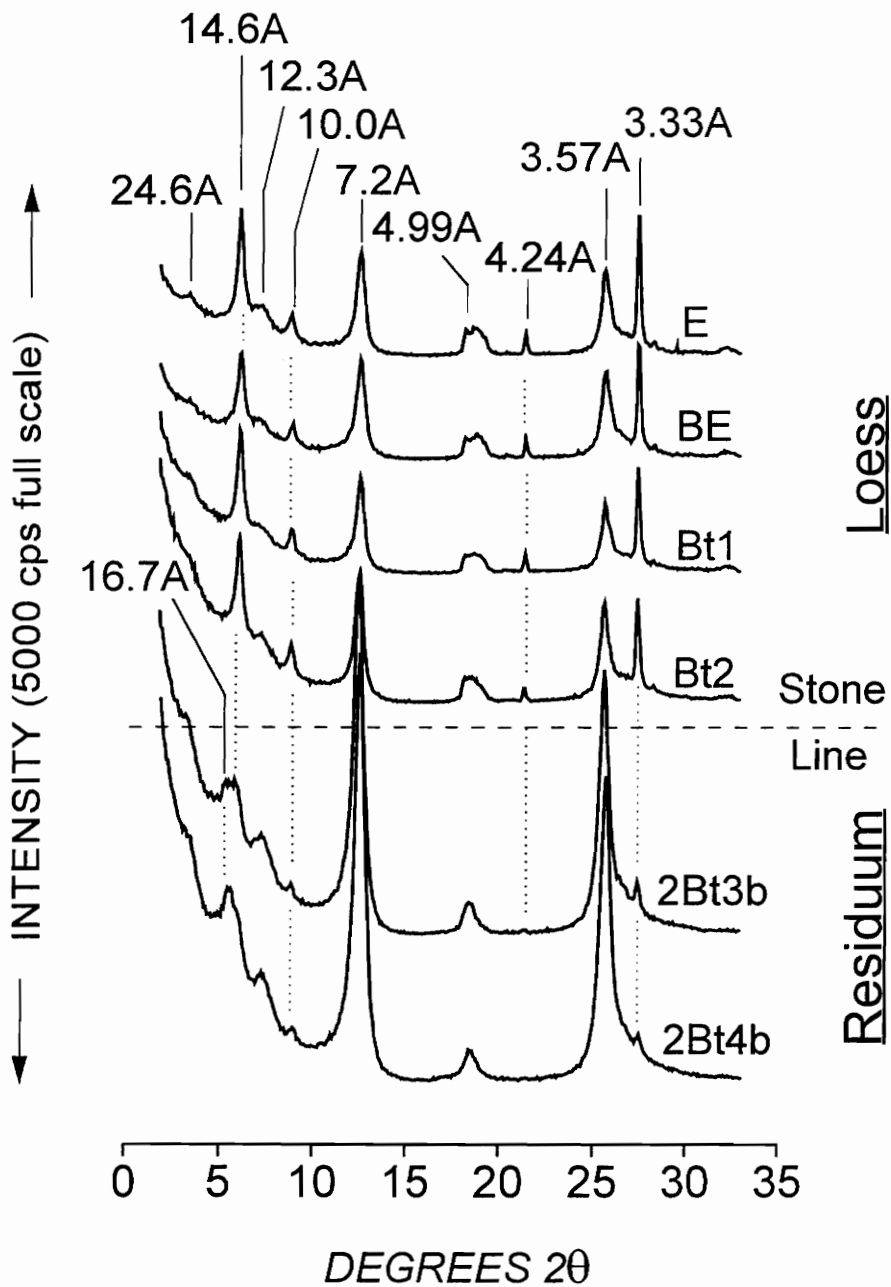


Fig. 3-5. Clay mineralogy of the pedon VA 1 loess cap and upper paleosol horizons. All samples were Mg-saturated, glycerol-solvated, and analyzed at 25°C.

identified in Figure 3-6 as a shift in the 7.2Å peak to 10.4Å corresponding to lattice expansion after selective intercalation with formamide. The clay fraction of the overlying 2C5b horizon at a depth of 400-411 cm shows considerably less expansion to 10.4Å, indicating that halloysite is transformed to kaolinite at this depth and higher in the profile.

In addition to being distinguished by its high kaolinite content, the paleosol also contains a high-charge smectite phase that is not present in the loess. This mineral is represented by a shift from the 14.6Å (HIV) peak above the stone line to the 16.7Å (smectite) peak below (Fig. 3-5). The smectite appears to be a product of mucovite weathering lower in the saprolite, and is responsible for the high CEC and high levels of exchangeable Mg in the upper paleoargillic horizons. Because the presence of this mineral is inconsistent with long-term weathering under the modern conditions of intense, acid leaching (Karathanasis and Hajek, 1983; 1984), smectite formation in the paleosol may be related to drier climatic conditions at or before the time of loess emplacement. Its preservation in the existing weathering environment may be an effect of i) inhibition of deep leaching through the paleosol resulting from the high silt content of the loess, which preferentially retains moisture until saturated conditions occur, and ii) slower chemical weathering rates resulting from lower temperatures and drier conditions during late Wisconsin time. Preferential water retention in the loess cap could also cause higher Si levels in solution, resulting in the observed cementation of silt particles after drying (Fig. 3-2c).

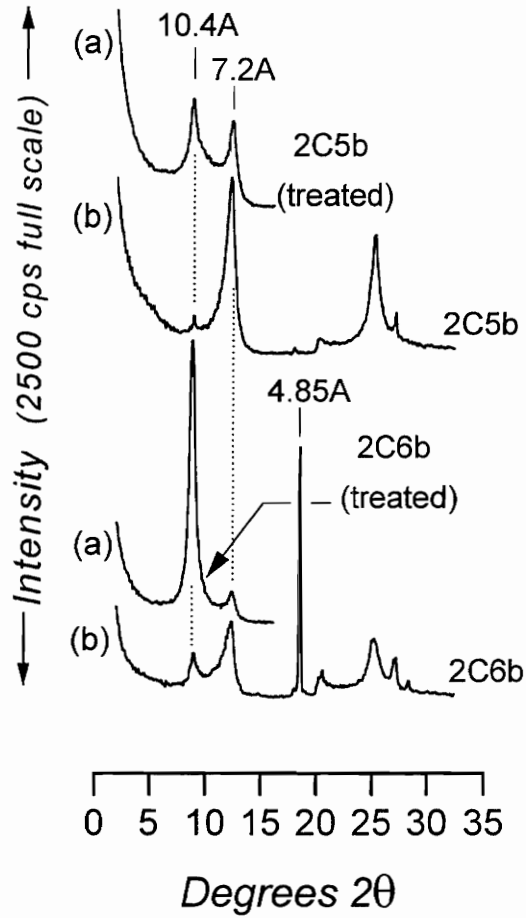


Fig. 3-6. X-ray diffractograms of clay fractions separated from the 2C5b (~400 cm depth) and 2C6b (~500 cm) saprolite horizons of pedon VA 1. Comparison of formamide-treated samples (a) with untreated ones (b) shows the dominance of halloysite deeper in the 2C6b horizon (expansion from 7.2Å to 10.4Å) and the change to a kaolinite-dominated clay fraction higher in the weathering profile.

Discussion

Periglacial Conditions at 13.8 ka BP

There is considerable geomorphological and palynological evidence for the existence of periglacial conditions in the Appalachian Highlands as far south as Georgia and South Carolina during cold periods of the Quaternary (Clark and Ciolkosz, 1988; Pewé, 1983), and there is a growing consensus that most of the widespread colluvial deposits in the Appalachians are the result of such periglacial activity (Braun, 1989).

A glacial climatic and geomorphic regime was dominant in the mid-Atlantic area from the late Wisconsin full glacial interval (23.0 to 16.5 ka BP) until at least 12.5 ka BP. Open tundra conditions and relict permafrost forms such as ice-wedge casts and pingos have been described in this area within 200 km of the glacial border (Marsh, 1987; Newell et. al., 1988). Alpine tundra has been documented during the late glacial interval (16.5 to 12.5 ka BP), for above elevations above 1000 m as far south as Cranberry Glades, West Virginia, more than 300 km from the glacial limit (Watts, 1983). Discontinuous permafrost has also been reported in Blue Ridge Province as far south as North Carolina and Tennessee (Delcourt and Delcourt, 1986). Periglacial areas adjacent to the southernmost limit of glacial advance in east-central Pennsylvania persisted until 12.5 ka BP, when tundra grasses and arctic herbs were replaced by open boreal woodland (Watts, 1983; Delcourt and Delcourt, 1986).

A cool, boreal forest dominated by spruce was reported during the late glacial at Criders Pond, Pennsylvania, ~150 km south of the ice advance, and ~150 km north of the present study site (Watts, 1983). Spruce vegetation that persisted as late as ~9 ka BP was also documented on the central Delmarva Peninsula of southern Delaware and Maryland at a similar latitude and distance from the ice as our northern Virginia Piedmont site (Sirkin et. al., 1977). This late persistence of boreal conditions in northern Virginia indicates that the time of loess deposition (13.8 ± 1.0 ka BP) considerably pre-dated climatic amelioration and the passage from a colluvial to a fluvial geomorphic regime in the mid-Atlantic area during the early Holocene. Under such a scenario, ample time was available for frost-driven slope processes to produce the matrix of loess reworked with colluvium that mantles the majority of upland soils in the study area.

Loess Deposition in the mid-Atlantic Area

An eolian origin for the silt cap we observed is corroborated by i) its being well-sorted, unstratified, and free of pebble stringers characteristic of sub-aqueous deposits (Pye, 1987), ii) a TL date of 13.8 ± 1.0 ka BP, which places it at a period of periglaciation (Péwe, 1983), iii) a degree of pedological development similar to known loess deposits on Maryland's Piedmont and Eastern Shore (Foss et al., 1978; Darmody and Foss, 1982), iv) high content of weatherable minerals which contrast with the mineralogy of the underlying material, v) restriction to level, high elevation sites with gentle slopes, and vi) its regional distribution across varying types of underlying regolith.

Conditions necessary for eolian transport include consistently strong winds with little or no rain, and the existence of bare or sparsely vegetated ground (Catt, 1988). The exposed, unstable geomorphic surfaces in a 50-150 km wide tundra directly adjacent to the ice border (Braun, 1989) no doubt provided a substantial source area for silt-size sediments. Indeed, at a distance of ~275 km from the glacial border, the northern Virginia loess is the likely product of glacial grinding and frost weathering, which are the only weathering mechanisms capable of producing appreciable quantities of mechanically abraded, silt-size (<50 μ) quartz particles (Smalley and Smalley, 1983).

The probable origin of much of the northern Virginia loess is the Susquehanna River basin, which traversed large outwash plains in central Pennsylvania and southern New York. The Potomac River basin, which drains the present study area, did not extend into glaciated terrain and could only be assumed to contribute nonglacial alluvium to the sediment source. Silt particles in the size range (5-50 μ) we observed in northern Virginia have previously been reported to come from a distance as great as 50-200 km away from the site of deposition (Pewé, 1984), validating the Susquehanna as a plausible source area. Vegetation reconstructions for the late Pleistocene also indicate that short-distance transport of loess was facilitated by drier conditions that were prevalent on the leeward side of the Blue Ridge Mountains in the mid-Atlantic area during the late Pleistocene (Sirkin et al., 1977; Delcourt and Delcourt, 1986; Wright, 1987).

The removal of much of the suspension load from the atmosphere, and the ultimate termination of substantial eolian transport in the area probably resulted from increasing

precipitation and cloud activity during late glacial summer growing seasons, more humid conditions associated with the establishment of closed boreal forests (Delcourt and Delcourt, 1984), and the abatement of wind frequency and strength as the ice sheet retreated northward beyond the Great Lakes at ~12 ka BP (Jacobson, et al., 1987). While the establishment of closed boreal forests may have been associated with the end of eolian activity during the late glacial, the record of post-depositional colluvial deposits and reworked loess on the landscape suggests that periglacial conditions persisted into the early Holocene.

Post-Depositional Colluviation

The lack of mineralogical differences attributable to pedogenesis within loessial horizons, and the sharp morphological, mineralogical, and chemical boundary between the intact loess cap and the underlying stoneline and paleosol at pedon VA 1 implies that the loess was deposited within a short time interval and that no appreciable basal mixing zone has formed between these two lithologies. Soils with preserved silty mantles in the study area are of very limited extent, however, commonly occupying only the broadest and most stable ridgetops. Where stone lines are absent or discontinuous (generally on non-stable or nonlevel summit positions), colluvial mixing of the loess with the underlying residuum has resulted in a surface mantle that is morphologically and lithologically similar to, and almost indistinguishable from, a soil formed entirely in place.

Pedon VA 2 typifies those soils that occupy summit positions and have mixed zones of colluvium with reworked loess incorporated into existing surface horizons. The surface horizon of pedon VA 2, for example, has a particle size distribution and a silt component nearly identical to the surface of pedon VA 1's uneroded loess cap (Fig. 3-1, 3-7). A weakly-defined stone line in VA 2 also identifies a buried subsoil horizon with a particle size distribution analagous to the buried paleoargillic horizon in VA1(Fig. 3-7, 3-8). Pedon VA 2 also has remnants of the dark, illuvial clay flows and organic coatings on prism faces similar to those found in the VA 1 paleosol. Colors of the colluvial material are also centered on yellowish browns, whereas they shift to yellowish reds in the residual subhorizons, indicating higher hematite contents. Finally, the distribution of exchangeable Mg and CEC is also similar in both pedons VA 1 and VA 2 above and below the zone of loess influence (Fig 3-8).

These lines of evidence suggest that the buried subsoil horizons of pedons VA 1 and VA 2 formed in a similar pedogenic environment and represent stratigraphic markers of a former erosional surface. Accordingly, much of the loess deposited in this region during the late glacial was not entirely lost through erosion, but remains on the landscape and is incorporated into the existing soil, probably through bioturbation, pedoturbation, and periglacially-driven colluvial processes, which persisted at least until boreal species were displaced by deciduous forests in the early to middle Holocene.

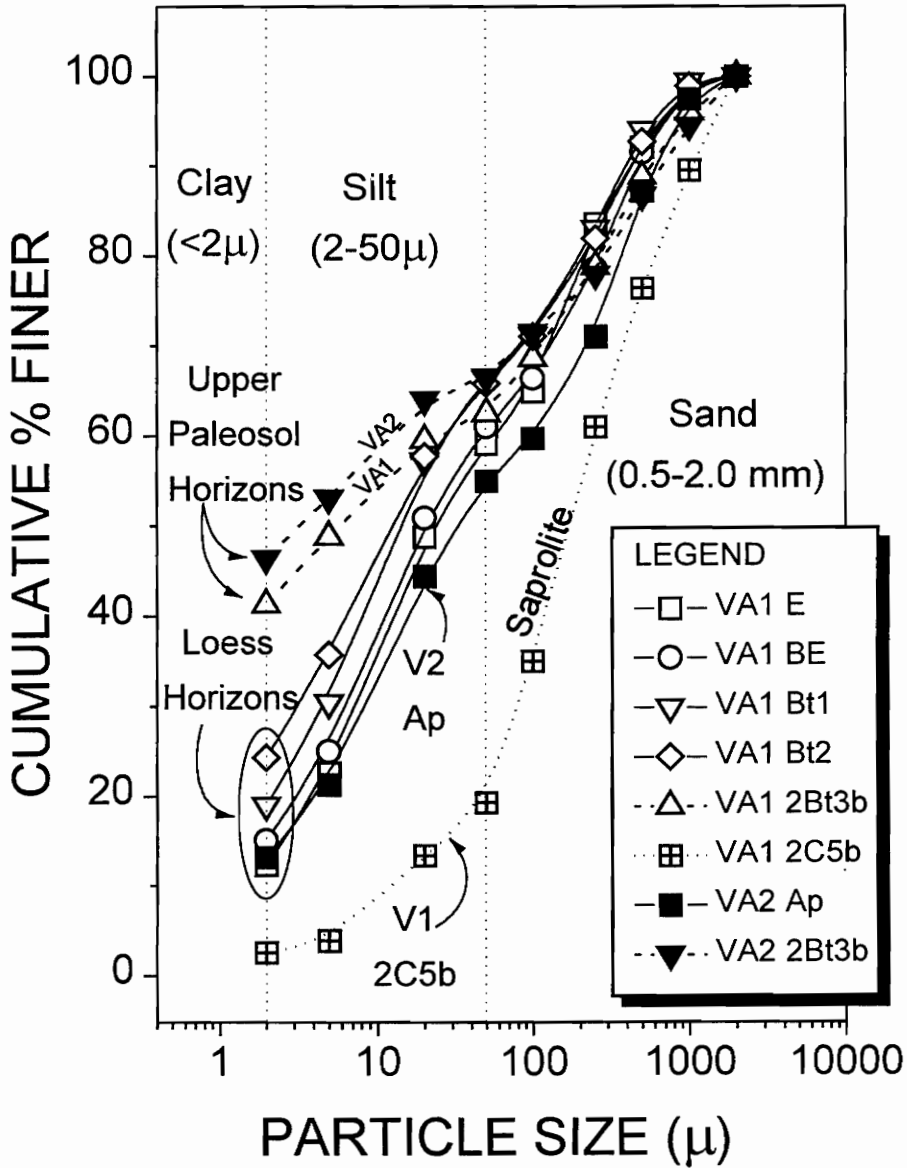


Fig. 3-7. Cumulative grain size frequency curves for the VA 1 loess cap, upper paleosol horizons, and an unoxidized saprolite horizon. Grain size distribution for the VA 2 surface horizon is also shown for comparison.

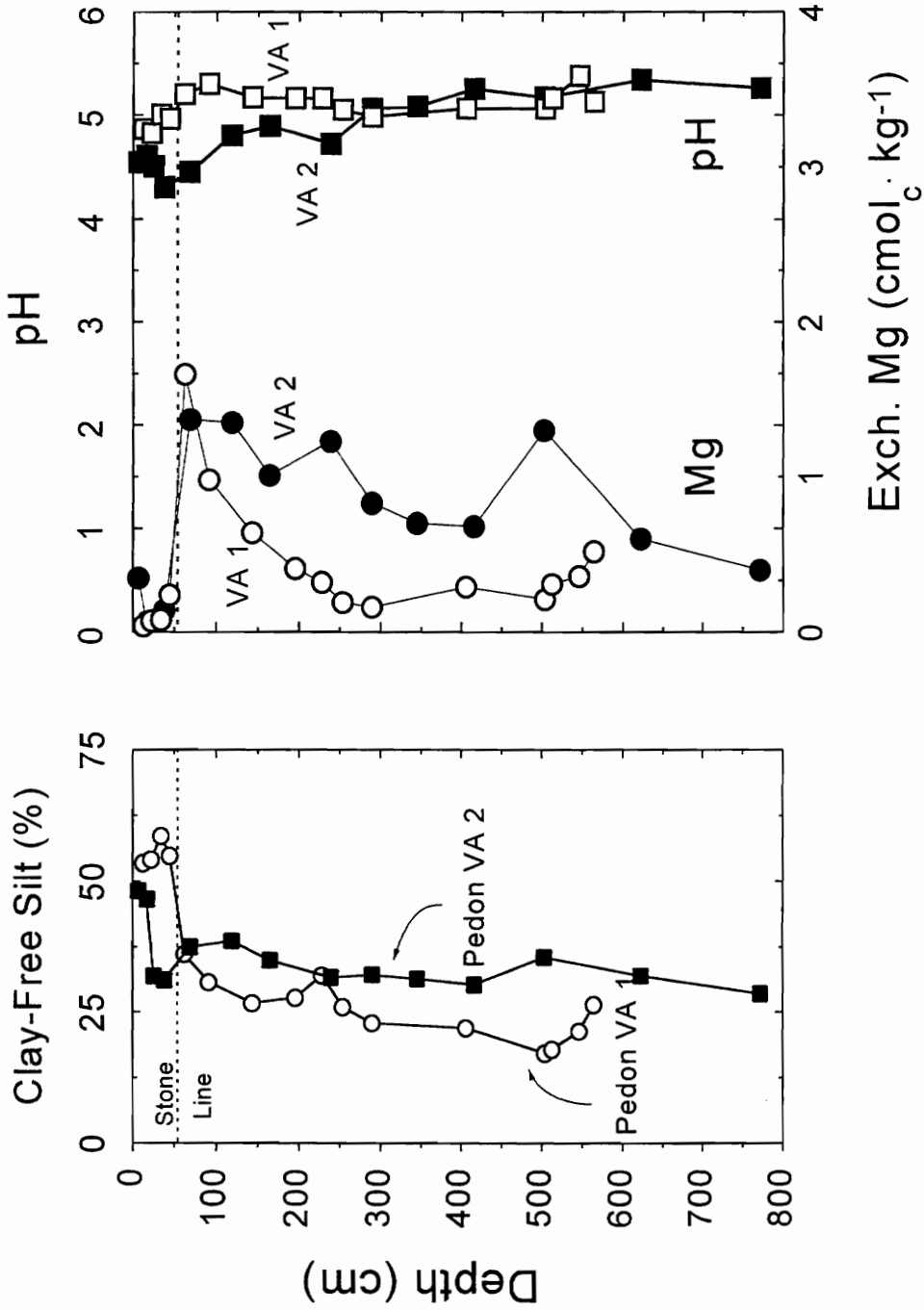


Fig. 3-8. Clay-free silt content, pH, and exchangeable Mg of pedons VA 1 and VA 2.

Summary and Conclusions

A truncated paleosol overlain by a dated loess deposit records a period of highly fluctuating climatic conditions south of the glacial border near the close of the Pleistocene. The continuous stone line in one soil we studied marks a period of aggravated sheet erosion and winnowing of fine (<2.0 mm) material from a highly developed residual soil containing resistant vein quartz. Although the timing of this event is unknown, we consider the formation of this stone line, which requires a high rainfall environment and exposed slopes, to pre-date the late glacial age of the loess deposit, which, in contrast, requires extreme aridity. Following loess deposition, a periglacial environment conducive to accelerated sediment transport and mass-wasting caused reworking of materials on the landscape, resulting in the admixture of colluvium and loess that mantles the majority of area soils today.

Evidence suggests that loess deposits in the mid-Atlantic area, possibly extending to south-central Virginia, were more extensive than previously believed. Additionally, the widespread occurrence of stone lines and colluvial caps on stable uplands as far south as Georgia also suggests that climatic oscillations highly influenced geomorphic processes throughout much of the Piedmont Province during the Pleistocene.

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Chapter IV

Mineralogy of a Paleoargillic Horizon and Its Contemporary Analogue on the Northern Virginia Piedmont

ABSTRACT

One dilemma commonly encountered in using soils for paleoenvironmental reconstruction is the inherent difficulty of separating relict pedogenic characteristics from those that have been imposed by weathering in the modern environment. We studied a loess-mantled paleosol in order to identify and elucidate complex polygenetic weathering mechanisms that contrast sharply between parent materials. In this paper, we describe the processes and products of secondary mineral neoformation and transformation in a dated loess deposit and underlying paleosol derived from granitic saprolite on the northern Virginia Piedmont. Additionally, we assess the magnitude of pedogenic (argillic horizon) development in a deposit of known age, and the influence of post-burial pedogenesis on the mineralogical composition of the underlying paleoargillic horizon.

Three distinct zones of weathering in this profile are described based on the distribution of minerals in both the clay and silt fractions. Surficial loess horizons in Zone 1 have a weakly-developed argillic horizon that formed in a humid-temperate

climate within the last ~13.8 ka. Clay formation has resulted at the expense of weathering of medium and fine silt minerals, but alteration of the coarse silt fraction has been only minimal within this time period.

Upper horizons of the paleosol (Zone 2) show a mineral suite characteristic of an advanced stage of weathering, suggesting that a relatively stable geomorphic surface that is probably pre-Wisconsinan. This zone exhibits evidence of strong desilication, depletion of feldspars, and a relict high-charge smectite phase that is inconsistent with long-term weathering under the modern conditions of intense, acid leaching.

Zone 3, beginning at a depth of ~500 cm, marks the transition from unoxidized saprolite to oxidized saprolite higher in the profile. This zone is characterized by a dramatic increase in plagioclase feldspars, and both gibbsite and halloysite. Halloysite is shown to form pseudomorphically from both biotites and albites in the silt fraction, and to recrystallize to form kaolinite higher in the profile. Gibbsite is resilicated to form kaolinite at a depth of ~400 cm in the profile. In addition to these processes, four mechanisms of kaolinite formation, and two processes of gibbsite formation are identified.

Introduction

Soils and paleosols have been used to estimate the age and origin of surficial deposits and geomorphic surfaces, assess landscape evolution and stability, and infer paleoenvironmental conditions during pedogenesis (Birkeland, 1990). One dilemma

commonly encountered in using soils for paleoenvironmental reconstruction is the inherent difficulty of separating relict pedogenic characteristics from those that have been imposed by weathering in the modern environment (Ruhe and Olson, 1980). This problem is exacerbated by a general lack of age control for deposits >40,000 yr (Mills and Delcourt, 1991), considerable uncertainty regarding rates of geochemical weathering under field conditions (Velbel, 1986), and consequently, our inability to quantitatively segregate such factors as the complex effects of climate and time on weathering intensity, soil profile development, mineral stability and transformation, and elemental flux within the pedoenvironment (Boardman, 1985; Feldman et al., 1995).

Although an understanding of the quantitative relationships between climate, time, and weathering processes is requisite to the correct interpretation of the continental geologic record with regard to paleoclimates, there is currently a scarcity of quantitative data regarding the specific effects of both climate and time on weathering processes, particularly in the nonglaciaded southeastern U.S. Without this information, we can neither identify the degree to which soil properties and secondary phases persist, alter, or become obliterated in the weathering environment as conditions change, nor can we predict with any certainty which soil features are reliable as indicators of retrospective or prospective climate/environmental change.

An excellent opportunity to study pedogenic relationships in soils formed both during and before the present interglacial period exists in an area of the northern Piedmont, where bisqual soils have formed on upland interfluves in about 0.8-1.0 m of Holocene

loess overlying a buried paleosol derived from granitic saprolite. In a previous paper (Feldman et al 1995, in review), we reported a thermoluminescence date of 13.8 ± 1.0 ka BP for this previously unidentified loess cap, and we provided evidence for the post-depositional reworking of this material throughout the landscape by periglacial processes during the late glacial and early Holocene. In this paper, we expand on our previous work by describing further i) the processes and products of secondary mineral neof ormation and transformation in both the loess and the residual soil/saprolite profile, ii) the magnitude of pedogenic development and argillic horizon formation in a deposit of known age, and iii) the influence of post-burial pedogenesis on the mineralogical composition of the underlying paleoargillic horizon. Finally, using a geochemical mass-balance approach, we evaluate the nature of volume change associated with soil and saprolite formation from crystalline rocks on this part of the Piedmont.

Materials and Methods

Location, Geologic Setting, and Bedrock Geology

The study area lies in the outer Piedmont Crystalline Province (Thornbury, 1965) in Prince William County, near Woodbridge, VA (Fig. 1-1). The area is underlain by the Occoquan Batholith, which consists of a medium-grained, light gray monzogranite with moderate to strong metamorphic foliation and a well-developed mineral lineation (Seiders et al., 1975). Major minerals include quartz, plagioclase (commonly with inclusions of sericite), and microcline with subordinate muscovite, biotite, and epidote (Table 4-1).

Plagioclase varies between albite (Ab₉₄₋₉₅) and a mixture of albite and oligoclase (Pavich, 1986). Microcline is generally free of inclusions and is commonly untwinned.

Table 4-1. Mean range of mineralogical composition of the Occoquon Granite.^{†‡}

Quartz	K-Feldspar	Plagioclase	Biotite	Muscovite	Epidote
29-42	24-30	17-22	1-4	8-15	2-5

[†] After Seiders et al (1975).

[‡] Includes the following accessory minerals: chlorite, sphene, apatite, garnet, opaques, allanite, pyrite, carbonates, and zircon.

The sampling site is located at the highest point of the local landscape, on a convex interfluvium at an elevation of 97 m. Major topographic and drainage features in the area are primarily controlled by the structure and foliation of the underlying lithology. Saprolite is thickest on upland divides and thinner on short, steep sideslopes. In the valley bottoms, third- and fourth-order streams are commonly incised into bedrock (Pavich, 1989).

Present climate of the Piedmont is humid-temperate to subtropical (Thornbury, 1965). Mean annual temperature (MAT) in the area is 12.2°C and mean annual precipitation (MAP) is 1024 mm yr⁻¹ that is uniformly distributed throughout the year (U.S. Dept. of Commerce, 1987) (Fig. 1-2). Calculated water balance data for the study area, based on MAT and calculated potential evapotranspiration (PET) values, show that soils currently undergo a summer moisture deficit and have only minimal capacity for deep percolation below the solum (Table 1-1).

Sampling and Laboratory Analyses

The soil we studied was described and sampled by genetic horizon at a representative location along a 2 m-deep vertical, recently-excavated cut-bank with an exposure of ~300 m along contour. Deeper samples (>5.5 m) were obtained at ~50 cm intervals using a modified bucket auger method (Stolt et al., 1991), which permitted bulk sampling and also sampling for bulk density determinations. Polished thin sections, prepared from oriented clods taken from the loess horizons of pedon VA1, were examined with a petrographic microscope.

Soil samples were air-dried, sieved to remove coarse fragments (>2 mm), and thoroughly mixed. Standard physical and chemical characterization included: i) particle-size analysis by the pipette method, ii) organic carbon by wet oxidation with dichromate and back-titration with 0.5 M $\text{Fe}(\text{NH}_4)_2(\text{SO}_4)_2$, iii) cation exchange capacity (CEC) by the sum of neutral M NH_4OAc -extractable bases (pH 7.0) plus total soil acidity by BaCl_2 -TEA at pH 8.2, and iv) free Fe by Na-dithionite-citrate-bicarbonate (DCB) analysis (Soil Survey Staff, 1984). Soil pH was measured in the supernatant portion of a 1:1 soil/deionized water suspension after a 1 h equilibration period.

Samples were prepared for mineralogical analysis by pretreating with 30% H_2O_2 , adjusted to pH 5 with M NaOAc to remove organic matter where necessary (Kunze, 1965), and with DCB to remove Fe-oxide coatings (Holmgren, 1967). Samples were then adjusted to pH 9.5 with M Na_2CO_3 to effect particle dispersion, and sand (>50 μ) was

separated by wet sieving. The remaining clay (<2 μ) and silt (2-50 μ) fractions were separated by repeated decantation after centrifugation.

Analysis of elemental Zr in the coarse silt fraction was performed using Spectrasol™ dissolution reagent for silicates (Spectrasol Inc., McAfee, NJ, 07428). Analysis by this method involves HF dissolution under pressure followed by the addition of a reagent which neutralizes the HF and stabilizes the analytes in solution without adding salts to the sample. This allows for the use of a glass concentric nebulizer and an all-glass ICPES transport system.

X-ray diffractograms of clay fractions were obtained from i) Mg-saturated, glycerol-solvated aliquots with no heat treatment (25°C) and after 4 hours of heating at 110°C, and ii) K-saturated samples with no heat treatment, and after 4 hours heating at 110°, 300°, and 550°C, respectively. Samples were scanned with CuK α radiation with fixed stepping parameters of 4.0s at 0.075° step⁻¹. Selected clay samples were also treated with formamide to differentiate kaolinite from halloysite by intercalation (Churchman et al., 1984). Kaolinite and gibbsite were quantified by mass-equivalent calibration of differential scanning calorimetry (DSC) endothermic peak areas using poorly crystalline Georgia kaolinite (KGA-2)¹ and Reynolds synthetic gibbsite (RH-31F)² as standards.

Selected samples of the deferrated very fine sand (0.05-0.10 mm) fraction were investigated by scanning electron microscopy (SEM) using a Camscan™ Series 2 SEM operated with a fixed working distance of 11 mm and an operating voltage of 30 kV.

¹ Source Clay Minerals Repository, Dept. of Geology, University of Missouri, Columbia, Mo 65211

² Reynolds Metals Co., Alumina Research Division, Bauxite, AR 72011

Randomly-oriented bulk samples were mounted onto aluminum stubs and coated with a ~20Å layer of Au-Pb in a vacuum evaporator. The system was also equipped with a HNU™ energy dispersive spectrometer, which was operated at a working distance of 27 mm and a voltage of 15 kV.

Results and Discussion

Physical and Chemical Characteristics

Abrupt transitions in chemical and physical properties between the loess and underlying paleosol point out the contrasting nature of these parent materials and indicates that mixing between them has been minimal. Clay-free silt content of all horizons in the loess cap exceeds 50% (Fig 4-1a). The loess is largely comprised of both medium (5-20µ) and coarse (20-50µ) silt particles that show evidence of mechanical abrasion, including subangular morphology, concoidal fracturing, and breakage features on grain surfaces (Feldman et al., 1995).

Quartz grains in the silt fraction of the paleoargillic horizon, in contrast, have irregular edges and dissolution features characteristic of *in situ* chemical weathering. Clay-free silt content decreases to a constant ~25% below the stone line, indicating that the paleosol formed entirely from the underlying saprolite.

Clay content, pedogenic Fe-oxides (Fed), CEC, and exchangeable Mg all have maxima occurring immediately below the stone line and lower limits of accumulation

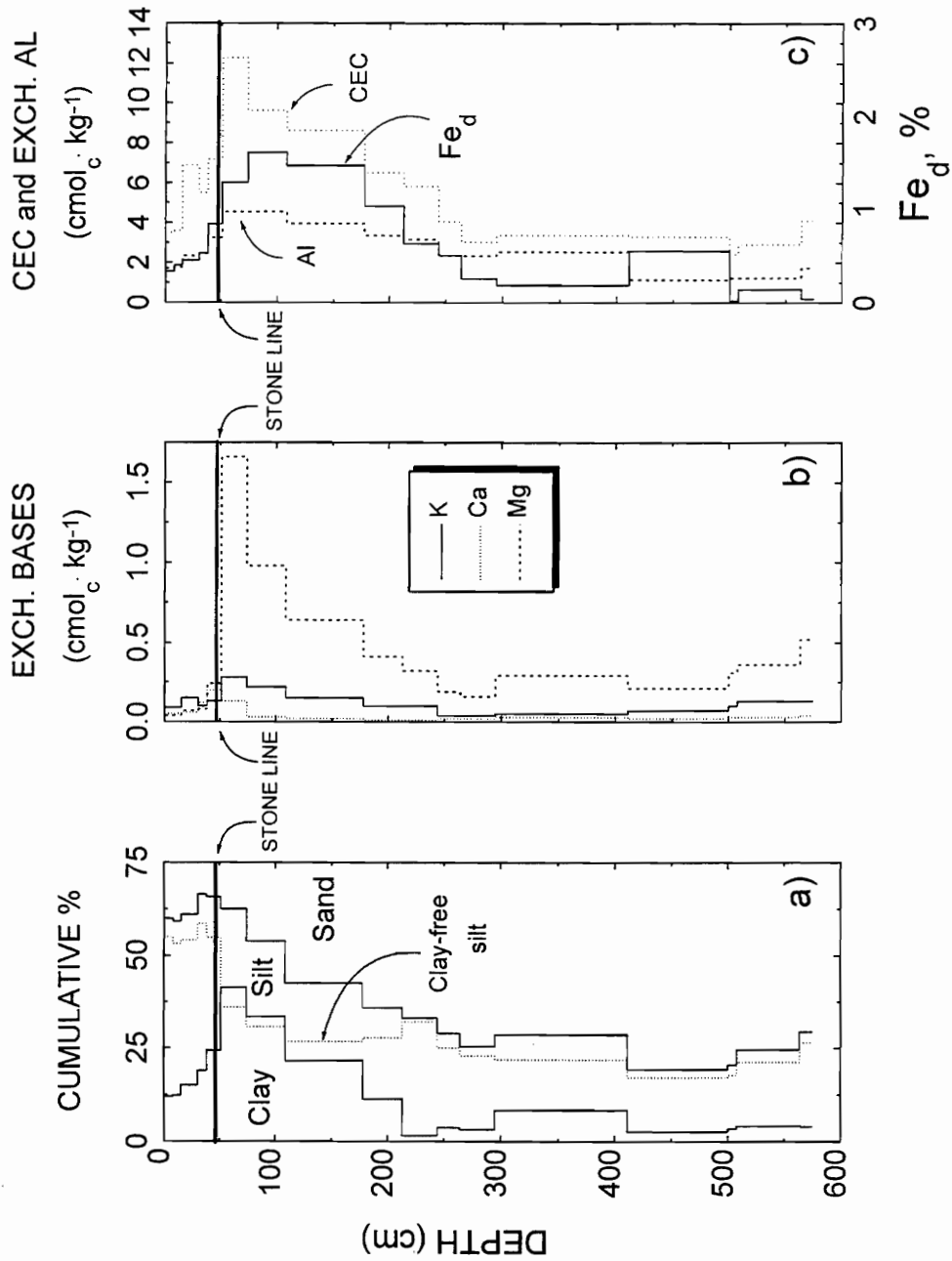


Fig. 4-1. Particle-size distribution, exchangeable cations, cation exchange capacity (CEC), and free Fe-oxide (Fe_d) trends showing dramatic changes below the stone line in the underlying paleosol.

between ~200-250 cm (Fig. 4-1a-c). This 200-250 cm depth essentially defines the major zone of pedogenic activity in the profile and is associated with the transformation of saprolite to soil. Because the granitic parent material in which this soil formed is free from inherited secondary clay minerals or oxides, this zone also effectively defines the depth of the wetting front under which both pedogenic clay and Fe_d have accumulated in the paleoargillic horizon.

Exchangeable Mg increases abruptly to an unusually high level in the paleosol, given the present high leaching conditions and low levels of exchangeable Ca and K shown (Fig. 4-1b). High levels of Mg may denote previous conditions of greater annual cycling of bases and biomass production under what might have been grassland vegetation (Birkeland, 1984).

A near doubling of the clay content within the loess cap, development of strong brown matrix colors, and evidence of argillans in Bt1 and Bt2 horizon thin section demonstrates that a weak argillic horizon has formed in the loess within the 13.8 ka since deposition. Maximum clay content of the paleosol, however, is nearly twice that of the average clay content of the loess cap, indicating that profile development has advanced to a much greater degree prior to loess deposition. This level of increased pedogenic activity in the paleosol is also reflected in the higher CEC values ($4-12 \text{ cmol}_c \cdot \text{kg}^{-1}$) and the dominance of high-charged 2:1 (constant charge) phyllosilicates in the paleoargillic horizons relative to those of both the saprolite horizons ($CEC \sim 3 \text{ cmol}_c \cdot \text{kg}^{-1}$) and the

loess cap (CEC 3-6 $\text{cmol}_c\cdot\text{kg}^{-1}$), which are dominated by 1:1 (variable charge) phyllosilicates.

Mineralogy

Three distinct zones of weathering in the profile can be inferred from the distribution of clay minerals shown in Fig. 4-2. The uppermost zone (Zone 1), above the stone line in loessial horizons, shows the least amount of weathering of all horizons in the profile. There is little or no evidence of mineral alteration within coarse silt (20-50 μ) fraction of the loess cap during the last ~10 ka of weathering in a humid-temperate climate (Fig. 4-3). The original primary weatherable minerals (i.e., muscovite and feldspars) present in the fine and medium silt fractions, however, have altered sufficiently to form a clay mineral suite dominated by kaolinite (7.2Å and 3.57Å) > hydroxy-interlayered vermiculite (HIV) (14.6Å) > quartz (4.24Å and 3.34Å) > mica (10.0Å) > regularly-interstratified mica/vermiculite (RMV) (24.6Å and 12.3Å) (Fig. 4-2). Dominant weathering mechanisms include i) solid-state transformation of muscovite to vermiculite through an interstratified phase, and hydroxy-Al interlayering in vermiculite to form HIV in surficial horizons, and ii) both hydrolysis of K-feldspars and desilication of clay-size 2:1 phyllosilicates to form kaolinite in surface and subsurface horizons. With the exception of an increase in kaolinite from 26% in the Ap to 43% in the Bt2 horizon, variation in clay mineral abundance is almost invariant with depth in the loess cap,

Pedon VA1 Clay Fraction

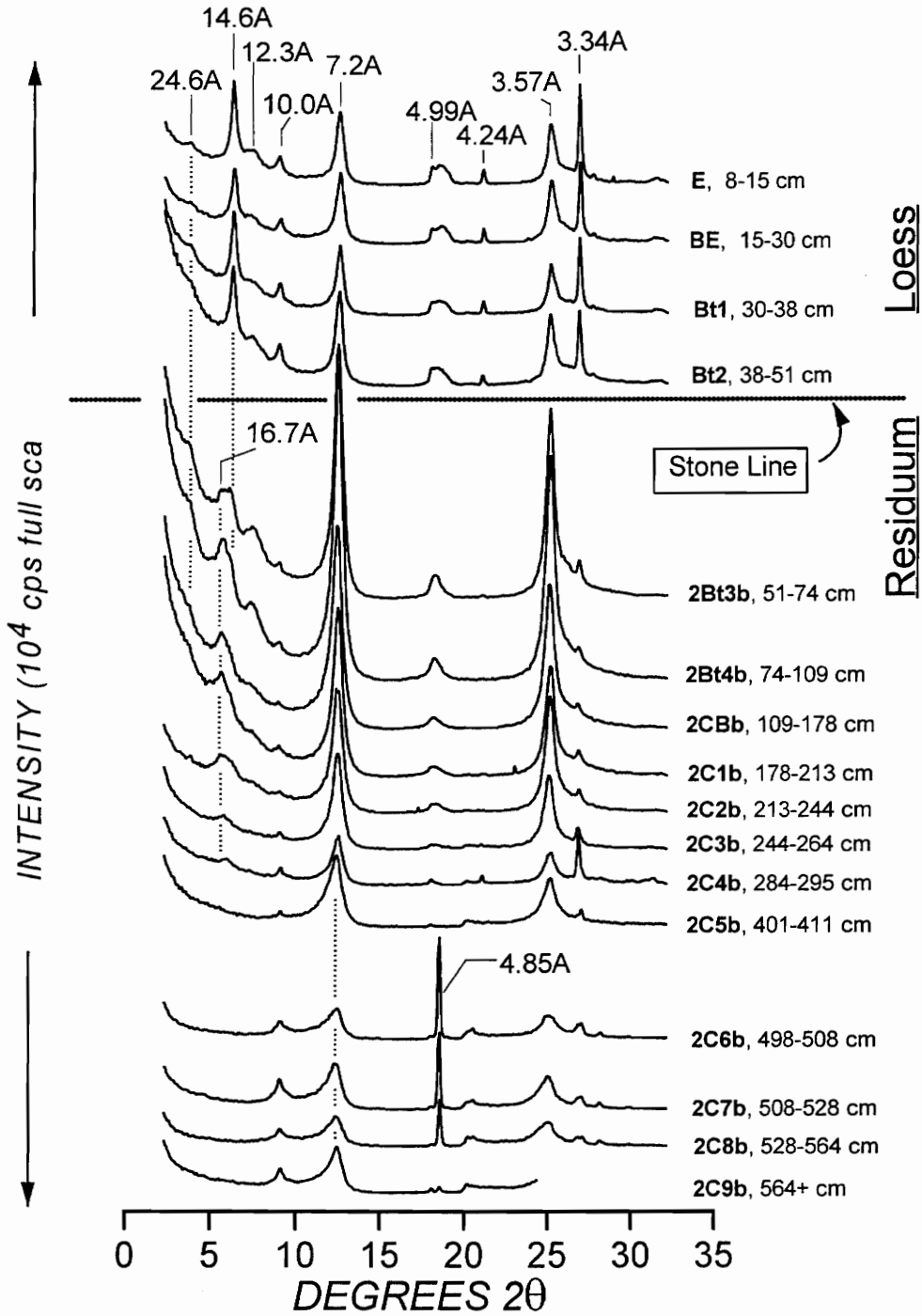


Fig. 4-2. X-ray diffractograms of oriented clay fractions, shown by horizon. Samples were Mg-saturated, glycerol-solvated, and scanned at 25°C.

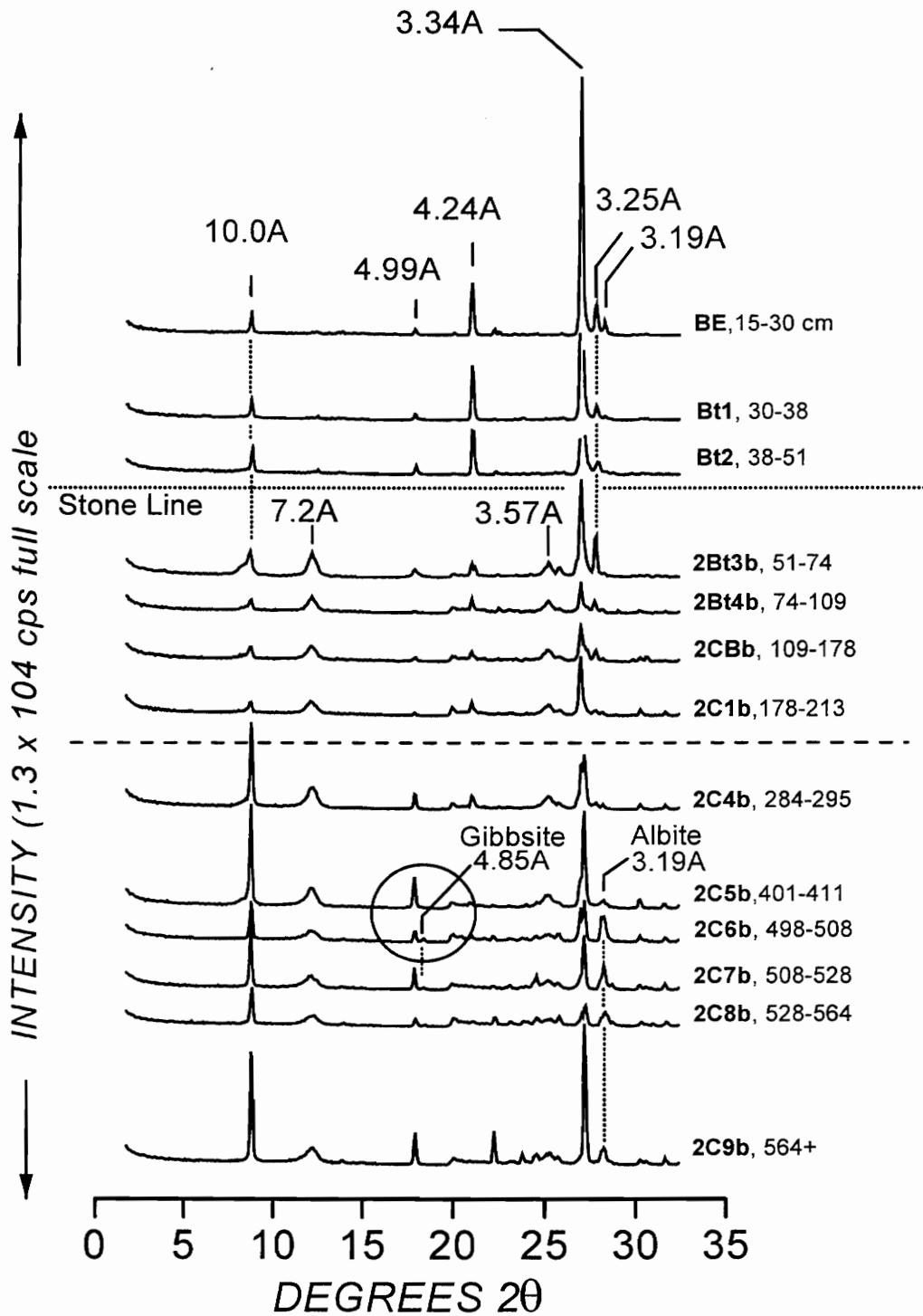


Fig. 4-3. X-ray diffractograms of the semi-oriented powder mounts of the coarse silt fractions.

further suggesting that only minimal pedogenic development and profile differentiation has taken place since the loess was deposited in late glacial time.

Below the stone line, the upper horizons of the paleosol mark the top of Zone 2, where the relatively advanced stage of weathering documents a geomorphic surface that was stable for a considerably longer than the late glacial loess of Zone 1. Kaolinite (coexisting with halloysite lower in the profile) is the dominant secondary phase in the paleosol, ranging from 70-90% of the clay fraction, with lesser amounts of HIV, mica, and only a trace amount of quartz (Fig. 4-2). Plagioclase feldspars are entirely depleted in these horizons, and the small amount of RMV and K-feldspars that are present in the 2Bb horizons are the probable result of silt infilling from the overlying loess through coarse prisms and desiccation cracks during dry seasons. A trace amount of gibbsite is also present in the 2Bt3b clay fraction, attesting to long periods of stability and desilication.

A distinct high-charge smectite phase, the product of muscovite weathering lower in the saprolite, is present in the paleosol and not the overlying loess cap. This mineral is represented by a shift from the glycolated 14.6\AA (HIV) peak above the stone line, to the expanded 16.7\AA (smectite) peak in the paleosol (Fig. 4-2). Smectite is associated with the high levels of exchangeable Mg and is responsible for the high CEC in the upper paleoargillic horizons (Fig. 4-1). The presence of this mineral is inconsistent with long-term weathering under the modern conditions of intense, acid leaching (Karathanasis et al., 1983; Karathanasis and Hajek, 1984) and may reflect drier climatic conditions at or before the time of loess emplacement. Its preservation in the existing weathering

environment may be an effect of i) inhibition of deep leaching through the paleosol owing to preferential water retention by the loess, and ii) slower chemical weathering rates resulting from lower temperatures and drier conditions during the late Wisconsinan (Feldman et al., 1995).

The lowermost weathering zone (Zone 3) of this profile, beginning at a depth of ~500 cm, marks the transition from unoxidized saprolite to oxidized saprolite higher in the profile. This zone is characterized by a distinct decrease and broadening of the kaolinite (7.2Å) peak, the absence of any expansible (weathered) 2:1 phyllosilicates, the appearance of gibbsite (4.85Å) in the clay fraction (Fig. 4-2), and an abundance of Na-plagioclases in the silt fraction. The first emergence of gibbsite in the 2C5b horizon clay fraction coincides with the depletion of silt-size plagioclase (3.19Å) in the saprolite below 5 m depth (Fig. 4-3), confirming the earlier work by Alexander et al. (1941) and Calvert et al. (1980) which showed that gibbsite is primarily formed on the Piedmont at the expense of feldspar dissolution.

The broadening and decrease in height of the the 7.2Å peak beginning at a depth of ~5 m in Zone 3 is related to the formation of dehydrated halloysite as a pseudomorphic alteration product of Na-plagioclase. Halloysite is identified in Figure 4-4 as a shift in the 7.2Å x-ray diffraction peak to 10.4Å, corresponding to lattice expansion after selective intercalation with formamide (Churchman et al., 1984). In the 2C6b horizon clay fraction at a depth of ~5 m, it is the dominant component of the 7.2Å diffraction peak. One meter higher in the 2C5b horizon, there is considerably less expansion to 10.4Å, suggesting

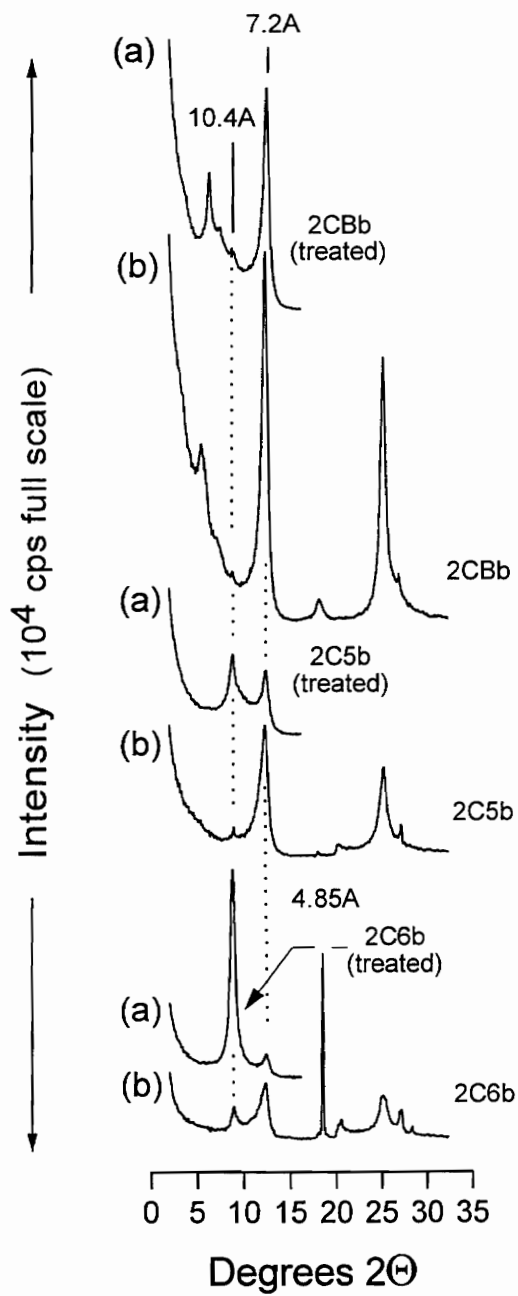


Fig. 4-4. X-ray diffractograms of (a) formamide-treated and (b) untreated clay fractions from selected horizons.

that halloysite is transformed to kaolinite between these depths. Higher still in the profile, in the lower part of the paleoargillic horizon, only trace amounts of halloysite remain (Fig. 4-4).

SEM

Halloysite originates in this profile as a secondary mineral from the pseudomorphic transformation of *both* biotite (Feldman, 1995, in preparation) and Na-bearing plagioclases. Most albite grains in the silt (2-50 μ) and fine sand (100-250 μ) fractions are almost entirely replaced by secondary halloysite pseudomorphs (Fig. 4-5). Halloysite tubes initially crystallize topotactically with respect to residual albite twin planes (Fig. 4-6a,b) and biotite planar (001) surfaces (Fig. 4-7). In contrast to the plagioclases, microcline surfaces show only deep dissolution etch pitting, with no evidence of secondary mineral formation (Fig. 8a-b). This contrasts with other studies that have reported the alteration of both plagioclases *and* potassium feldspars to kaolin minerals (Gilkes et al 1979; Tazaki and Fyfe 1987).

Well-crystalline gibbsite particles, shown as high contrast euhedral crystals of ~1-3 μ across, occur in intimate association with halloysite in both the albite and biotite (Fig. 4-5, 4-6a-b, 4-7), but the coprecipitation of these minerals is not likely. The timing of gibbsite precipitation with respect to halloysite genesis remains unclear, but XRD and SEM evidence suggests that formation of halloysite precedes that of gibbsite during the production of saprolite. Halloysite is present in the youngest (2C9b) horizon, which is

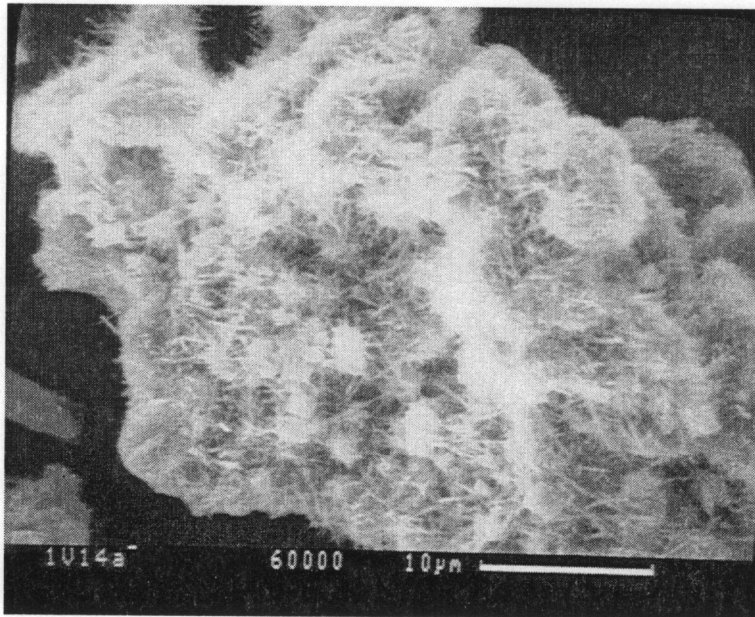


Fig. 4-5. Scanning electron micrographs of an entirely halloysitized, coarse silt-sized albite grain from the 2C6b horizon.

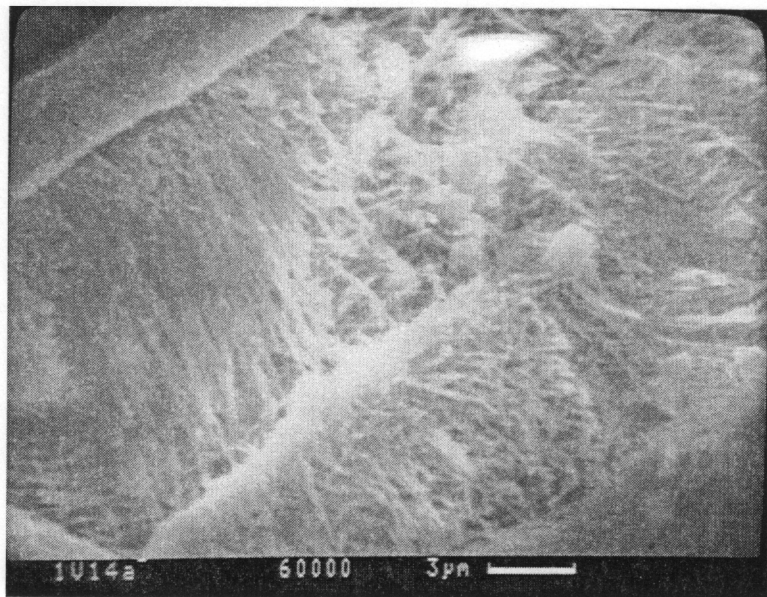
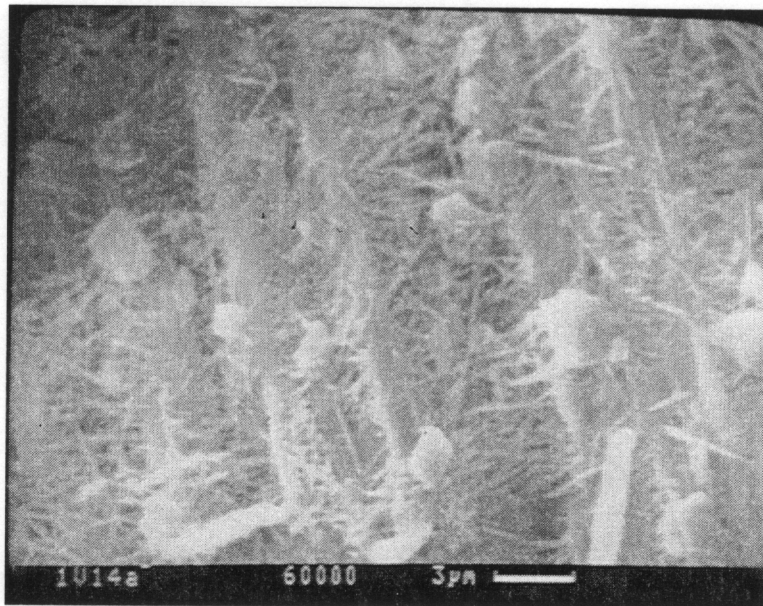


Fig. 4-6. Topotactic growth of halloysite within residual albite twin planes. Well crystalline gibbsite is shown as high-contrast, euhedral particles of $\sim 1 \mu$ across.

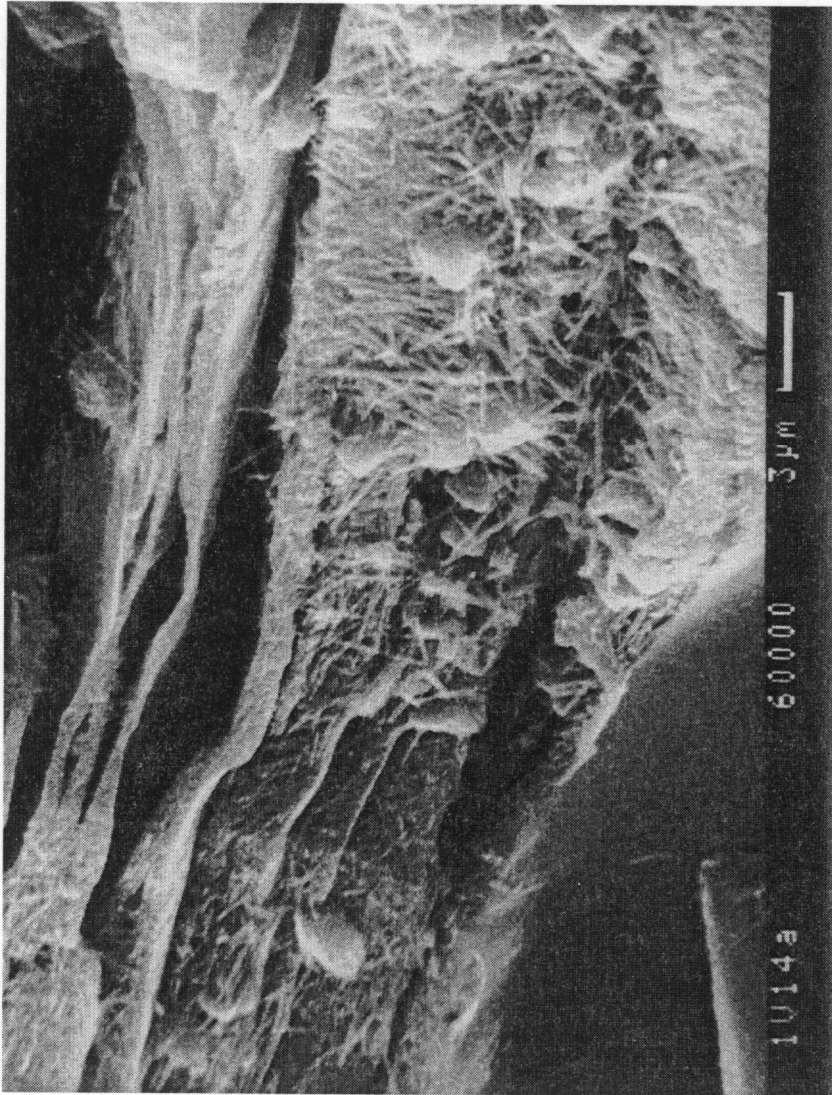


Fig. 4-7. Topotactic halloysite with respect to biotite planar (001) surfaces.

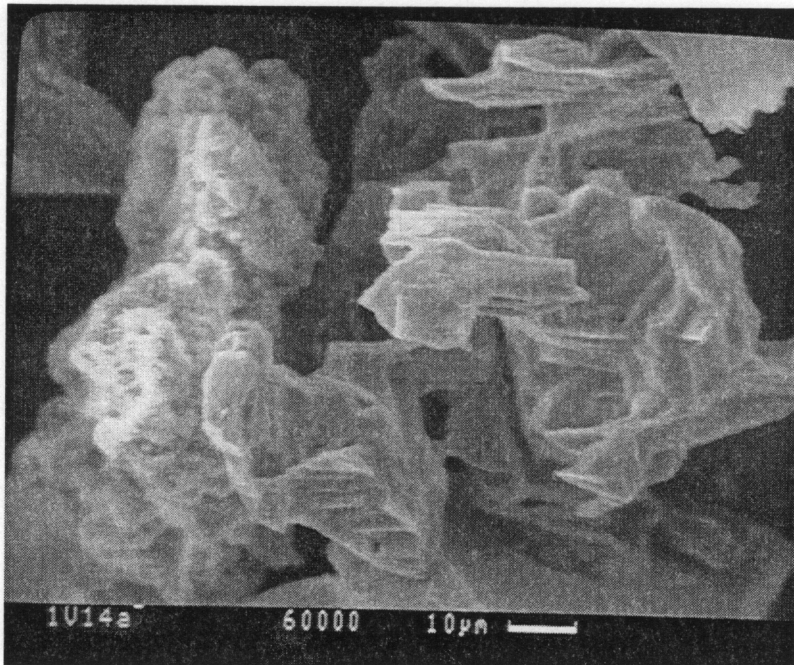
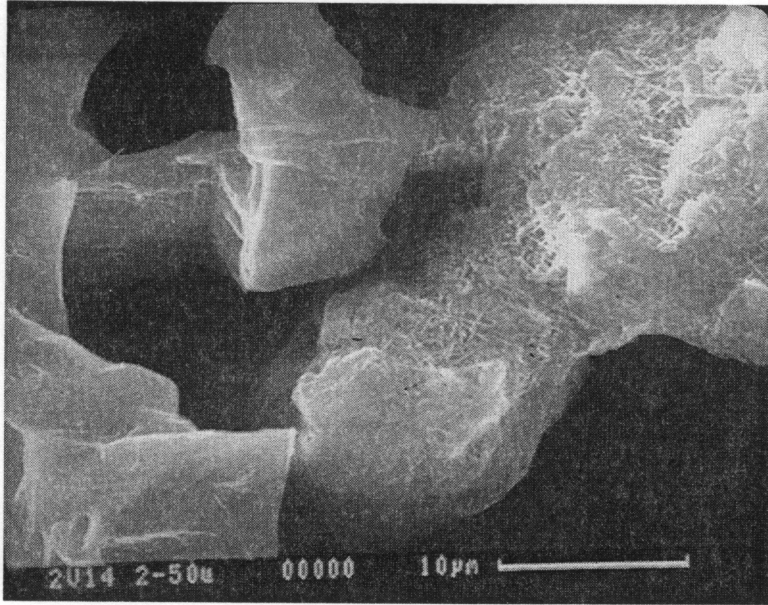


Fig. 4-8. Contrast between albite grains entirely pseudomorphed by halloysite, and microcline grains showing only deep dissolution etch pits withno residual secondary products.

the deepest, most recent saprolite horizon formed as the weathering front descends toward bedrock. Gibbsite begins to appear in the profile in the overlying and more highly weathered 2C8b horizon (Fig. 4-2). Moreover, halloysite tubes of $\sim 1 \mu$ in length commonly bisect gibbsite crystals (Fig. 4-9a-b), further suggesting that gibbsite has precipitated around an existing halloysite framework.

Micrographs of polished thin sections show that the initial alteration of plagioclase grains in these deep saprolite horizons may initially involve dissolution of the grain interior (Fig. 4-10a-b) followed by reprecipitation of secondary halloysite within the framework (Fig. 4-10b) until the host is entirely replaced by the alteration product (Fig. 4-10c).

Tubular halloysite derived from albite alteration in coarser fractions dominates the clay fraction of horizons below a depth of ~ 5 m (Fig. 4-11b). One meter higher in the 2C5b horizon, however, halloysite diminishes appreciably in abundance and kaolinite becomes the dominant clay mineral, as indicated by the profusion of pseudo-hexagonal plates in association coexisting with halloysite tubes (Fig. 4-11a). Two possible mechanisms may explain this trend. In one case, the decrease in halloysite upward in the profile and the increasing intensity of the 7.2\AA diffraction peak (Fig. 4-2) is simply the result of dissolution of halloysite and reprecipitation to the more stable kaolinite phase. Analysis of samples with formamide has shown this to be the case (Fig. 4-4). The transformation of halloysite to kaolinite occurs not only in the clay-size fractions, but also

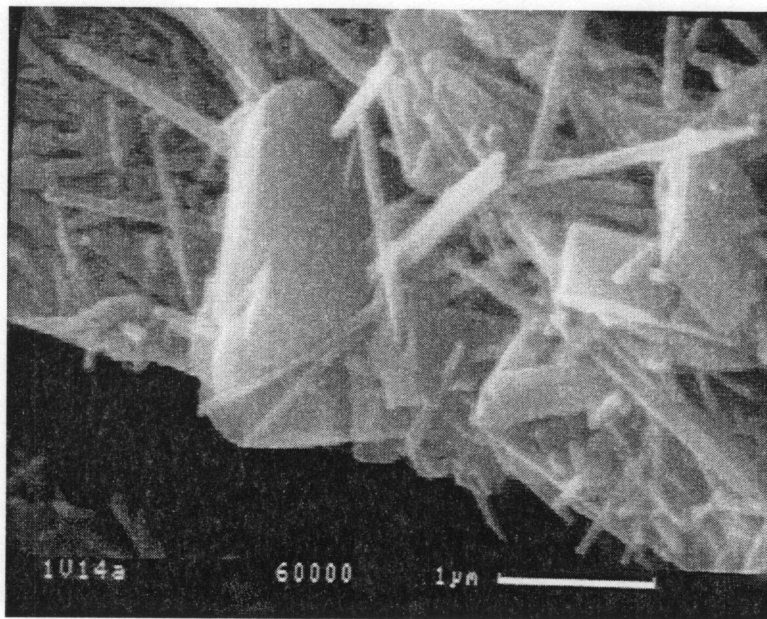
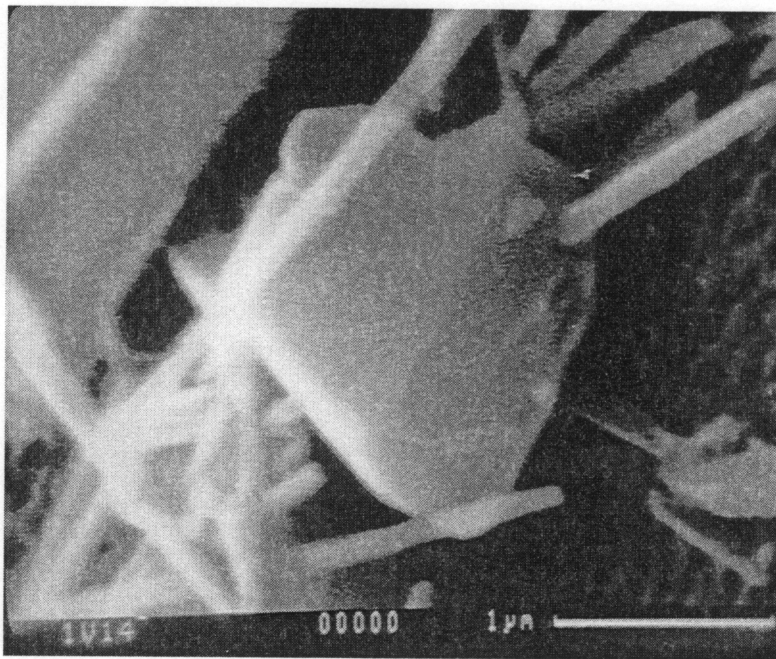


Fig. 4-9. Halloysite tubes of $\sim 1 \mu$ in length bisecting gibbsite crystals, suggesting that gibbsite has precipitated around an existing halloysite framework.

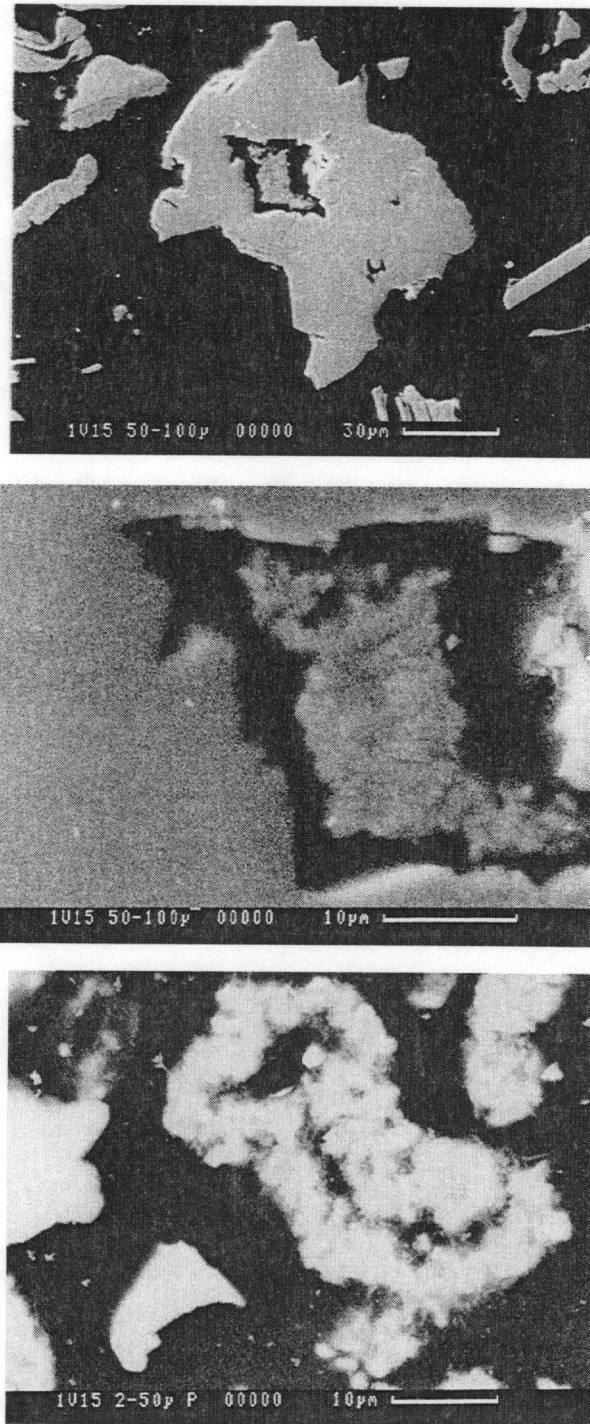


Fig. 4-10. (a) Initial dissolution of an albite grain interior followed by (b) reprecipitation of secondary halloysite within the framework. In another grain (c), the host is entirely replaced by halloysite.

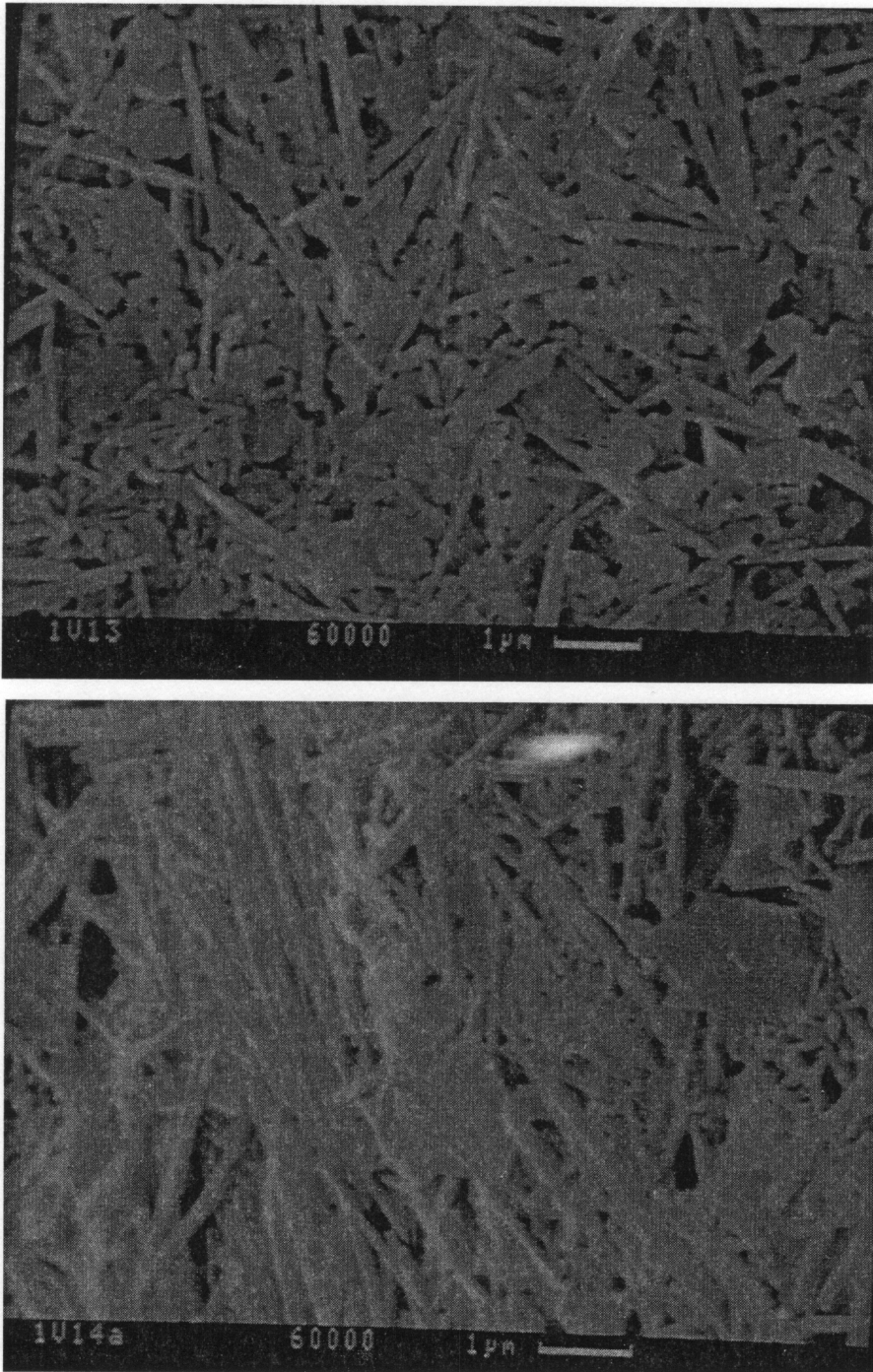


Fig. 4-11. Micrographs of (a) the kaolinite-dominated 2C5b horizon, and (b) the halloysite-dominated 2C6b horizon clay ($<2 \mu$) fractions.

within silt- and sand-size halloysitized albite pseudomorphs, ultimately resulting in the concentric nucleation of kaolinite pseudo-hexagonal plates on the surface of host grains (Fig. 4-12a-b).

In another scenario, kaolinite is formed from the dissolution and resilication of gibbsite at, or just above, a depth of ~5 m in the profile. The gibbsite content of the 2C6b and 2C7b horizon clay fractions, determined by DSC, ranges from 11-17%. The 2C5b horizon clay fraction at <5 m depth is devoid of any trace of gibbsite as shown by DSC analysis and by the complete loss of the 4.85Å gibbsite x-ray diffraction peak (Fig. 4-2, 4-3).

However, the concurrent near tripling of the 7.2Å kaolinite diffraction peak in this same horizon suggests that the influx of Si-enriched leachate from overlying horizons reacts with destabilized gibbsite at a pH of ~4.9-5.0 (Nordstrom and May 1989) to form kaolinite.

Strain Calculations for Pedon VA 1

Volumetric change associated with the transformation of weathered rock to saprolite, and saprolite to soil can be assessed by using depth trends of the concentration of an immobile element, Zr, with an assumed flux of zero (Chadwick et al, 1990; Brimhall et al 1991). Strain, or volume change, results whenever a change in soil bulk density is not compensated by an inversely proportional change in the concentration of the immobile element:

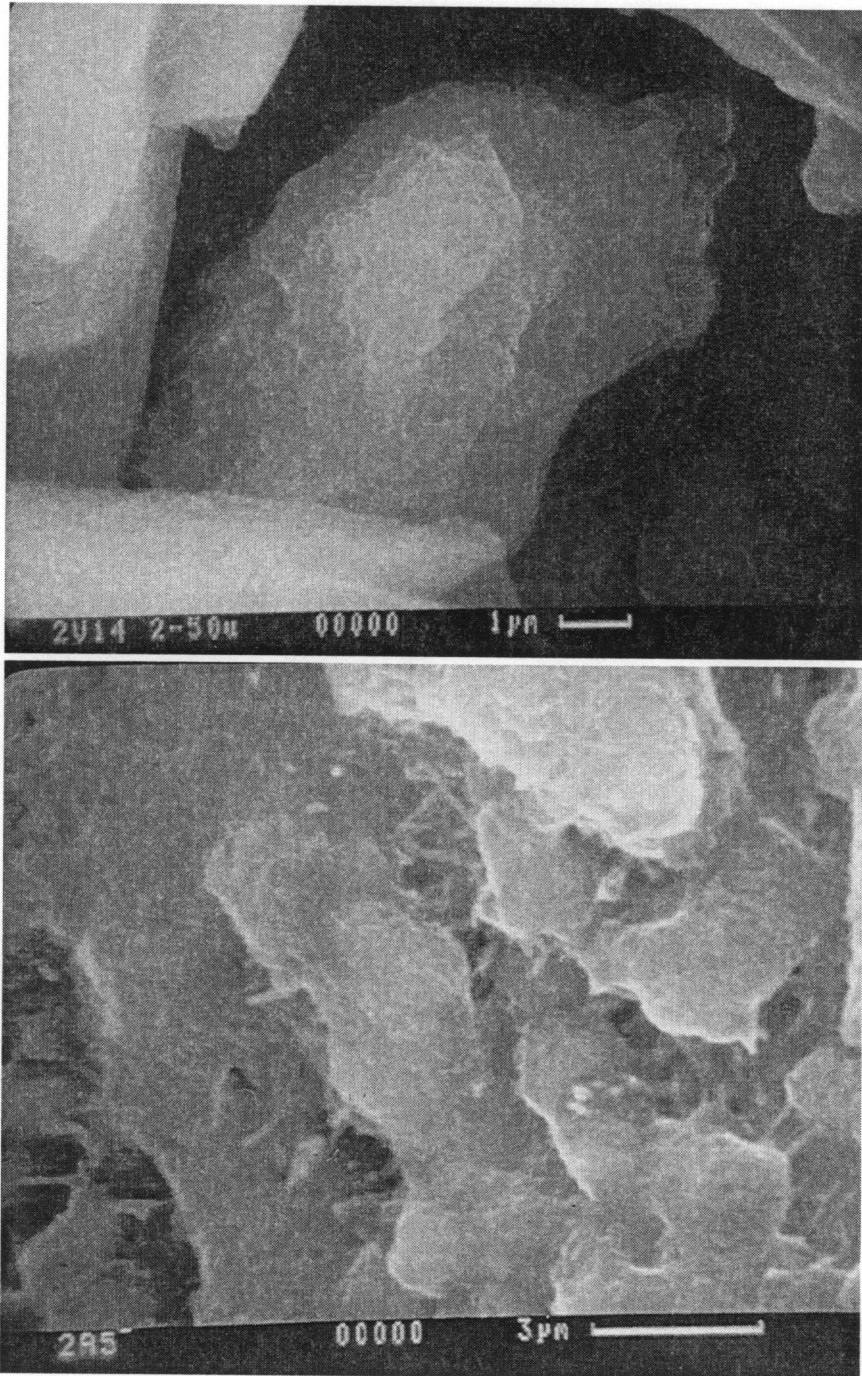


Fig. 4-12. Formation of kaolinite higher in the profile from gibbsite or halloysite precursors involves dissolution and reprecipitation as concentric layers (a), often while maintaining the integrity of the host grain. Note the pseudohexagonal kaolinite crystals forming on a matrix of tubular halloysite (b).

$$\epsilon_{i,w} = \frac{\rho_p C_{i,p}}{\rho_w C_{i,w}} - 1$$

Bulk density (ρ_b) and density ratio (ρ_p/ρ_w) plots show that saprolite horizons are noticeably less dense than weathered rock as would be expected after intense solution weathering and loss of mass (Fig. 4-13a,b). Elemental Zr values show little variation in the loess cap decreasing uniformly with depth in the residual subsoil and saprolite horizons (Fig. 4-13c). The strain profile ($\epsilon_{zr,w}$), indicates that oxidized saprolite undergoes a volume expansion of up to 50% relative to the underlying unoxidized material (Fig 4-13d-e). Nonisovolumetric weathering of saprolite is conceivable when viewed in light of a 300-400% volume increase in biotite grains upon weathering (Feldman, 1995, in preparation) and the structural rearrangement of Al from four-fold to six-fold coordination upon weathering and release to solution.

At a depth of about 200-300 cm, a significant volume reduction upward in the profile marks the transition from saprolite to soil. This depth corresponds to the lower limit of pedogenic clay illuviation, Fe_d accumulation, CEC, and exchangeable Mg in the profile as described previously.

Applying density ratios and strain relationships to the flux of Fe_d within the profile indicates that the weight percent of Fe_d in the profile (Fig. 4-14a), the enrichment of Fe_d

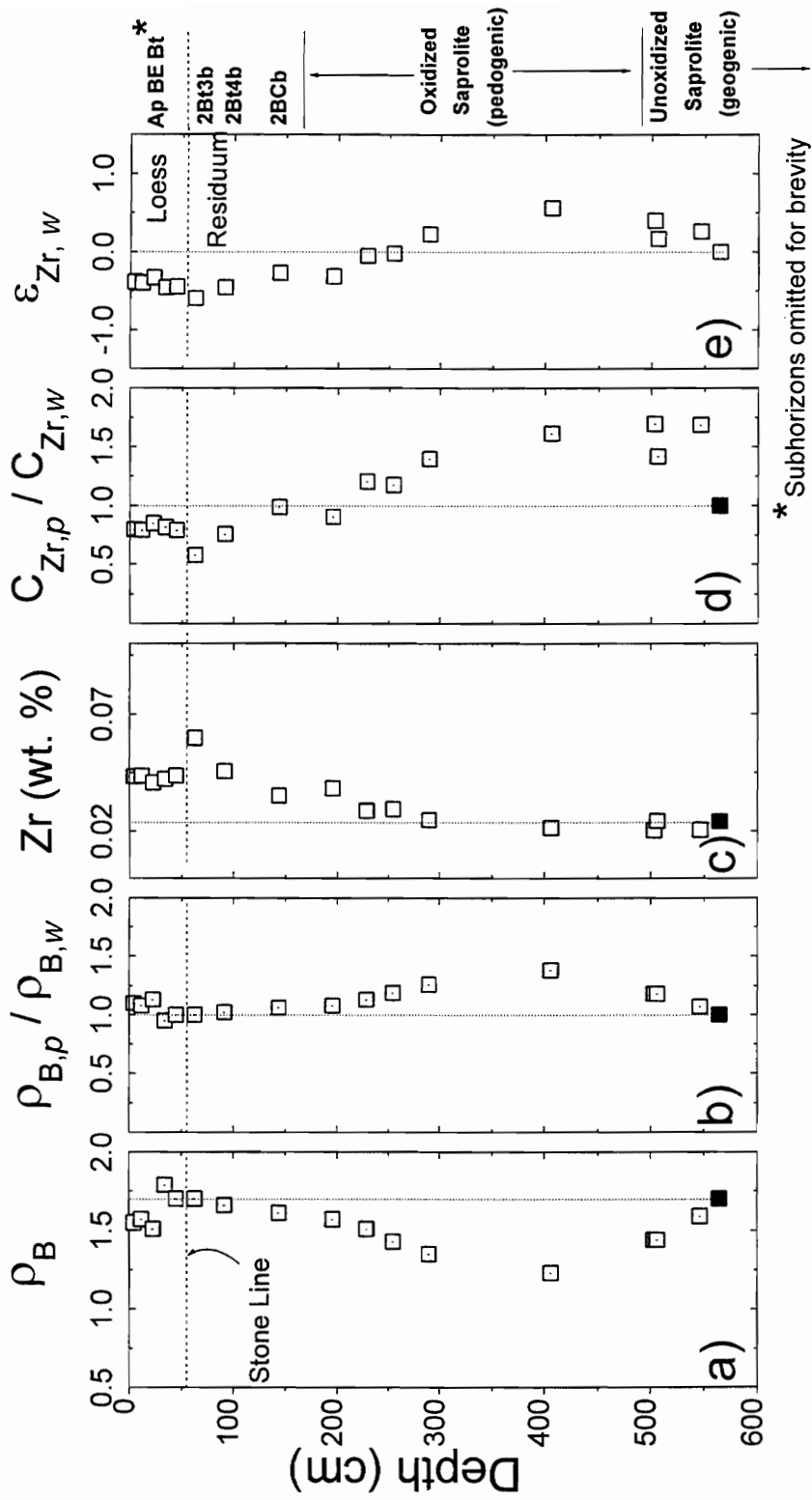


Fig. 4-13. Determination of strain ($\epsilon_{Zr,w}$) associated with the formation of saprolite, and the transformation of saprolite to soil.

in the profile relative to the parent material (Fig. 4-14b), and the Fe_d values corrected for strain and density effects (Fig. 4-14c) are not appreciably different from one another. This similarity results because there are no inherited primary Fe-oxides in the granitic parent material — all Fe_d values reflect the *in situ*, pedogenic development of secondary Fe-oxides. Clay flux in the profile is very similar to Fe_d trends (Fig. 4-15), suggesting that it is entirely pedogenic in nature.

The source of much of the Fe in this profile is weathered biotite, as shown by the decreasing upward amounts of elemental Fe_2O_3 (Fig. 4-14d), beginning at a depth that corresponds to the initial accumulation of Fe_d in the profile (Fig. 4-1c). Trends in K_2O and MgO also corroborate biotite as a dominant source mineral for Fe (Fig. 4-14d).

Summary and Conclusions

Soils formed in a dated loess deposit overlying a paleosol derived from granitic saprolite on the northern Virginia Piedmont exhibit complex polygenetic weathering trends that contrast sharply between parent materials. The climate following deposition of the loess included a period of periglacial conditions at the close of the Pleistocene (Feldman et al., 1995), followed by humid-temperate conditions in the Holocene. A weakly-developed argillic horizon has formed in the loess cap from alteration of weatherable minerals in the fine and medium silt fractions within the last 10 ka, when conditions were warm and the landscape was relatively stable in the mid-Atlantic area. The minimal alteration of coarse silt-size minerals during this period and the lack of

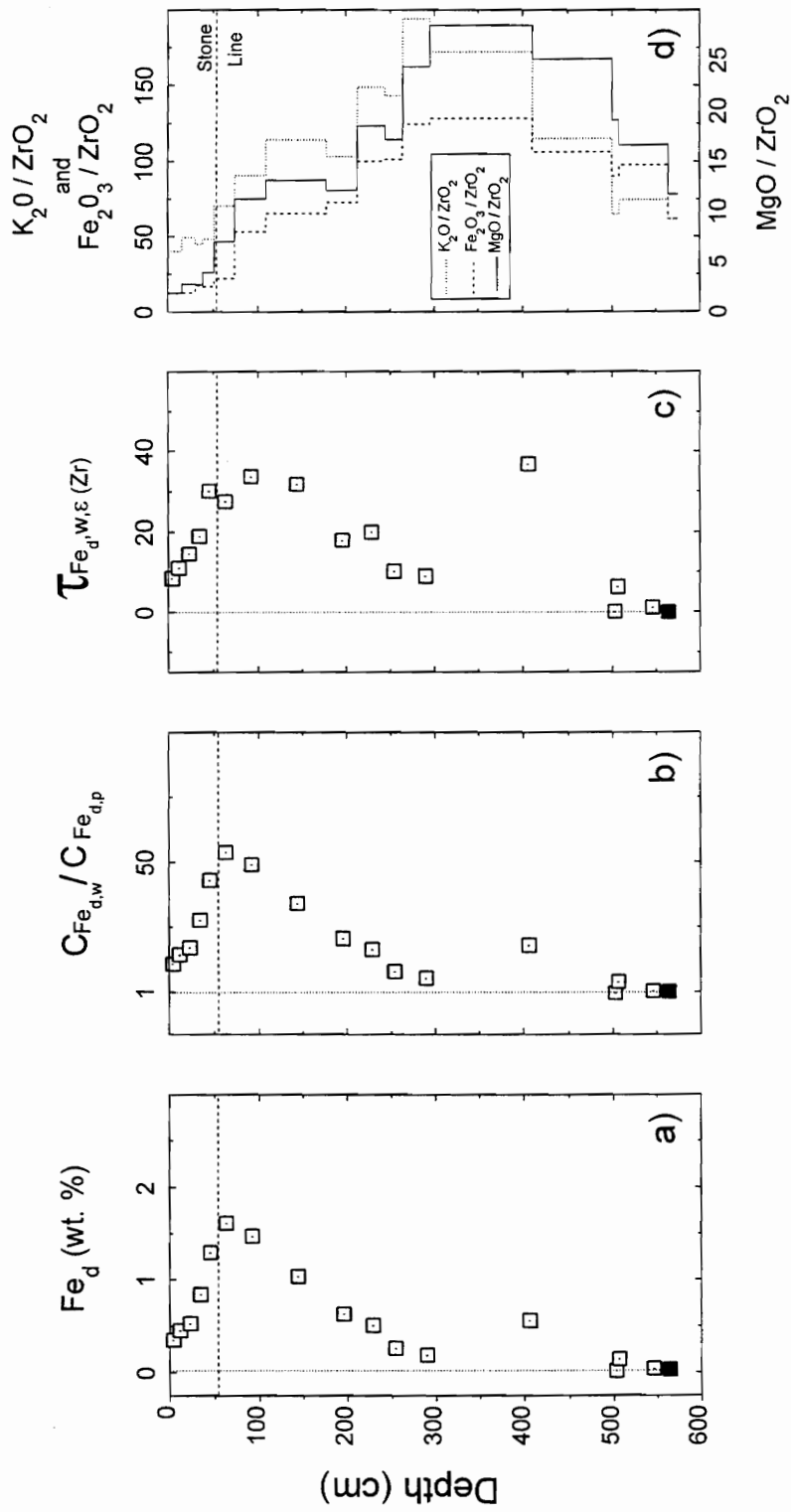


Fig. 4-14. (a) Weight percent Fe_d in the profile, (b) concentration of Fe_d in the soil relative to the parent material, (c) distribution of Fe_d corrected for strain, and (d) distributions of K_2O , Fe_2O_3 , and MgO in the profile relative to ZrO_2 .

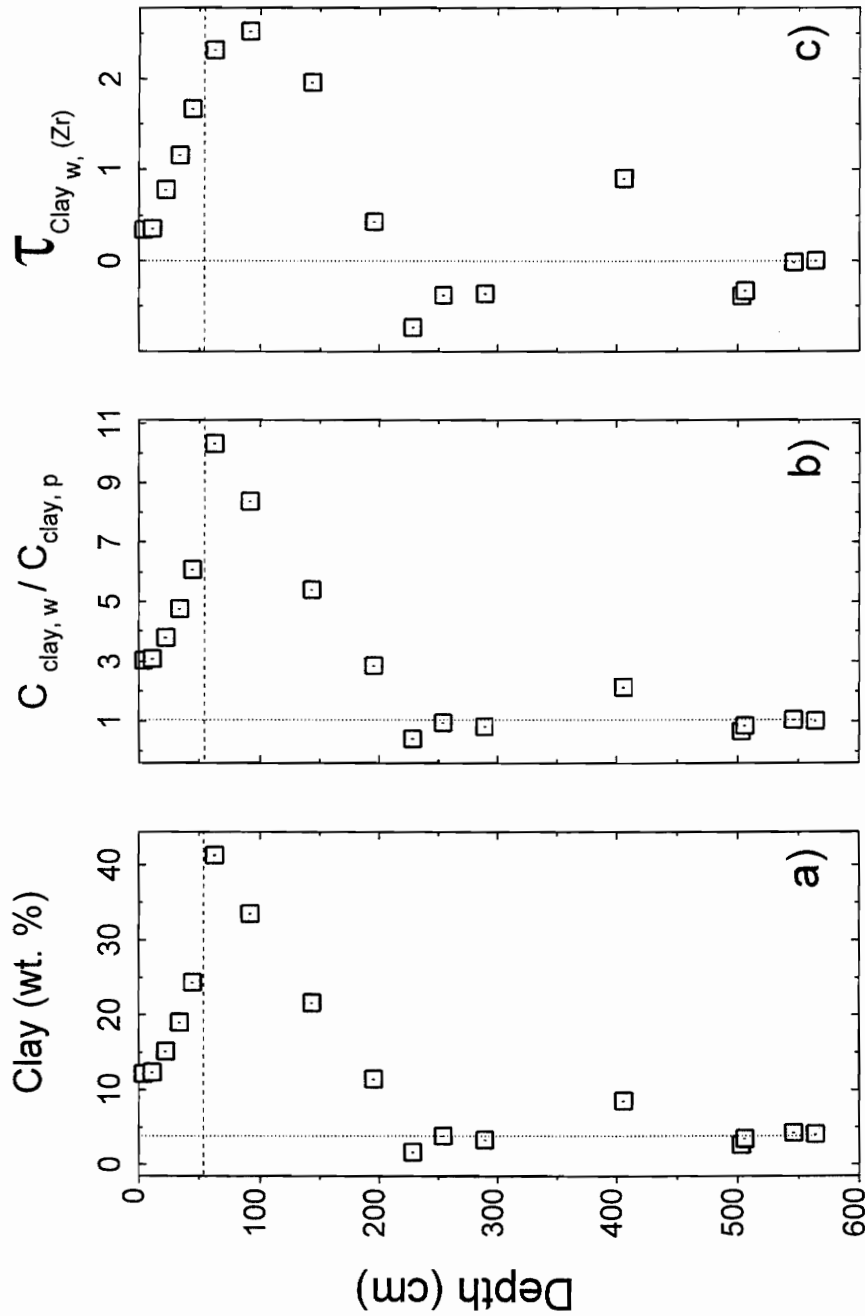


Fig. 4-15. (a) Weight percent clay in the profile, (b) concentration of clay in the soil relative to the parent material, and (c) distribution of clay corrected for strain.

variation in the clay mineral suite with depth in the loess also suggest a relatively limited duration of weathering.

In the underlying paleosol, high levels of exchangeable Mg and a high-charge smectite phase in the clay fraction are inconsistent with 10 ka of weathering under the humid-temperate conditions of intense, acid leaching. We consider these features to be relict soil properties related to drier climatic conditions and lower leaching intensity that could only have been present during or before the time of loess emplacement. Moreover, the paleoargillic horizon exhibits pedogenic features that, when compared with the weathering characteristics of the dated loess deposit, could not have formed within the ~13.8 ka since deposition. Abrupt increases in pedogenic clay content, CEC, and Fe_d , and the development of secondary mineral phases within the coarse silt fraction of the paleosol contrast sharply with the overlying loess cap. It follows that either weathering intensity and leaching rates were markedly greater before the time of loess deposition when the paleosol was developing, or there was a long period of geomorphic stability prior to that time. In either case, the degree of profile development and the mineralogical characteristics are consistent with a pre-Wisconsinan age for this soil.

Because of their complex mechanisms of formation, the abundance of kaolinite and gibbsite in the residual paleosol is not a useful index of soil age, weathering intensity, or paleoenvironment on the northern Virginia Piedmont. Kaolinite forms from at least four separate mechanisms in this soil, including: i) desilication of 2:1 phyllosilicates resulting from intense leaching in surface or near-surface horizons ii) transformation of K-feldspars

in the loess cap silt fraction, iii) resilication of gibbsite formed from plagioclase deep in the saprolite, and iv) transformation of halloysite at a depth of ~5 m in the profile. Gibbsite similarly forms from both desilication of kaolinite in surface horizons, and also by reprecipitation after Na-feldspar dissolution at a depth of >5 m. Halloysite is produced in this soil from both feldspar and biotite precursors, which illustrates the importance of microenvironment in determining the nature of the secondary minerals produced, irrespective of the composition of the primary mineral undergoing transformation.

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Chapter V

Mineralogy, Weathering, and Elemental Distribution Along a Piedmont Soil Climosequence

ABSTRACT

The Southern Piedmont region may be particularly sensitive to potential changes in climate because of high population density and agricultural drought hazard. Changes in temperature and precipitation on the order predicted climate by models through the next century may affect soil physicochemical properties by influencing mineral stabilities and rates of both chemical reactions and leaching. These factors may ultimately cause a northward shift in the distribution and extent of kandic, oxidic, and kaolinitic soils in the southeastern US.

We studied a climosequence of soils formed from structurally and lithologically similar granitic plutons on the Southern Piedmont in order to i) assess the response of soils to variations in weathering intensity, and ii) quantify, for a range of pedogenic processes, the magnitude of soil physical, chemical, and mineralogical changes induced by the natural range of existing climatic regimes across the region.

Morphological, physical, chemical, and mineralogical measures of weathering all increased dramatically from north to south on the Piedmont over similar rock types.

Indices of soil rubification, extent of desilication features, accumulation and distribution of pedogenic clay and Fe-oxides, and the depth at which pseudomorphic gibbsite and halloysite forms from feldspar and biotite in these profiles, is nearly twice that of soils in Virginia and North Carolina. Of all the soils studied, only those in Alabama appear to have reached a steady-state with regard to pedogenic features such as clay mass. This indicates a condition of either greater stability and longer weathering, or more intense weathering farther south on the Piedmont.

Weathering trends for all locations closely correspond to calculated effective leaching indices developed for each site based on modern climatic averages. This suggests that despite widespread absolute climatic fluctuations during the Quaternary, the relative differences in mean annual temperature, precipitation, and potential evapotranspiration that now exist from north to south on the Piedmont may be analagous to the long-term climatic signal under which these soils formed.

Introduction

Strong relationships between climate, primarily temperature and precipitation, and soil properties such as solum thickness, depth to carbonates, N and clay contents, pH, and CEC have long been known to exist (Jenny, 1941; Arkley, 1963, Ruhe, 1984a). In the southeastern U.S., soil properties are dramatically influenced by different temperature and moisture regimes, largely through effects on solubilities of solid and gaseous phases, chemical reaction rates, and speciation of ions in soil solution (Buol et al., 1990). These

factors, and additionally the rate at which drainage water leaches through the soil, dictate to a large extent the nature and products of primary and secondary mineral weathering, degree of soil profile development, magnitude and depth of clay illuviation, extent of soil rubification, and the surface charge properties of the soil exchange complex.

In the nonglaciaded Southern Piedmont, soil rubification, weathering intensity, and degree of profile development on similar parent materials increase noticeably from north to south (Kaster, 1960; Porter et al., 1963; Frost, 1979) as do reported increases in both hematite and gibbsite, and decreases in goethite in this same direction (Southern Regional Project S-14, 1959). A similar change from soils having dominantly mixed mineralogy to those having kaolinitic mineralogy (Soil Survey Staff, 1990) also closely coincides with a shift from mesic soils (MAST¹ 8° to 15°C) in the northern part of the Piedmont to thermic soils (MAST 15° to 22°C) in the southern part.

The Piedmont region may be particularly sensitive to potential changes in climate because of high population density and agricultural drought hazard. Properties of Piedmont soils that would be expected to change in response to changes in the weathering environment within 10 to 100 years are generally restricted to levels of soluble and exchangeable nutrients, amount and quality of organic C and N levels, water availability and soil hydrologic properties, and effects on microbial processes (Rounsevell and Loveland (1994). In the longer-term, soil properties that may be influenced by different temperature and moisture regimes include solubilities of solid and gaseous phases,

¹ Mean annual soil temperature (at 50 cm depth)

chemical reaction rates, and speciation of ions in soil solution (Buol et al., 1990). Millennia-scale effects could include changes in Si solubility, alterations to both crystalline and noncrystalline Fe and Al oxides, and increases in the amount of hydroxy-Al or -Fe interlaying in vermiculite interlayers. These effects on Fe- and Al-oxides and vermiculite interlayering may be especially important because increases in these components would significantly modify soil physicochemical properties resulting, partly, in reduction of cation exchange capacity (CEC), increases in anion retention characteristics, increases in K^+ selectivity, and changes in flocculation/dispersion phenomena which influence soil structure, surface crusting, and rates of infiltration, runoff, and erosion. These factors may potentially cause a northward shift in the distribution and extent of kandic, oxidic, and kaolinitic soils (Soil Survey Staff, 1990) in the southeastern U.S. Expanded areas of these low-fertility soils would result in greater potential for groundwater pollution through higher rates of fertilizers, herbicides, and insecticides required to sustain agricultural productivity.

Despite the acknowledged dominion of temperature, precipitation, and (or) potential evapotranspiration over many Piedmont soil properties, there is currently a scarcity of quantitative data regarding the specific effects of both climate and time on weathering processes, owing largely to a lack of age control for deposits >40,000 yr (Mills and Delcourt, 1991). We presently cannot differentiate between the combined effects of climate and geomorphic age on weathering intensity, soil profile development, mineral stability and transformation, and elemental flux within the pedoenvironment (Boardman,

1985; Chapter V, this volume), nor do we understand how these parameters will be affected by future changes in the chemical and physical climate. Additionally, we know relatively little about rates of geochemical weathering and pedogenic processes under field conditions (Velbel, 1986), and even less about minimum climatic thresholds required to initiate changes in steady-state soil systems and processes. Without this information, we can neither identify the degree to which soil properties and secondary phases persist, alter, or become obliterated in the weathering environment as conditions change, nor can we predict with any certainty which soil features are reliable as indicators of retrospective or prospective environmental change.

The principal objectives of this study are to i) investigate whether the response of soils to variations in the weathering environment follows predictable, systematic patterns related to climate on the Southern Piedmont, and ii) quantify, for a range of pedogenic processes, the magnitude of soil physical, chemical, and mineralogical changes induced by the natural range of existing climatic regimes across the region.

Materials and Methods

Setting, Site Selection, and Field Sampling

A climosequence approach was employed to study soil development from north to south on the Piedmont. While it is virtually impossible to exert rigorous experimental controls on open, natural soil systems, the similarity of soil parent materials and landscape features selected for study in this project should permit accurate analysis of soil

properties as a function of climate. Duplicate soil pedons were sampled at each of four sites located in northern Virginia, north-central North Carolina, northeastern Georgia, and east-central Alabama, respectively (Fig. 1-1). These sites span the modern climatic extremes on the Piedmont (Fig. 1-2, 1-3; Table 1-1). They also encompass the range of temperature and precipitation predicted by climate models under several global warming scenarios through the next century (Houghton et al., 1990).

Topographic and parent material effects were minimized by sampling exclusively on stable upland interfluves underlain by lithologically and structurally similar granitic plutons which have been well-characterized (Adams, 1933; Gault, 1945; Deininger, 1975; Neathery et al., 1975; Seiders et al., 1975; Briggs et al., 1978; Stormer and Whitney, 1980; Drake 1986; Drake and Froelich, 1986; Drummond and Green, 1987). These lithostratigraphic units exhibit a relatively high degree of homogeneity both within and between sites, particularly with regard to bulk chemical composition and mineralogy (Tables 1-2 and 1-3). Total Fe content is very similar between all four granites, which will permit independent determination of the effects of climate on the formation and transformation of secondary Fe-oxides.

Duplicate soil pedons, separated by not less than 2 km, were sampled at each of the four study areas (Fig. 1-1). All pedons were located on high, broad, nearly level or gently convex interfluves to ensure that areas selected for detailed study have formed in residuum and are representative of soils exhibiting maximum pedogenic development in the landscape with minimal likelihood of lithologic discontinuity. These areas are neither

highly erosive nor subject to overwash from other landscape positions and thus the transport of recent sediments by slope processes, either to or from each site, was expected to be minimal. Representative soil pits were excavated with a backhoe to depths of ≥ 2 m and described using standard terminology (Soil Survey Staff, 1981). Samples for characterization and bulk density analysis (Blake and Hartge, 1986) were collected by genetic horizon from one area of the pit chosen to be representative of conditions in the pit. Deeper samples (>5.5 m) were obtained at ~ 50 cm intervals using a modified bucket auger method (Stolt et al., 1991), which also permitted bulk sampling and also sampling for bulk density determinations.

Summary of Bedrock Geology

Mean chemical composition of representative rock samples from each study site is very similar with respect to both total Fe and FeO contents for all but the Alabama rocks, which are significantly higher in these parameters (Table 1-2). Silica and Al levels are also nearly equal for rocks of each of the three northernmost sites, but the Alabama rocks are slightly more mafic, with less Si and more alkaline earth metals.

Occoquan Granite, Virginia

The Occoquan Batholith in Fairfax and Prince William Counties, Virginia, consists of medium-grained, light gray monzogranite with moderate to strong metamorphic foliation and a well-developed mineral lineation. Major minerals include quartz, plagioclase (commonly with inclusions of sericite), and microcline with subordinate muscovite,

biotite, and epidote (Seiders et al., 1975) (Table 1-3). Plagioclase varies between albite (Ab_{94-95}) and a mixture of albite and oligoclase (Pavich, 1986). Microcline is generally free of inclusions and is commonly untwinned. Accessory minerals include apatite, sphene, garnet, chlorite, allanite, calcite, and zircon.

Roxboro Formation, North Carolina

The Roxboro Metagranite is located in Person and Caswell Counties, North Carolina. A foliation and preferred orientation of biotite, muscovite, chlorite, and epidote was apparently developed in this pluton during a deformational event at ~575 Ma B.P., which also resulted in a significant modification of the original igneous mineralogy. Quartz and nearly equal amounts of both alkali and plagioclase feldspars comprise 84 to 95% of the total mineral suite with biotite, muscovite, chlorite, and epidote making up the bulk of the remainder (Briggs et al., 1978) (Table 1-3). All K-feldspars are nearly pure microcline (Or_{97}), and all plagioclase is nearly pure low albite (Ab_{97-99}).

Elberton Granite, Georgia

The Elberton Batholith intrudes into the high-grade metamorphic terrain of the Inner Piedmont region of eastern Georgia in a northeast-striking area approximately 60 km long and 10 km wide, extending from south of Lexington, Georgia to north of Elberton, Georgia. The Elberton Granite is a fine- to medium-grained light-gray granite that shows no evidence of metamorphic deformation or metamorphic fabric (Stormer et al., 1980). The entire pluton is reported to be the product of a single intrusive event that occurred

~330 Ma B.P., during the peak of tectonic and metamorphic events in the area (Stormer et al., 1980).

The Elberton Granite exhibits extreme mineralogical and chemical homogeneity despite minor variations in color and grain-size. Darker colors of the rock are attributed to impurities in the feldspars (Ramspott, 1964). Variations in grain-size across the entire 500 sq. km exposure are slight and gradual. Compositionally, the rock contains ~30% quartz, 30-35% plagioclase, and 30-35% potassium feldspar (Ramspott, 1964; Stormer et al., 1980) (Table 1-3). Biotite content ranges from about 4-7%. Common accessory minerals include sphene, zircon, allanite, magnetite, apatite, and ilmenite-hematite. Muscovite, epidote, and chlorite may be locally important, but these minerals are reported to be the secondary products of post-emplacement alteration (Whitney et al., 1980). Plagioclase is predominantly unzoned oligoclase (Ab_{74-86}) and K-feldspar is mainly microcline with a composition of $\sim Or_{95}$ (Table 1-3).

Elkahatchee Quartz Diorite, Alabama

Igneous and metamorphic rocks that dip under the Alabama Coastal Plain sediments to the south are at the southern limit of the crystalline Appalachian system. The Elkahatchee Quartz Diorite makes up the largest part of the Pickneyville Quartz Diorite Complex, occupying an area about 65 km long and up to 20 km wide that strikes northeast in eastern Coosa and northwestern Tallapoosa Counties, Alabama (Gault, 1945; Deninger, 1975). The Elkahatchee is a medium- to dark-gray, coarse-grained mesocratic rock with well-defined foliation, lineation, and local textural or light-dark banding

ranging from millimeters to meters. Essential minerals, comprising >90% of the mineral suite, include plagioclase, quartz, and biotite which are recognizable in hand specimen. Smaller amounts of muscovite and epidote make up the remainder (Gault, 1945) (Table 1-3). Accessory minerals include apatite, sphene, zircon, pyrite, allanite, garnet, and K-feldspar. Plagioclase ranges from labradorite (Ab₄₅) to albite (Ab₉₅), but sodic andesine and oligoclase are most abundant. Light-colored quartzitic, granitic dikes up to about 30 cm thick are widespread in the Elkahatchee.

Emplacement of the Elkahatchee is reported to have occurred during the Late Cambrian period (~520 Ma B.P.), and a regional metamorphic event is thought to have occurred at ~300 Ma B.P. (Neathery et al., 1975).

Modern Climate and Predicted Change

Modern climate of the Southern Piedmont is humid, temperate to subtropical, spanning a considerable temperature, precipitation, and evapotranspiration gradient from about 32° to 41°N latitude (Thornbury, 1965). Mean annual temperature across the region ranges from about 12° to 17°C and is highly correlated with latitude (Table 1-1). Southern sites experience temperatures about 7°C higher than northern areas during winter months but only about 3°C higher during the summer (Fig. 1-2). Mean annual precipitation ranges from about 1000 mm in the north to about 1500 mm in the south but shows considerably more variability than temperature trends. Precipitation is relatively uniformly distributed throughout the year at each site, with maximum values occurring

during the late fall and relative increases occurring in early spring and mid-summer (Fig. 1-2). Calculated potential evapotranspiration trends closely parallel those of temperature, but these values essentially converge for all sites during periods of maximum summer temperature (Fig. 1-2).

U.S. Department of Energy models project the equilibrium climate response to a doubling of atmospheric CO₂ (or its radiative equivalent from all greenhouse gases) to range from 1.5° to 4.5°C (Schlesinger and Mitchell, 1985). Under the scenario of current emissions of greenhouse gases, the World Meteorological Organization through the Intergovernmental Panel on Climate Change (IPCC), estimates that global mean temperatures will increase at a rate of about 0.3° to 0.4°C per decade (Houghton et al., 1990). At this rate, temperatures will increase about 1°C above the present value by the year 2025, with a projected increase of about 4.5°C by the end of the next century. If emissions are subject to controls, the most optimistic IPCC estimates indicate that average rates of temperature increase can be limited to about 0.1°C per decade, or an increase of about 2.0°C by the year 2100.

Water balance data for the study areas indicates that soils in lower latitudes experience less summer moisture deficit and have higher capacity for deep percolation below the solum than soils farther north (Fig. 1-3, Table 1-1). Comparison of these values with global changes predicted by climatic models for the end of the next century suggests that the proposed study area covers the range of temperature increase expected for the Piedmont through the year 2100.

Physical, Chemical, and Mineralogical Analyses

Samples of genetic soil horizons were air-dried, sieved to remove coarse fragments (>2 mm), and thoroughly mixed. The following analyses were made using standard methods (Soil Survey Staff, 1984): particle-size analysis by the pipette method, organic carbon by wet oxidation with Na-dichromate and back-titration with 0.5 M $\text{Fe}(\text{NH}_4)_2(\text{SO}_4)_2$, and CEC by the sum of neutral M NH_4OAc -extractable bases (pH 7.0) plus total soil acidity by BaCl_2 -TEA (pH 8.2). Soil pH was measured in the supernatant portion of a 1:1 soil/deionized water suspension after a 1 h equilibration period.

Samples were prepared for mineralogical analysis by pretreating with 30% H_2O_2 , adjusted to pH 5 with M NaOAc to remove organic matter (Kunze, 1965), and with Na-dithionite-citrate-bicarbonate (DCB) to remove Fe-oxide coatings (Holmgren, 1967). Samples were adjusted to pH 9.5 with M Na_2CO_3 to effect particle dispersion. Sand was separated by wet sieving; clay and silt fractions were separated by repeated decantation after centrifugation. The dominant sand fractions of selected horizons were examined petrographically to quantify mineral suites by the line count method (Galehouse, 1971). X-ray diffractograms were obtained from semi-oriented powder mounts of the coarse (20-50 μ) and medium (5-20 μ) silt fractions, and from oriented clay (<2 μ) fractions using i) Mg-saturated, glycerol-solvated treatments at both room temperature and after 4 hours of heating at 110°C, and ii) K-saturated treatments involving scans collected at room temperature, and after 4 hours heating at 110°, 300°, and 550°C, respectively. Selected clay samples were treated with formamide to differentiate kaolinite from halloysite by

intercalation (Churchman et al., 1984). Kaolinite and gibbsite were quantified by mass-equivalent calibration of differential scanning calorimetry (DSC) endothermic peak areas using poorly crystalline Georgia kaolinite and Reynolds synthetic gibbsite (RH-31F) as standards.

Whole-soils were extracted for reductant-soluble and poorly crystalline Fe and Al using Na-dithionite-citrate-bicarbonate (Holmgren, 1967), and 0.2 M NH_4^+ -oxalate solution in darkness (Hodges and Zelazny, 1980), respectively. Total elemental analysis of both whole soils (<2 mm) and whole sand separates (0.05-2.0 mm) was performed using total digest in HF followed by solution analysis by ICPEs.

Results and Discussion

Physical and Chemical Characterization

Morphological characteristics (Appendix A) and classification of the soils sampled (Table 5-1) denote a distinct zonation in soil color and depth of profile development from north to south. Soils formed on similar granitic parent materials in Virginia and North Carolina are Kanhapludults with colors suggesting goethitic (yellowish-brown) mineralogy, whereas those from Georgia and Alabama are more highly rubified, with rhodic epipedons that indicate hematitic (yellowish-red) mineralogy. Because deep bucket auger sampling made it impossible to differentiate horizonation at depth in the soils sampled, solum depths based on field descriptions are not accurately defined and do not reflect weathering intensity or degree of profile development at each site (Fig. 5-1a).

However, assuming that soil colors of 7.5YR or redder are solely the result of pedogenic processes involved with Fe-oxide formation, the depth to which 10YR hues become the dominant matrix in each profile color can be used as an weathering index. Using this approach, the effective depth of soil profile development in Georgia and Alabama is nearly twice that of Virginia and North Carolina soils (Fig. 5-1b). Effective depth of pedogenesis is closely correlated with leaching indices developed for each site based on modern climatic averages (Table 1-1), suggesting that despite widespread Pleistocene climatic fluctuation, the range of modern climatic conditions may be an appropriate analogue for long-term weathering signal under which Piedmont soils formed.

Table 5-1. Classification of the 8 primary pedons sampled.

Pedon	Classification
VA 1	Typic Kanhapludult, fine-loamy, mixed, mesic
VA 2	Typic Kanhapludult, clayey, mixed, mesic
NC 1	Typic Kanhapludult, clayey, kaolinitic, thermic
NC 2	Typic Kanhapludult, clayey, kaolinitic, thermic
GA 1	Typic Rhodudult, clayey, oxidic, thermic
GA 2	Typic Rhodudult, clayey, oxidic, thermic
AL 1	Typic Rhodudult, clayey, oxidic, thermic
AL 2	Typic Rhodudult, clayey, oxidic, thermic

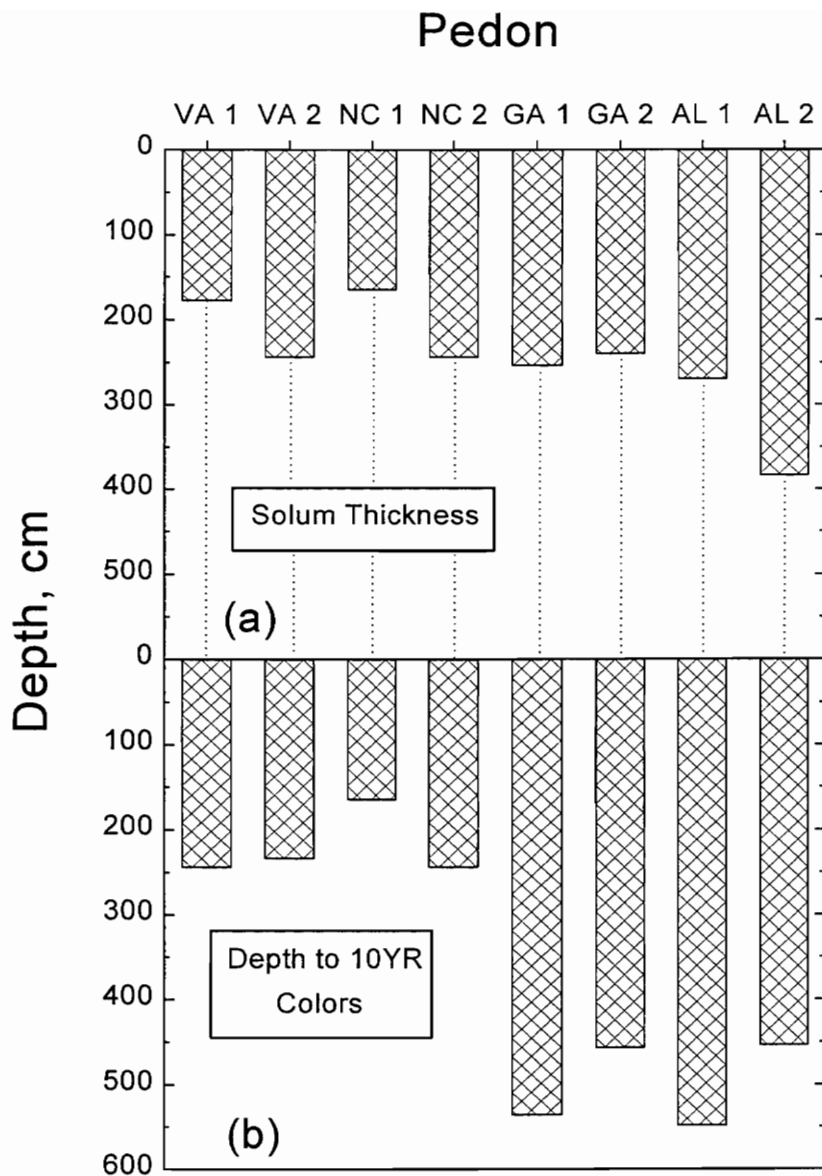


Fig. 5-1. Effective depth of profile development assessed by (a) field estimates of solum thickness, and (b) depth to 10YR color dominance.

All soils have clayey particle size control sections except for pedon VA 1, which is overlain by a 0.8-1.0 m late glacial loess cap (Table 5-1). An apparent gradient in temperature and leaching intensity exists with respect to the mineralogical control section, which is mixed in Virginia, kaolinitic in North Carolina, and oxidic in both Georgia and Alabama.

Maximum clay content ranges to ~65% in each pedon and decreases to ~5% for all soils in the saprolite, below the zone of pedogenic activity (Fig. 5-2,a-d). The lowermost depth of clay illuviation reaches ~200-250 cm in Virginia and North Carolina soils, and increases to ~300-400 cm in Georgia and Alabama. Maximum pedogenic clay, expressed as a weight percent, is not appreciably greater in Georgia and Alabama soils despite deeper weathering depths and both higher temperatures and moisture availability. Expressed on the basis of mass per unit area of soil (weight % x bulk density x horizon thickness, summed to a uniform depth of 200 cm) clay trends increase slightly toward the south in close association with pedogenic Fe-oxides (Fe_d) (Fig. 5-3a).

Both clay content and Fe_d , which are significantly related (Fig. 5-3b), are important because they are both introduced into these soils entirely by pedogenic processes largely involving the alteration of primary Fe-bearing minerals. They therefore directly reflect the relative age and degree of weathering of the soils. When grouped by state, the relationship between clay content and Fe_d in Virginia, North Carolina, and Georgia soils is highly linear, with decreasing slopes at each location, respectively (Fig. 5-3, a-c).

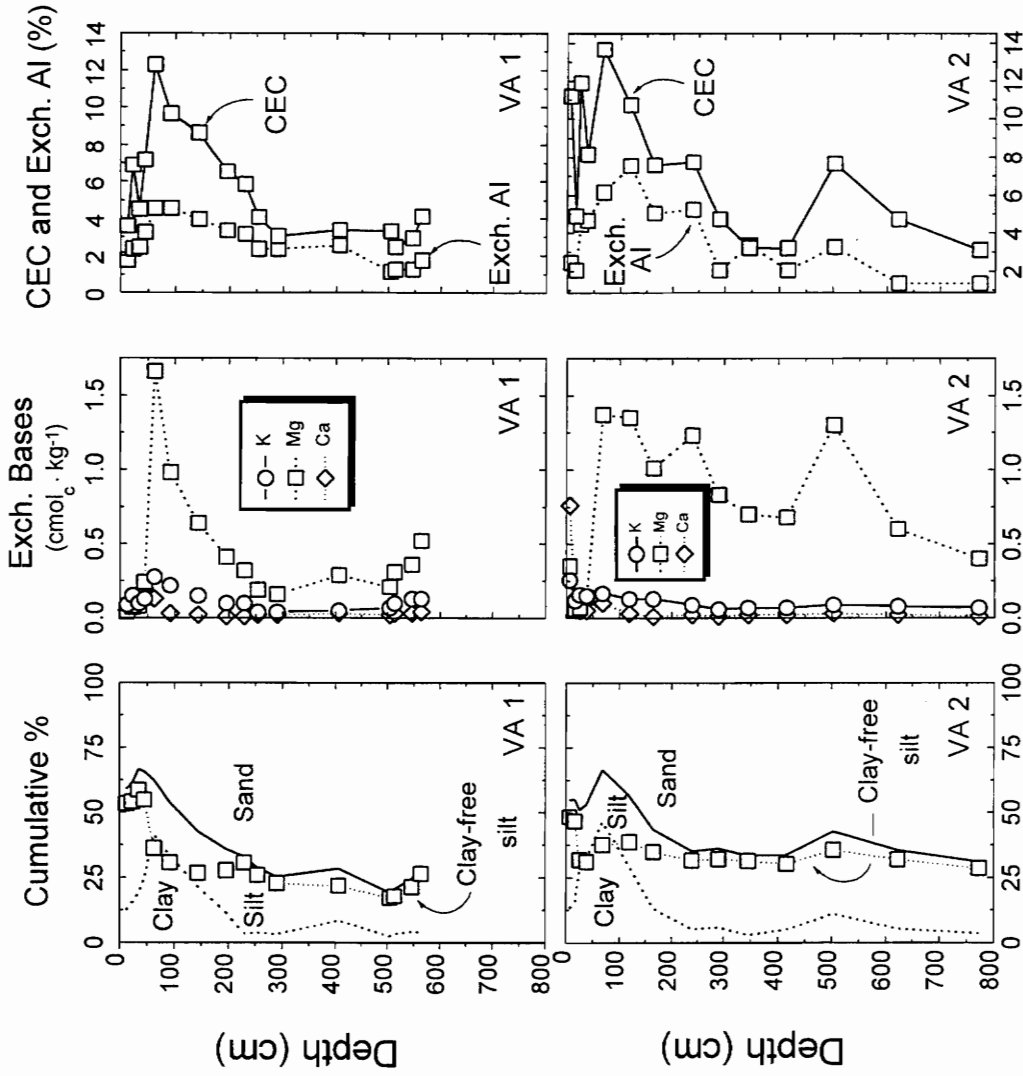


Fig. 5-2(a). Depth functions of significant physical and chemical properties for each of the two pedons studied in Virginia.

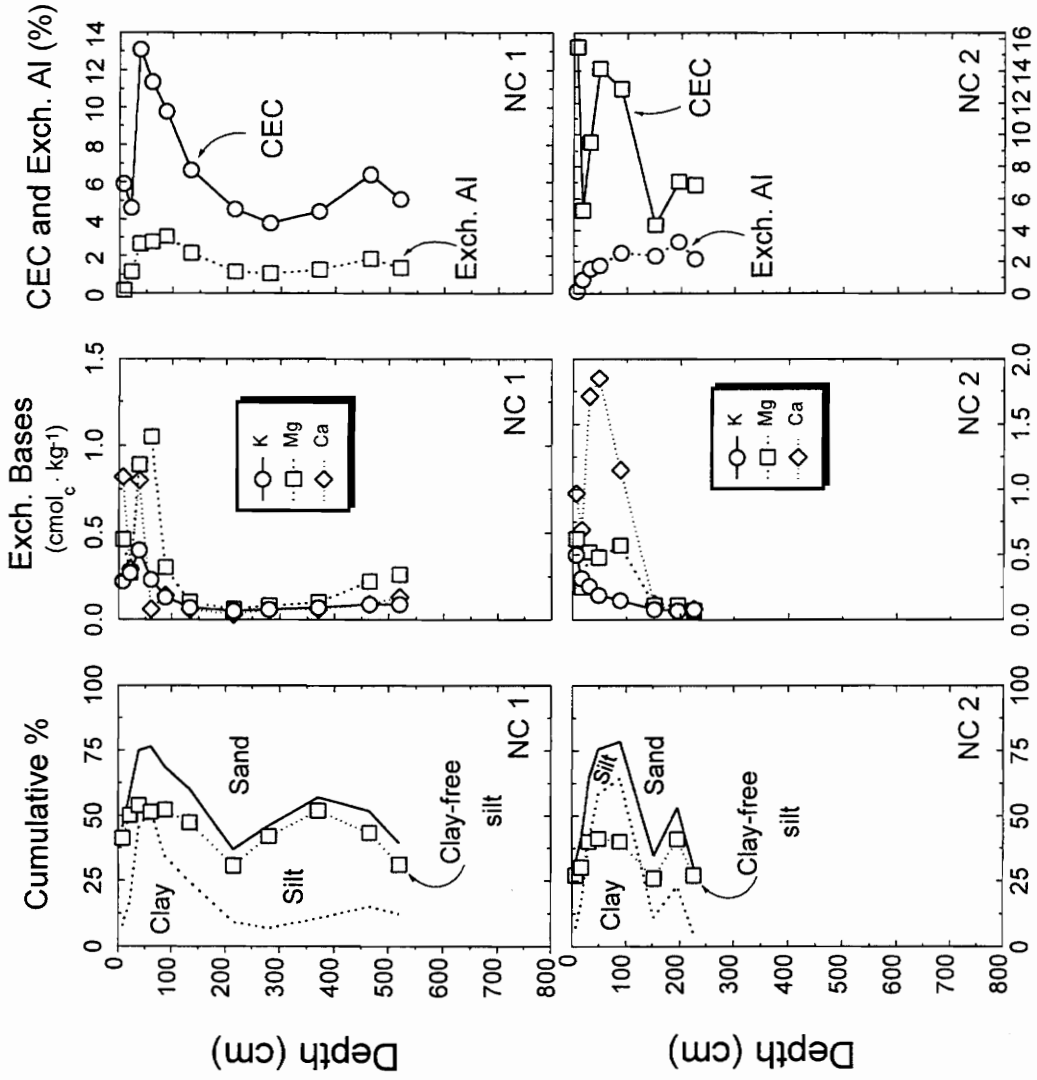


Fig. 5-2(b). Depth functions of significant physical and chemical properties for each of the two pedons studied in North Carolina.

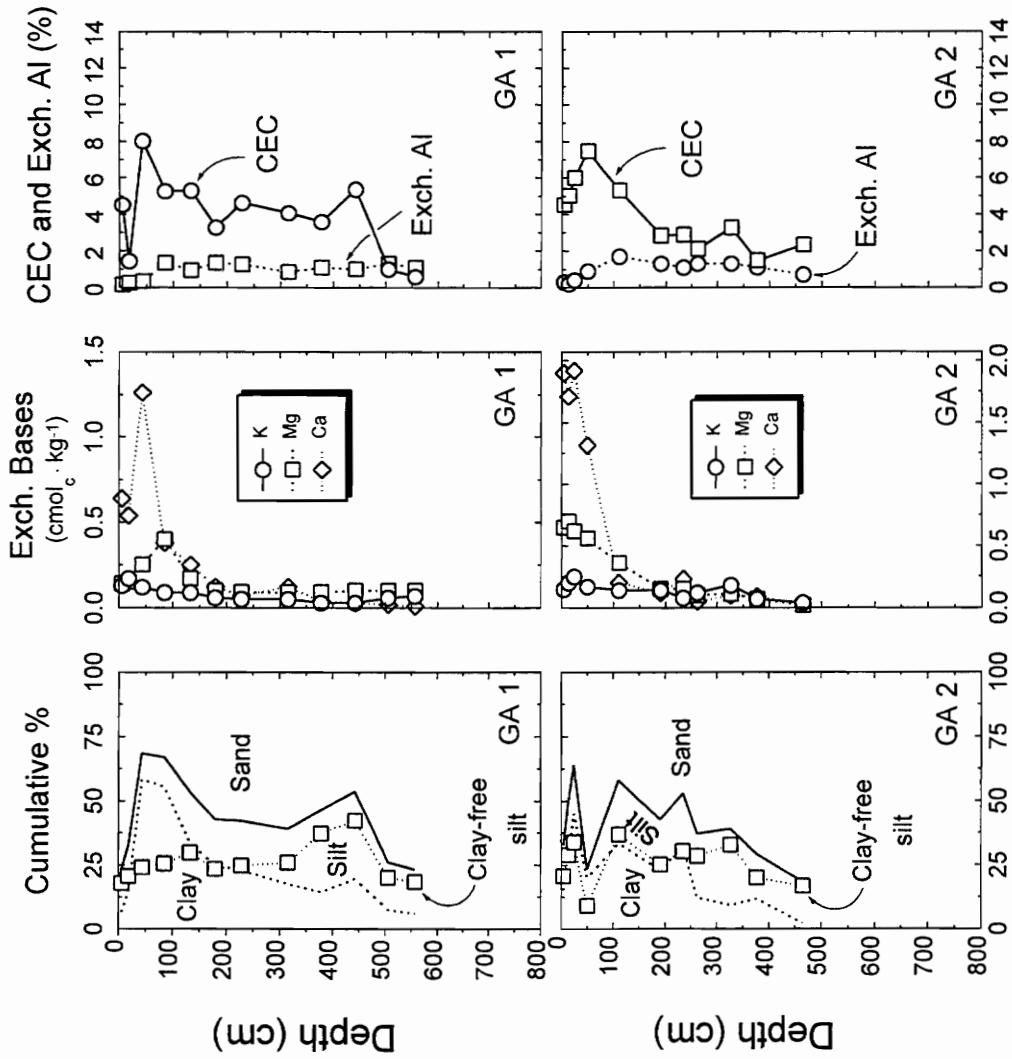


Fig. 5-2(c). Depth functions of significant physical and chemical properties for each of the two pedons studied in Virginia.

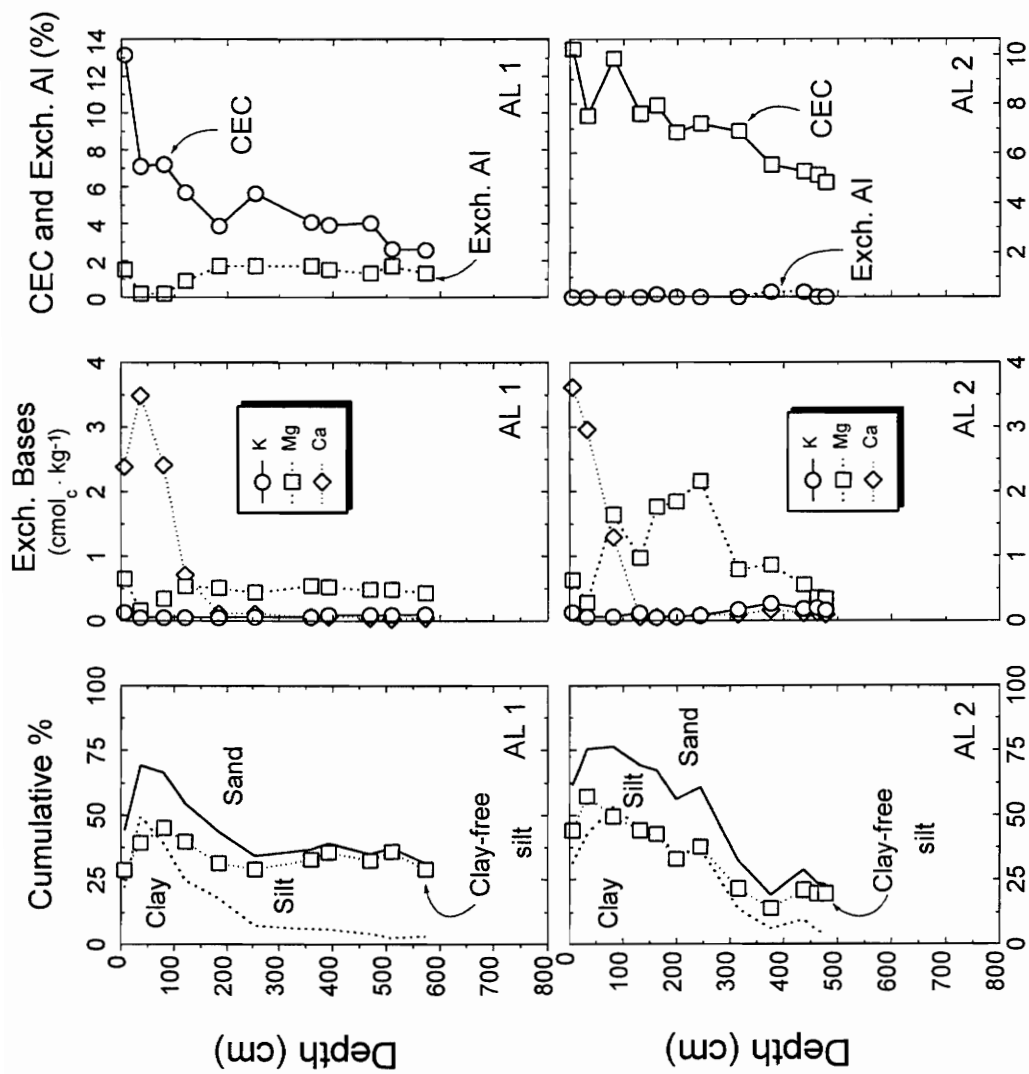


Fig. 5-2(d). Depth functions of significant physical and chemical properties for each of the two pedons studied in Alabama.

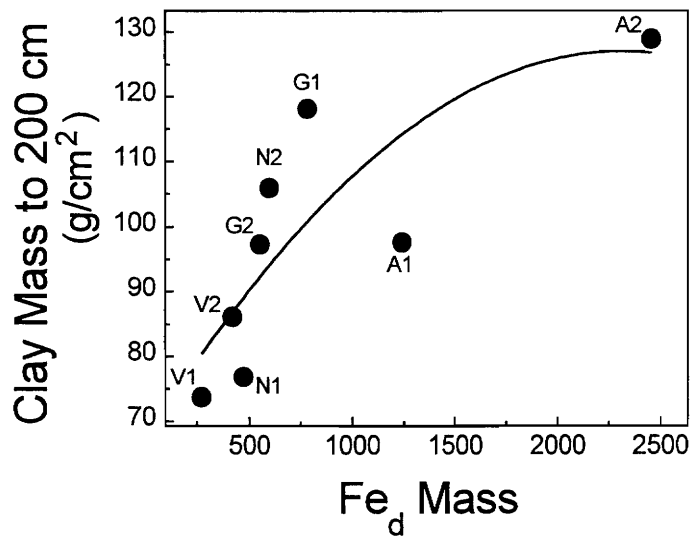
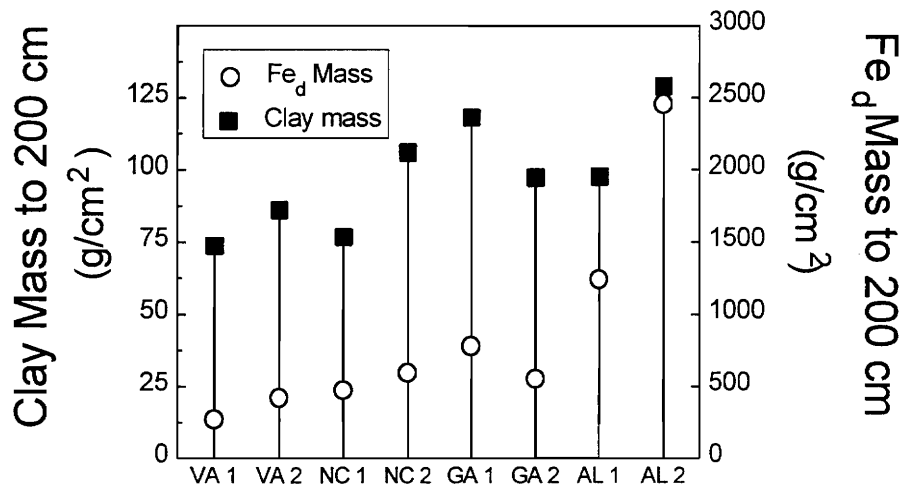


Fig. 5-3. (a) Increasing trends in both clay content and Fe_d, expressed on the basis of mass per unit area of soil (weight x bulk density x horizon thickness, summed to a uniform depth of 200 cm), and (b) the relationship between clay mass and Fe_d.

This suggests that i) Virginia soils, in the most moderate climate, are considerably older than most Holocene soils, which typically have a low correlation between clay content and Fe_d (McFadden and Hendricks, 1985), and ii) a steady-state has not been reached in these three sites with respect to clay production and the continued alteration of unstable Fe-bearing minerals. It should be noted that three adjacent saprolite horizons from pedon GA 2 (shown as crosshatched symbols) were considered to be outliers and omitted from the analysis based on variation in the parent material composition (e.g., a pocket of mafic material) at this site.

In Alabama, soils appear to be approaching steady-state conditions as indicated by the decreasing clay production at higher Fe_d levels (Fig. 5-4d). This suggests that clay formation and illuviation are self-limiting processes attained as some intrinsic threshold value of clay content is reached, thereby increasing available water holding capacity, limiting percolation rates, decreasing mineral dissolution, and promoting quasi-equilibrium conditions with respect to reactive minerals. The fact that Alabama soils formed in the most mafic of all bedrock types in this study indicates that the limiting factor in clay production over time is leaching intensity, and not the concentration of Fe-bearing primary minerals.

Based on the limited availability of exchangeable bases, all soils are highly leached. The relatively high levels of Ca and Mg in Alabama soils (Fig. 5-2d) reflects the mafic character of the parent Elkahatchee Formation (Table 1-2). Virginia, North Carolina, and Georgia soils show moderate amounts of exchangeable K in surface horizons, reflecting

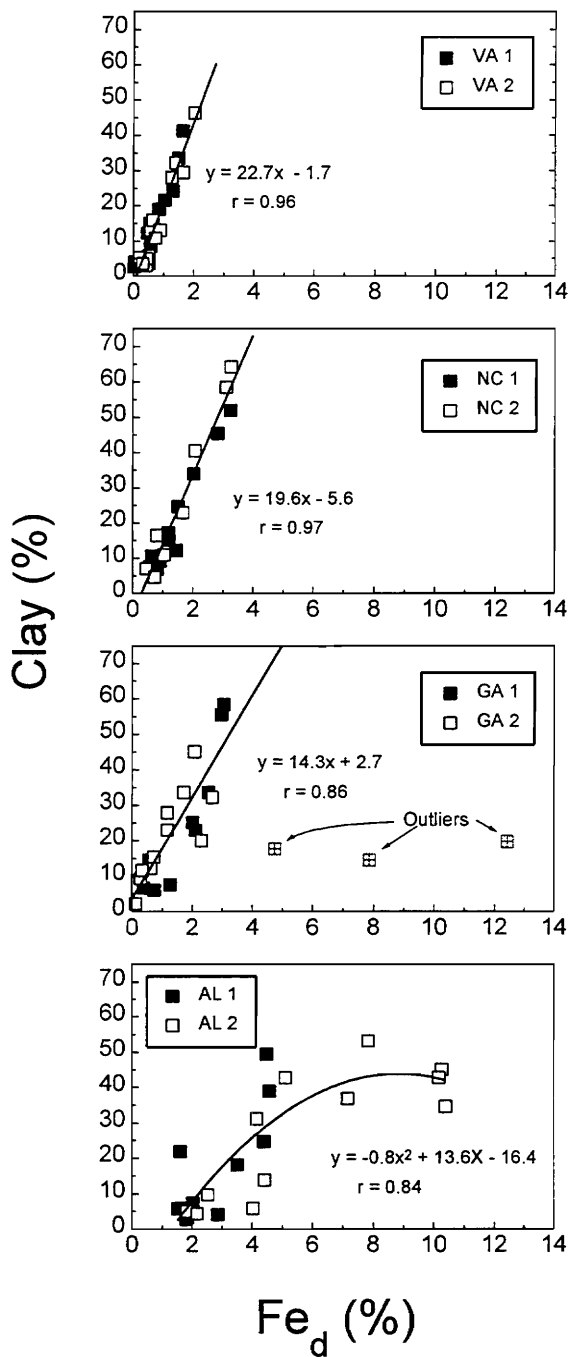


Fig. 5-4. Relationship between clay mass and Fe_d for each study site (a-d).

an abundance of hydroxy-interlayered vermiculite (HIV) which is highly selective for K. Alabama soils, however, are entirely depleted of exchangeable K, suggesting that HIV diminishes in clay fractions farther south. Exchangeable Ca is present at depths of <1 m in all soils, probably as a result of liming practices. The high levels of exchangeable Mg reported for pedon VA 1 by (Chapter III, this volume) are also present in pedon VA 2, but not elsewhere in the soils studied. Higher levels of Mg in Virginia may be indicative of Pleistocene climatic fluctuations and a greater annual cycling of bases and biomass production under what might have been grassland vegetation (Birkeland, 1984).

Cation exchange capacity values show little variation with latitude, with maxima ranging from ~8-16 $\text{cmol}_c \cdot \text{kg}^{-1}$ in surface or near-surface horizons, to ~2-6 $\text{cmol}_c \cdot \text{kg}^{-1}$ at depth. Higher CEC trends in surface horizons reflect the influence of high charged mica weathering products higher in the profile and the dominance of gibbsite and 1:1 phyllosilicates at depth.

Silt Mineralogy

The relative trends reported for silt mineralogy in Tables 5-2 (a-i) represent raw x-ray diffraction peak heights. In the case where silt minerals lack interstratification, poor crystallinity, or compositional variability that gives rise to the broad peaks characteristic of clay minerals, the use of peak heights is an acceptable substitute over the conventional peak area analysis.

Table 5-2a. Mineralogy of the VA 2 pedon silt (2-50 μ) fraction.

Pedon	Hor- izon	Depth	Verm- iculite †	Inter- stratified	Micas	Kaolins	Gibbsite	Quartz	Goethite	K-Spar	Plag- io- clase	XRD Peak Height (counts sec ⁻¹)	
												cm	(110) ‡
VA 2	A	0-13	--	--	10Å	~7.2Å	4.85Å	4.24Å	4.18Å	3.25Å	3.19Å	--	92
	BA	13-20	--	--	21	tr	--	319	--	53	--	--	53
	Bt1	20-30	--	--	26	50	--	231	tr	36	tr	tr	36
	Bt2	30-46	--	--	38	56	--	225	tr	38	tr	tr	38
	2Bt3b	46-91	--	--	52	53	--	91	tr	35	37	37	35
	2Bt4b	91-147	--	--	41	71	--	115	tr	40	38	38	40
	2BC1b	147-182	--	--	66	90	--	62	tr	36	38	38	36
	2BC2b	234-244	--	--	78	63	--	85	tr	38	38	38	38
	2C1b	284-295	--	--	93	83	--	79	tr	tr	39	39	tr
	2C2b	340-350	--	--	77	120	--	68	tr	36	44	44	36
	2C3b	411-421	--	--	59	116	--	64	tr	38	35	35	38
	2C4b	498-508	--	--	58	84	--	80	tr	38	49	49	38
	2C5b	615-625	--	--	65	156	--	101	tr	41	38	38	41
	2C6b	767-777	--	--	59	138	--	73	tr	113	41	41	113

† Samples are untreated powder mounts. Vermiculite includes all 14Å phases (e.g., hydroxy-interlayered vermiculite); interstratified includes all phases with a superstructure reflection at >20Å and/or ~12 Å; micas are dominantly muscovite, but can include some biotite; kaolins include both kaolinite and dehydrated halloysite; K-spars include K feldspars (dominantly microcline); plagioclase is dominantly albite.

‡ (hkl) reflection

§ Also includes a significant contribution from the K-spar (130) peak at 3.7Å.

Table 5-2b. Mineralogy of the VA 1 pedon medium silt (5-20 μ) fraction.

Pedon	Horizon	Depth	Vermiculite †	Inter-stratified	Micas	Kaolins	Gibbsite	Quartz	Goethite	K-Spar	Plagioclase
			14Å	12Å	10Å	~7.2Å	4.85Å	4.24Å	4.18Å (110) ‡	3.25Å	3.19Å
		cm	XRD Peak Height (counts sec ⁻¹)								
VA 1	A	0-8	NS§	--	--	--	--	--	--	--	--
	E	8-15	--	--	60	14	--	345	--	47	30
	BE	15-30	--	--	76	24	--	395	--	64	36
	Bt1	30-38	--	--	42	15	--	363	--	73	27
	Bt2	38-51	--	--	104	38	--	379	--	75	18
	2Bt3b	51-74	--	--	57	202	5	93	19	42	10
	2BCb	74-109	--	--	60	280	6	42	--	31	18
	2CBb	109-178	--	--	64	247	--	44	--	14	16
	2C1b	178-213	--	--	112	194	--	45	--	15	17
	2C2b	213+	--	--	51	134	--	50	--	11	16
	2C3b	244-264	--	--	184	231	--	38	--	17	25
	2C4b	284-295	--	--	179	225	--	32	--	11	23
	2C5b	401-411	--	--	261	188	--	35	29	--	20
	2C6b	498-508	--	--	395	177	77 (5%)	35	--	--	72
	2C7b	508-518	--	--	462	184	41 (3%)	38	21	--	58
	2C8b	528-564	--	--	436	195	32 (2%)	41	--	--	50
	2C9b	564+	--	--	312	131	4 (Tr)	36	--	--	29

† Samples are untreated powder mounts. Vermiculite includes all 14Å phases (e.g., hydroxy-interlayered vermiculite); interstratified includes all phases with a superstructure reflection at >20Å and/or ~12 Å; micas are dominantly muscovite, but can include some biotite; kaolins include both kaolinite and dehydrated halloysite; K-spars include K feldspars (dominantly microcline); plagioclase is dominantly albite.

‡ (hkl) reflection

§ Horizon not sampled.

† Bold numbers in parentheses are quantitative values for selected horizons obtained by using Differential Scanning Calorimetry (DSC) for the mineral indicated.

Table 5-2c. Mineralogy of the VA 1 pedon coarse silt (20-50 μ) fraction.

Pedon	Horizon	Depth	Vermiculite †	Inter-stratified	Micas	Kaolins	Gibbsite	Quartz	Goethite	K-Spar	Plagioclase	XRD Peak Height (counts sec ⁻¹)	
												cm	(110) ‡
VA 1	A	0-8	NS§	--	--	--	--	--	--	--	--	--	--
	E	8-15	--	tr	185	21	--	593	--	128	38	--	38
	BE	15-30	--	--	162	tr	--	422	--	242	98	--	98
	Bt1	30-38	--	--	129	10	--	412	--	121	tr	--	tr
	Bt2	38-51	--	--	179	35	--	388	--	81	tr	--	tr
	2Bt3b	51-74	--	tr	51	27	--	257	--	54	tr	--	tr
	2BCb	74-109	7	tr	71	101	--	104	18	94	19	--	19
	2CBb	109-178	--	--	94	100	--	66	16	83	21	--	21
	2C1b	178-213	--	--	79	88	--	97	14	35	16	--	16
	2C2b	213+	19	--	270	136	3	49	--	35	24	--	24
	2C3b	244-264	--	--	189	96	5	38	--	21	38	--	38
	2C4b	284-295	--	--	609	183	--	99	19	44	29	--	29
	2C5b	401-411	--	--	696	167	--	39	--	--	48	--	48
	2C6b	498-508	--	--	272	74	29	41	--	--	200	--	200
	2C7b	508-518	--	--	452	115	26	tr	25	--	232	--	232
	2C8b	528-564	--	16	288	59	17	29	--	--	105	--	105
	2C9b	564+	--	--	878	139	--	22	--	--	147	--	147

† Samples are untreated powder mounts. Vermiculite includes all 14Å phases (e.g., hydroxy-interlayered vermiculite); interstratified includes all phases with a superstructure reflection at >20Å and/or ~12 Å; micas are dominantly muscovite, but can include some biotite; kaolins include both kaolinite and dehydrated halloysite; K-spars include K feldspars (dominantly microcline); plagioclase is dominantly albite.

‡ (hkl) reflection

§ Horizon not sampled.

Table 5-2d. Mineralogy of the NC 1 pedon silt (2-50 μ) fraction.

Pedon	Hor- izon	Depth	Verm- iculite †	Inter- stratified	Micas	Kaolins	Gibbsite	Quartz	Goethite	K-Spar	Plag- io- clase	XRD Peak Height (counts sec ⁻¹)	
												14Å	12Å
NC 1	Ap	0-15	--	--	--	13	--	404	--	137	--	--	
	BA	15-28	--	--	--	21	--	398	--	104	--	--	
	Bt1	28-48	--	--	--	55	--	323	--	82	--	--	
	Bt2	48-74	--	--	--	69	--	282	--	82	--	--	
	BC1	74-102	--	--	--	73	--	238	--	119	--	--	
	BC2	102-165	--	--	tr	83	--	192	--	139	--	--	
	C1	165-265	--	--	tr	82	--	264	tr	131	--	--	
	C2	265-295	--	--	--	30	--	173	--	121	--	--	
	C3	356-381	--	--	tr	79	--	218	tr	147	--	--	
	C4	457-483	--	--	tr	63	--	238	tr	106	--	--	
	C5	508-533	--	--	--	71	--	189	tr	173	--	--	

† Samples are untreated powder mounts. Vermiculite includes all 14Å phases (e.g., hydroxy-interlayered vermiculite); interstratified includes all phases with a superstructure reflection at >20Å and/or ~12 Å; micas are dominantly muscovite, but can include some biotite; kaolins include both kaolinite and dehydrated halloysite; K-spars include K feldspars (dominantly microcline); plagioclase is dominantly albite.

‡ (hkl) reflection

§ Also includes a significant contribution from the K-spar (130) peak at 3.7Å.

Table 5-2e. Mineralogy of the NC 2 pedon silt (2-50 μ) fraction.

Pedon	Hor- izon	Depth	Verm- culite †	Inter- stratified	Micas	Kaolins	Gibbsite	Quartz	Goethite	K-Spar	Plag- io- clase	XRD Peak Height (counts sec ⁻¹)	
												Goethite	K-Spar
NC 2	Ap	0-13	--	--	16	15	--	487	--	33	--		
	BA	13-20	--	--	14	14	--	522	--	24	tr		
	Bt1	20-41	--	--	17	30	--	386	--	33	--		
	Bt2	41-56	--	--	18	58	tr	256	--	26	--		
	Bt3	56-122	--	--	tr	109	tr	214	tr	26	--		
	BC1	122-183	--	--	tr	79	--	161	tr	93	--		
	BC2	183-208	--	--	tr	134	--	111	tr	45	--		
	BC3	208-244	--	--	tr	81	--	129	tr	116	--		

† Samples are untreated powder mounts. Vermiculite includes all 14Å phases (e.g., hydroxy-interlayered vermiculite); interstratified includes all phases with a superstructure reflection at >20Å and/or ~12 Å; micas are dominantly muscovite, but can include some biotite; kaolins include both kaolinite and dehydrated halloysite; K-spars include K feldspars (dominantly microcline); plagioclase is dominantly albite.

‡ (hkl) reflection

§ Also includes a significant contribution from the K-spar (130) peak at 3.7Å.

Table 5-2f. Mineralogy of the GA 1 pedon silt (2-50 μ) fraction.

Pedon	Hor- izon	Depth	Verm- cullite †	Inter- stratified	Micas	Kaolins	Gibbsite	Quartz	Goethite	K-Spar	Plag- io- clase
			14Å	12Å	10Å	~7.2Å	4.85Å	4.24Å	4.18Å (110) ‡	3.25Å	3.19Å
			XRD Peak Height (counts sec ⁻¹)								
cm											
GA 1	Ap	0-10	--	--	--	23	--	285	--	137	--
	BA	10-25	--	--	--	25	--	267	--	62	--
	Bt	25-61	--	--	26	193	34	173	--	139	--
	BC	61-109	--	--	tr	281	42	129	--	31	--
	CB1	109-157	--	--	--	216	tr	77	76	tr	--
	CB2	157-201	--	--	tr	247	--	84	81	--	--
	CB3	201-254	--	--	--	217	--	90	82	38	--
	C1	307-323	--	--	--	153	--	69	60	--	--
	C2	370-386	--	--	--	137	--	65	78	--	--
	C3	434-450	--	--	17	92	--	55	79	--	--
	C4	497-513	--	--	--	188	--	84	84	45	--
	C5	536-549	--	--	--	147	--	88	tr	77	--

† Samples are untreated powder mounts. Vermiculite includes all 14Å phases (e.g., hydroxy-interlayered vermiculite); interstratified includes all phases with a superstructure reflection at >20Å and/or ~12 Å; micas are dominantly muscovite, but can include some biotite; kaolins include both kaolinite and dehydrated halloysite; K-spars include K feldspars (dominantly microcline); plagioclase is dominantly albite.

‡ (hkl) reflection

§ Also includes a significant contribution from the K-spar (130) peak at 3.7Å.

Table 5-2g. Mineralogy of the GA 2 pedon silt (2-50 μ) fraction.

Pedon	Hor- izon	Depth	Verm- iculite †	Inter- stratified	Micas	Kaolins	Gibbsite	Quartz	Goethite	K-Spar	Plag- io- clase	XRD Peak Height (counts sec ⁻¹)	
												cm	(110) ‡
GA 2	Ap	0-8	--	--	12	56	53	147	--	201	--		
	BA	8-18	--	--	25	51	24	186	--	270	--		
	Bt1	18-30	--	--	21	127	15	128	tr	113	--		
	Bt2	30-69	--	--	46	312	tr	81	tr	82	--		
	BC1	69-152	--	--	38	252	tr	79	27	81	--		
	BC2	152-229	--	20	14	77	tr	51	103	70	tr		
	CB	229-240	--	--	15	260	--	90	84	40	tr		
	C1	254-270	20	50	20	119	--	109	tr	163	tr		
	C2	318-336	16	17	18	153	--	62	27	117	tr		
	C3	368-384	13	37	22	86	--	114	36	178	tr		
	C4	457-473	--	21	26	34	246	61	14	228	tr	10	

† Samples are untreated powder mounts. Vermiculite includes all 14Å phases (e.g., hydroxy-interlayered vermiculite); interstratified includes all phases with a superstructure reflection at >20Å and/or ~12 Å; micas are dominantly muscovite, but can include some biotite; kaolins include both kaolinite and dehydrated halloysite; K-spars include K feldspars (dominantly microcline); plagioclase is dominantly albite.

‡ (hkl) reflection

§ Also includes a significant contribution from the K-spar (130) peak at 3.7Å.

Table 5-2h. Mineralogy of the AL 1 pedon silt (2-50 μ) fraction.

Pedon	Hor- izon	Depth	Verm- iculite †	Inter- stratified	Micas	Kaolins	Gibbsite	Quartz	Goethite	K-Spar	Plag- io- clase	XRD Peak Height (counts sec ⁻¹)	
												(110) ‡	(hkl) §
AL 1	Ap	0-13	--	--	10Å	~7.2Å	4.85Å	4.24Å	4.18Å	3.25Å	3.19Å	--	31
	Bt1	13-60	--	--	46	157	tr	108	tr	--	15	--	15
	Bt2	60-100	--	tr	45	116	45	175	40	--	tr	--	tr
	BC1	100-140	--	tr	38	153	50	93	52	--	10	--	10
	BC2	140-228	--	tr	44	134	36	92	52	--	12	--	12
	CB	228-270	--	tr	34	139	51	117	58	--	tr	--	tr
	C1	351-367	--	26	32	130	tr	64	57	--	tr	--	tr
	C2	384-400	--	21	36	147	12	76	69	--	tr	--	tr
	C3	462-478	--	20	38	97	20	54	55	--	tr	--	tr
	C4	502-518	--	18	39	121	21	62	56	--	15	--	15
	C5	549-565	tr	24	55	116	22	101	27	--	17	--	17

† Samples are untreated powder mounts. Vermiculite includes all 14Å phases (e.g., hydroxy-interlayered vermiculite); interstratified includes all phases with a superstructure reflection at >20Å and/or ~12 Å; micas are dominantly muscovite, but can include some biotite; kaolins include both kaolinite and dehydrated halloysite; K-spars include K feldspars (dominantly microcline); plagioclase is dominantly albite.

‡ (hkl) reflection

§ Also includes a significant contribution from the K-spar (130) peak at 3.7Å.

Table 5-2i. Mineralogy of the AL 2 pedon silt (2-50 μ) fraction.

Pedon	Hor- izon	Depth	Verm- cullite †	Inter- stratified	Micas	Kaolins	Gibbsite	Quartz	Goethite	K-Spar	Plag- io- clase	XRD Peak Height (counts sec ⁻¹)	
												cm	
AL 2	Ap	0-10	--	--	17	tr	32	288	--	47	--		
	Bt1	10-55	--	--	20	tr	33	305	--	21	--		
	Bt2	55-108	--	--	23	48	117	97	--	--	--		
	Bt3	108-155	--	tr	10	42	62	46	--	--	--		
	Bt4	155-170	--	tr	20	62	tr	41	83	--	--		
	Bt5	170-228	--	tr	tr	43	tr	40	85	--	--		
	BC	228-260	--	34	22	68	57	45	55	--	--		
	CB1	307-323	tr	33	23	34	325	49	35	--	33		
	CB2	368-384	tr	43	21	73	44	55	50	--	35		
	C1	429-445	tr	23	tr	25	477	100	tr	--	46		
	C2	454-470	tr	32	17	25	524	133	tr	--	45		
	C3	470-486	tr	22	23	35	509	tr	30	--	50		

† Samples are untreated powder mounts. Vermiculite includes all 14Å phases (e.g., hydroxy-interlayered vermiculite); interstratified includes all phases with a superstructure reflection at >20Å and/or ~12 Å; micas are dominantly muscovite, but can include some biotite; kaolins include both kaolinite and dehydrated halloysite; K-spars include K feldspars (dominantly microcline); plagioclase is dominantly albite.

‡ (hkl) reflection

§ Also includes a significant contribution from the K-spar (130) peak at 3.7Å.

Silt mineralogy of all soils shows typical weathering patterns for residuum, with quartz dominating upper horizons, both mica and feldspars increasing in abundance with depth, and the secondary kaolinite and goethite minerals occupying the solum and upper saprolite. With the exception of pedon VA 2 and NC 1, gibbsite is present in all soils, in one or both of two distinct layers. The first one occurs at the top of the profile near the surface. The second one is located toward the bottom of the profile, close to unweathered parent rocks.

The increase in gibbsite in surface horizon has been termed 'bauxitic gibbsite' by Tardy (1992) and is the result of desilication of kaolinite under stable landscape conditions giving rise to long-term weathering, and intense leaching. There is only minimal evidence of gibbsite formation by this mechanism in Virginia and North Carolina soils (Table 5-2, a-c), suggesting that soils there are either younger or not as highly leached as soils farther south.

Gibbsite also forms deeper in the saprolite at the expense of Na-plagioclase alteration (Chapter IV, this volume). Pedons GA 2, AL 1, and AL 2 show substantial amounts of gibbsite formed by this mechanism at depths >3 m relative to amounts in overlying horizons (Table 5-2, f-h). Pedon VA 1 has only pseudomorphic gibbsite after feldspar, where amounts range from 2-5% of the silt fraction. Pedon GA 1 shows only desilication gibbsite beginning at a depth of 25 cm instead of at the surface. This suggests that some type of surface disturbance may have occurred in the past. The North Carolina soils, because of the relatively moderate climate and lack of plagioclase in the

parent material at the sampling locations, have no gibbsite whatsoever. The distribution of kaolinite minerals in all soils is also somewhat bimodal in nature, with dehydrated halloysite forming in lower saprolite horizons as a pseudomorphic replacement of Na-plagioclase, and also in subsurface horizons, where kaolinite forms both from the recrystallization of halloysite and the desilication of 2:1 phyllosilicates. There are no significant trends in kaolin content with respect to latitude or weathering intensity.

Virginia and North Carolina soils contain no measurable silt-sized alteration products of mica. Farther south, pedons GA 2, AL 1, and AL 2 have appreciable amounts of interstratified 2:1 species and trace amounts of vermiculite in lower horizon silt fractions. This also indicates that soils in GA and AL are older or more highly leached than soils in VA and NC.

Goethite is present in small amounts in the silt fraction of all soils, usually in greatest abundance near the contact between saprolite and the lower solum.

Clay Mineralogy

The dominant weathering products and processes described for deep soil/saprolite profiles on the northern Virginia Piedmont (Chapters III and IV, this volume) were also observed in North Carolina, Georgia, and Alabama. With few minor exceptions, most profiles displayed three distinct zones of weathering: 1) a highly weathered upper zone which is dominated by low activity clays that result from conditions of intense desilication and high Al activity, 2) a lower, unoxidized saprolite zone which is usually

>5 m in depth and close to unweathered rock, and 3) an intermediate zone where initial secondary weathering products are transformed to more stable species.

Dominant weathering mechanisms in zone 1 include interlayering of hydroxy-Al species in vermiculite to form HIV in surficial horizons, and desilication of phyllosilicates to form gibbsite in surface and kaolinite subsurface horizons. Zone 3 is characterized by the formation of both gibbsite (4.85Å) and poorly crystalline dehydrated halloysite (7.2Å) as pseudomorphic alteration products of feldspars and biotite. In the intermediate zone (Zone 2), dominant weathering processes include formation of kaolinite by replacement of halloysite and by resilication of gibbsite, and alteration of primary micas to form vermiculitic or smectite phases. Further details regarding these weathering mechanisms can be found in Chapters III and IV of this volume.

The trends of increased weathering intensity and desilication toward the south are even more strongly indicated by soil clay mineralogy than the soil properties discussed in previous sections. Bauxitic, or desilication gibbsite increases measurably in surface horizons from north to south, ranging from <1% in Virginia and 2% in North Carolina soils, to between 4% and 10% of the clay fraction of Georgia and Alabama soils, respectively (Fig. 5-5, a-h). The 14Å/7Å peak ratio, indicative of the relative proportion of HIV to kaolinite in these soils, also decreases from unity, or near unity, in Virginia and North Carolina (Fig. 5-5b,c,f,g,h), to well below unity in Georgia and Alabama. Kaolinite content of the clay fraction ranges from 40-60% in Virginia and North Carolina sola, to between 70-85% of the sola in Georgia and Alabama soils. Mica (10Å) and quartz

Pedon VA1 Clay Fraction

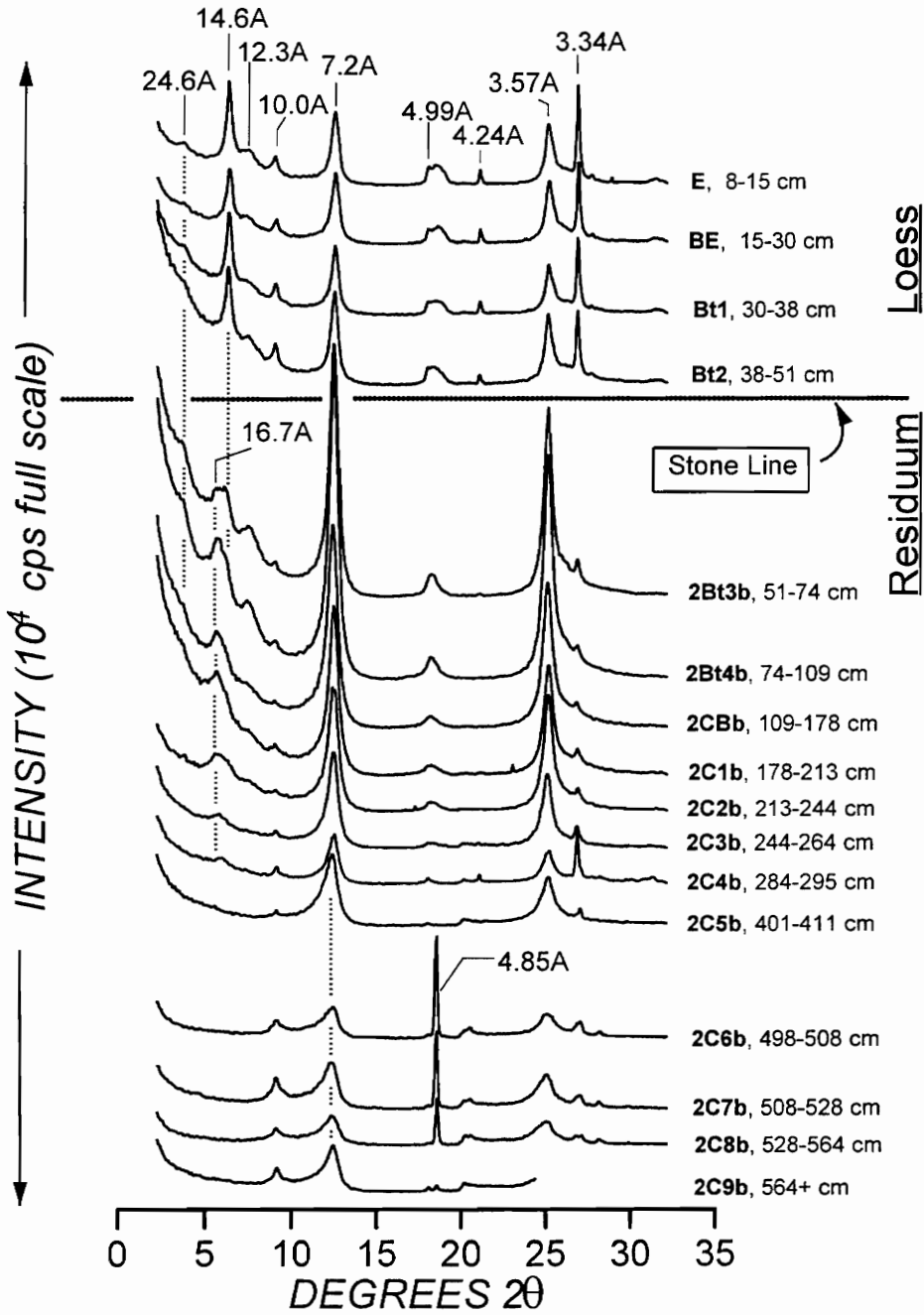


Fig. 5-5(a). X-ray diffractograms of oriented clay fractions from pedon VA 1. Samples were Mg-saturated, glycerol-solvated, and scanned at 25°C.

Pedon VA 2 Clay Fraction

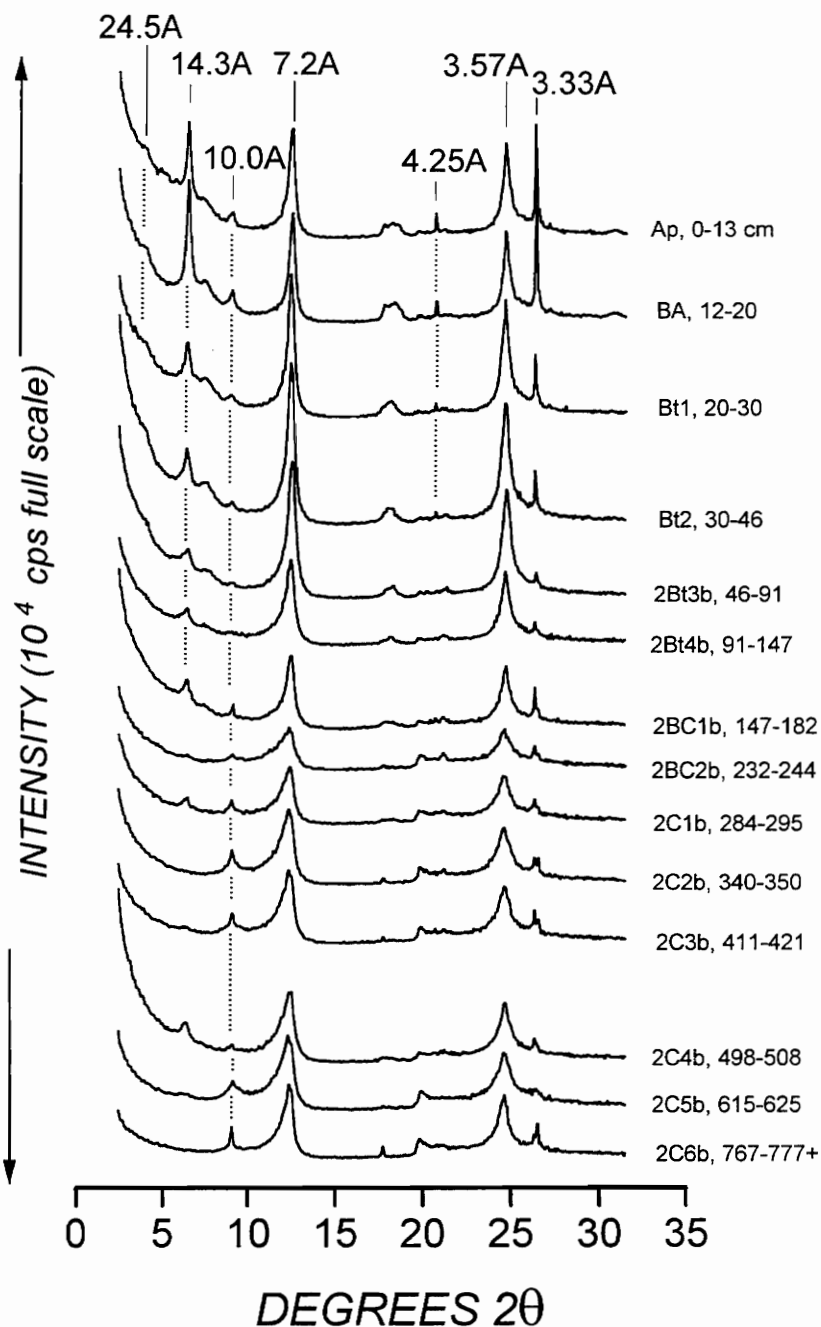


Fig. 5-5(b). X-ray diffractograms of oriented clay fractions from pedon VA 2. Samples were Mg-saturated, glycerol-solvated, and scanned at 25°C.

Pedon NC 1 Clay Fraction

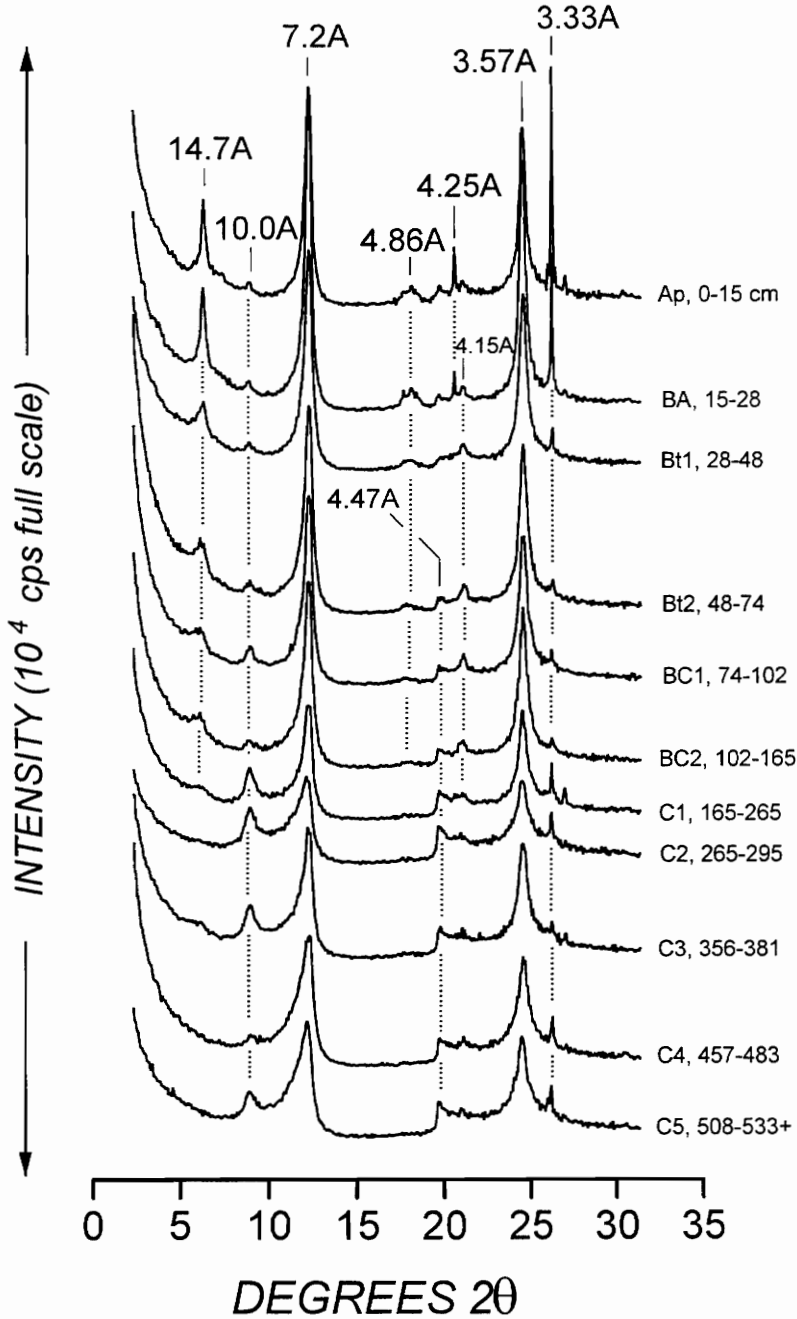


Fig. 5-5(c). X-ray diffractograms of oriented clay fractions from pedon NC 1. Samples were Mg-saturated, glycerol-solvated, and scanned at 25°C.

Pedon NC 2 Clay Fraction

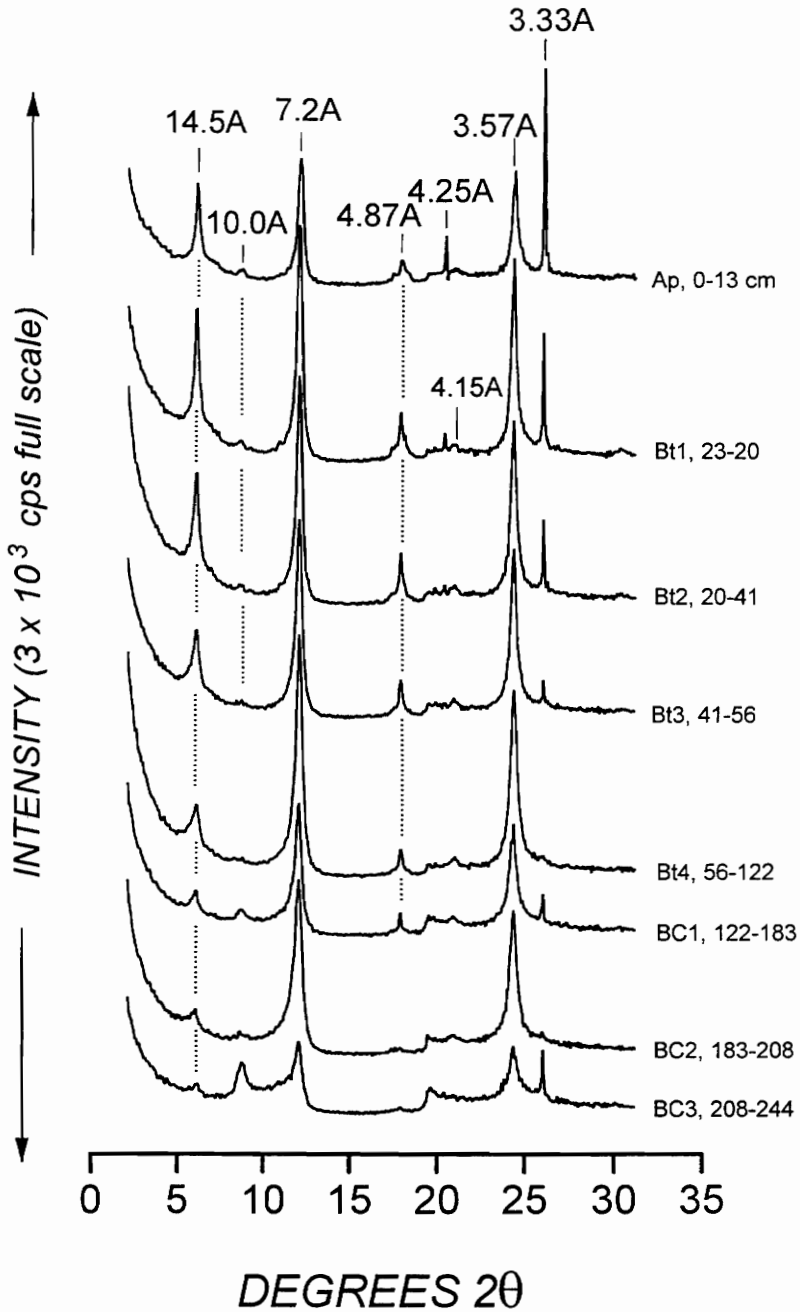


Fig. 5-5(d). X-ray diffractograms of oriented clay fractions from pedon NC 2. Samples were Mg-saturated, glycerol-solvated, and scanned at 25°C.

Pedon GA 1 Clay Fraction

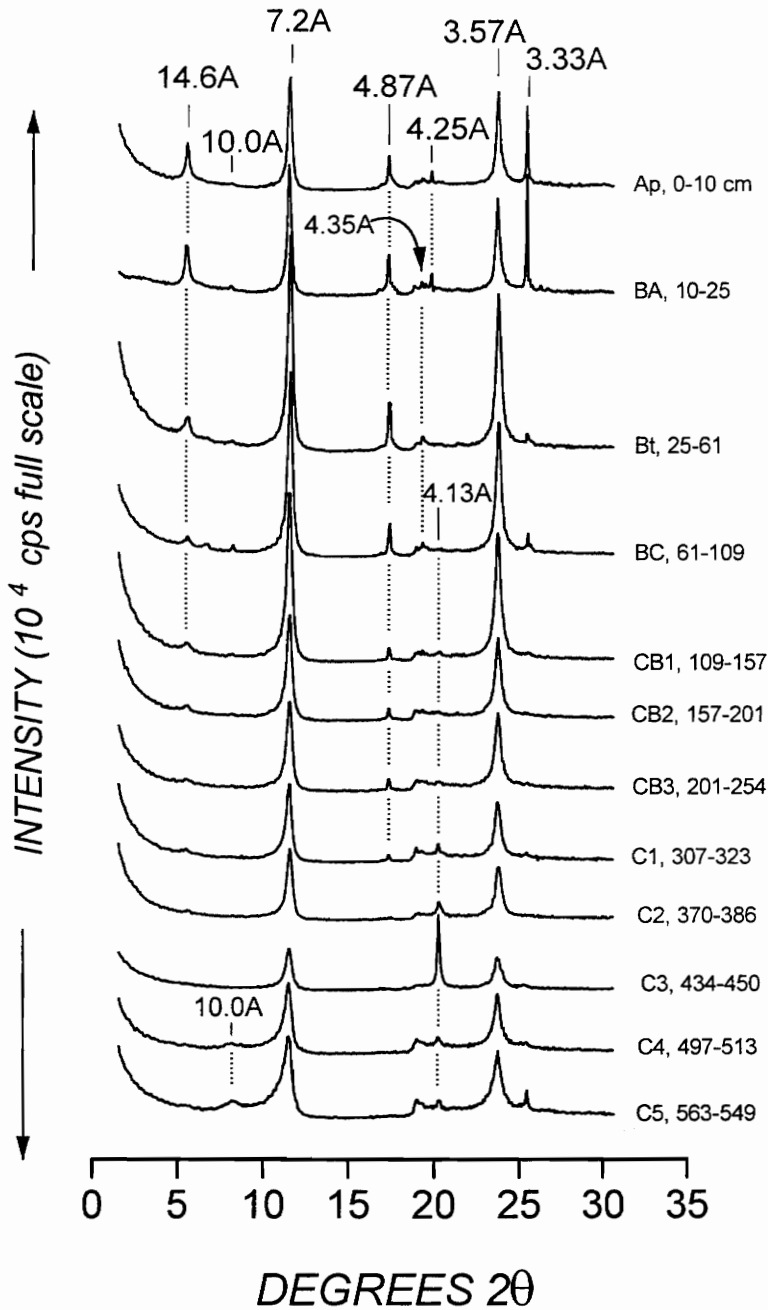


Fig. 5-5(e). X-ray diffractograms of oriented clay fractions from pedon GA 1. Samples were Mg-saturated, glycerol-solvated, and scanned at 25°C.

Pedon GA 2 Clay Fraction

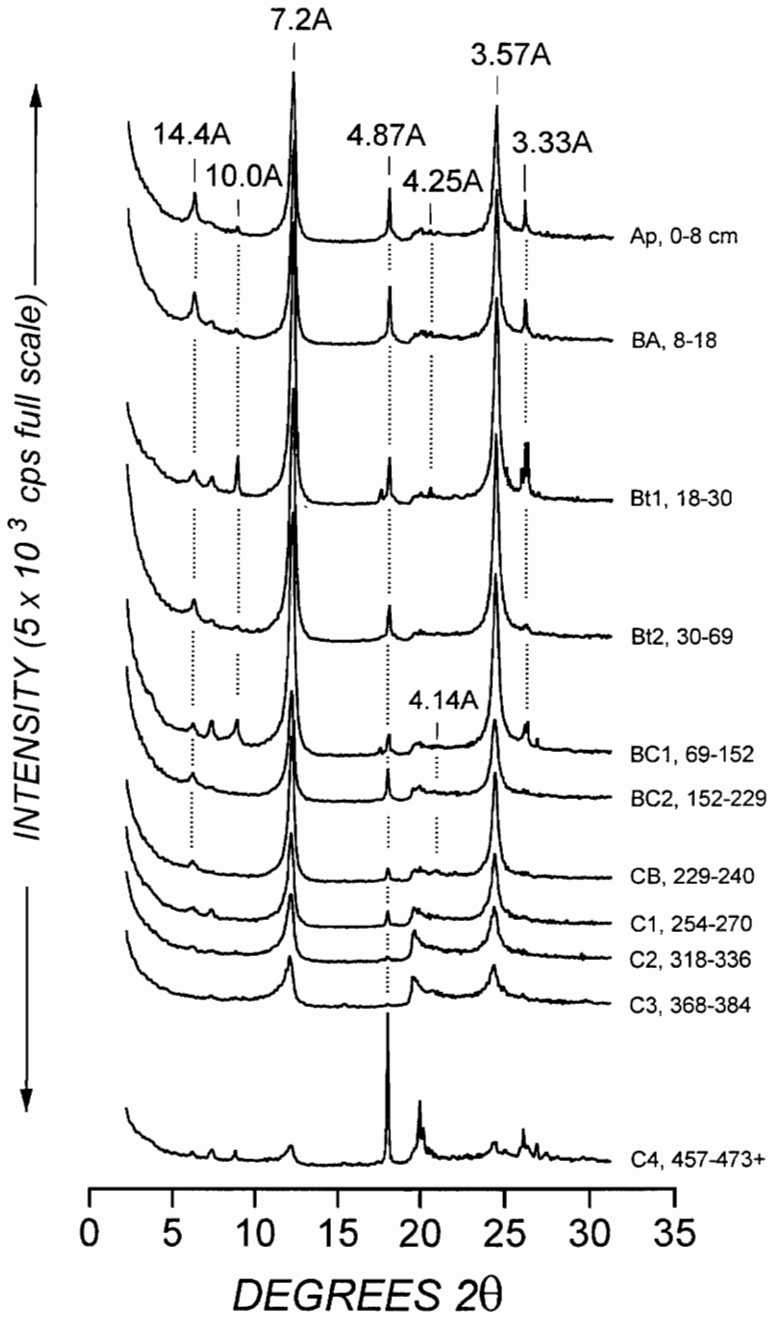


Fig. 5-5(f). X-ray diffractograms of oriented clay fractions from pedon GA 2. Samples were Mg-saturated, glycerol-solvated, and scanned at 25°C.

Pedon AL 1 Clay Fraction

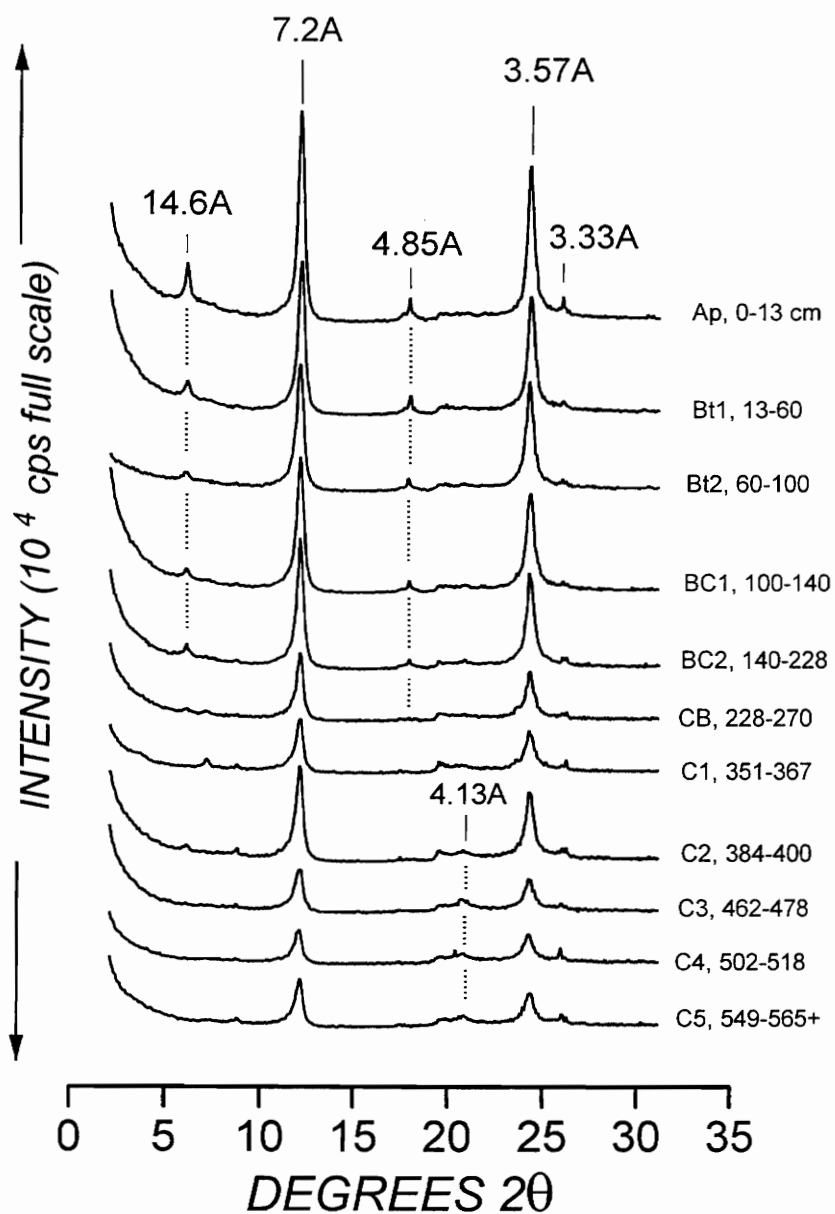


Fig. 5-5(g). X-ray diffractograms of oriented clay fractions from pedon AL 1. Samples were Mg-saturated, glycerol-solvated, and scanned at 25°C.

Pedon AL 2 Clay Fraction

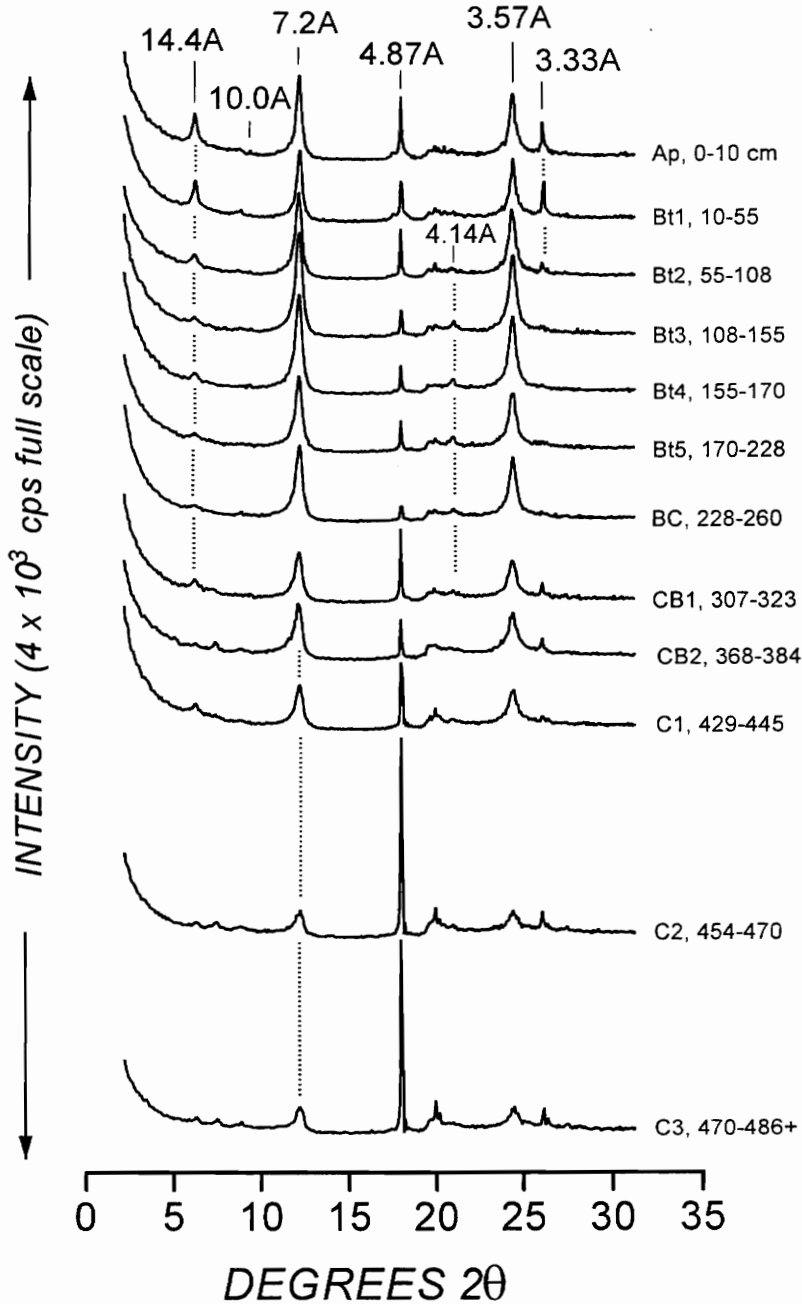


Fig. 5-5(h). X-ray diffractograms of oriented clay fractions from pedon AL 2. Samples were Mg-saturated, glycerol-solvated, and scanned at 25°C.

(4.24Å and 3.33Å) are present in only trace amounts in Georgia and Alabama in contrast to the northern sites. Goethite is present in subsurface horizons of North Carolina soils, and deeper in the profile in soils of Georgia and Alabama. Because of interferences from other mineral phases present, hematite trends require special techniques for measurement.

In all profiles, the depth at which micas begin to weather is closely related to the depth at which halloysite becomes the dominant phase in the clay fraction. This is shown by the decrease of the HIV 14Å peak downward in the profile, which coincides with the broadening of the 7.2Å peak, and the appearance of the 4.47Å (11 l and 02 l) peaks. This critical depth increases from ~200 cm in Virginia and North Carolina to >350 cm in Georgia and Alabama, suggesting that the effective depth of leaching and desilication is greater farther south.

Whole-Soil Elemental Analysis

Of the eight soils studied, only pedons VA 2, NC 1, GA 1, and AL 1 were analyzed for total elemental composition. As expected, much of the loss of mass associated with the formation of soil from saprolite was attributable to the leaching of soluble Si from the solum of all profiles (Fig. 5-6a). Insoluble elements (Al, Fe, and Ti) were conserved in these horizons (Fig. 5-6a,c,d). Similar trends are observed in SiO₂/Al₂O₃ and Fe₂O₃+Al₂O₃/SiO₂ ratios (Fig. 5-6b). Pedons GA 1 and AL 1 showed anomolous Si enrichment and Al depletion at depths of 157 cm and 228 cm, respectively, suggesting that these zones may represent former erosional surfaces. These pedons also show a

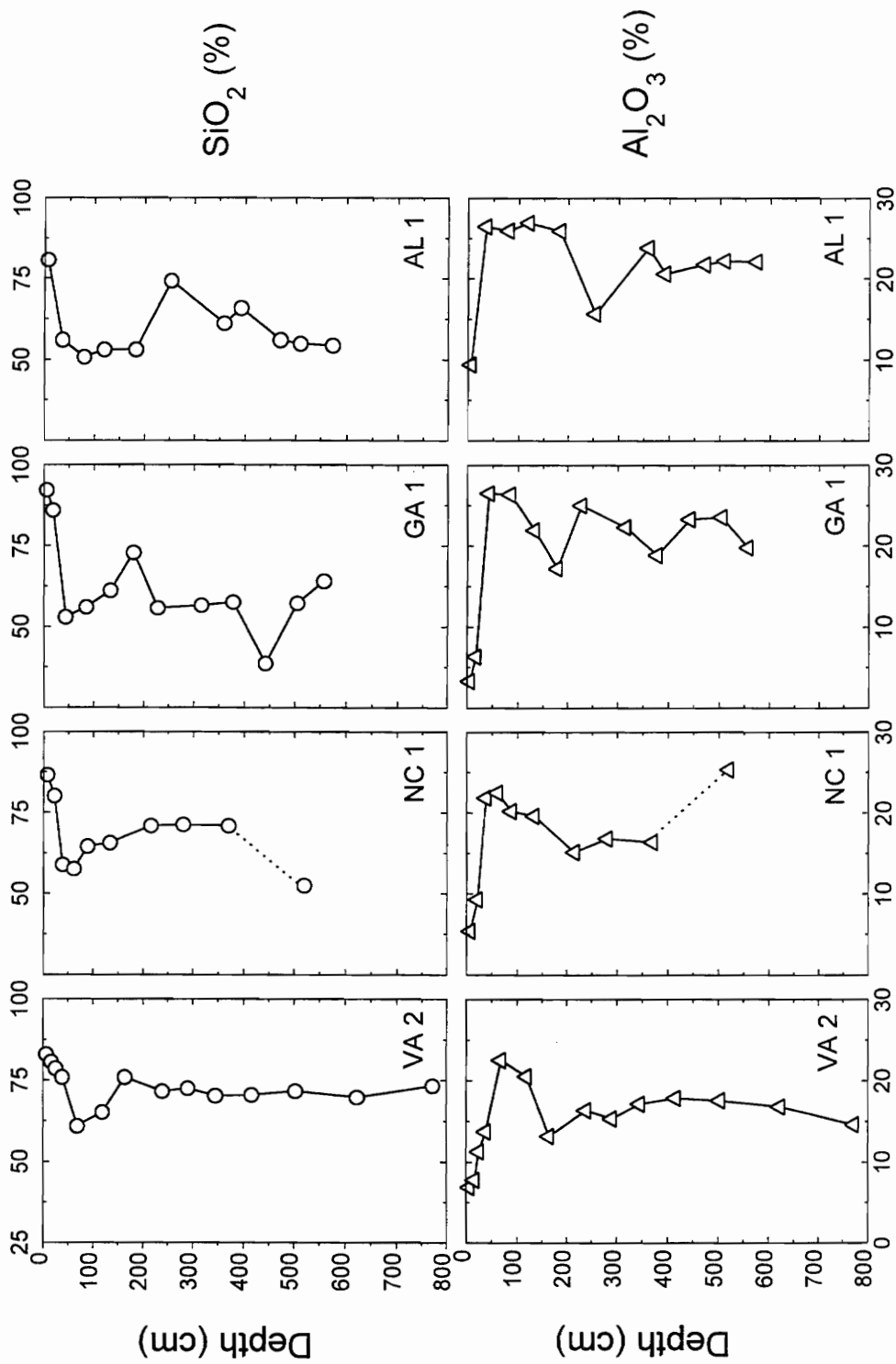


Fig. 5-6(a). Depth distributions of whole-soil (<2.0 mm) elemental properties for selected pedons.

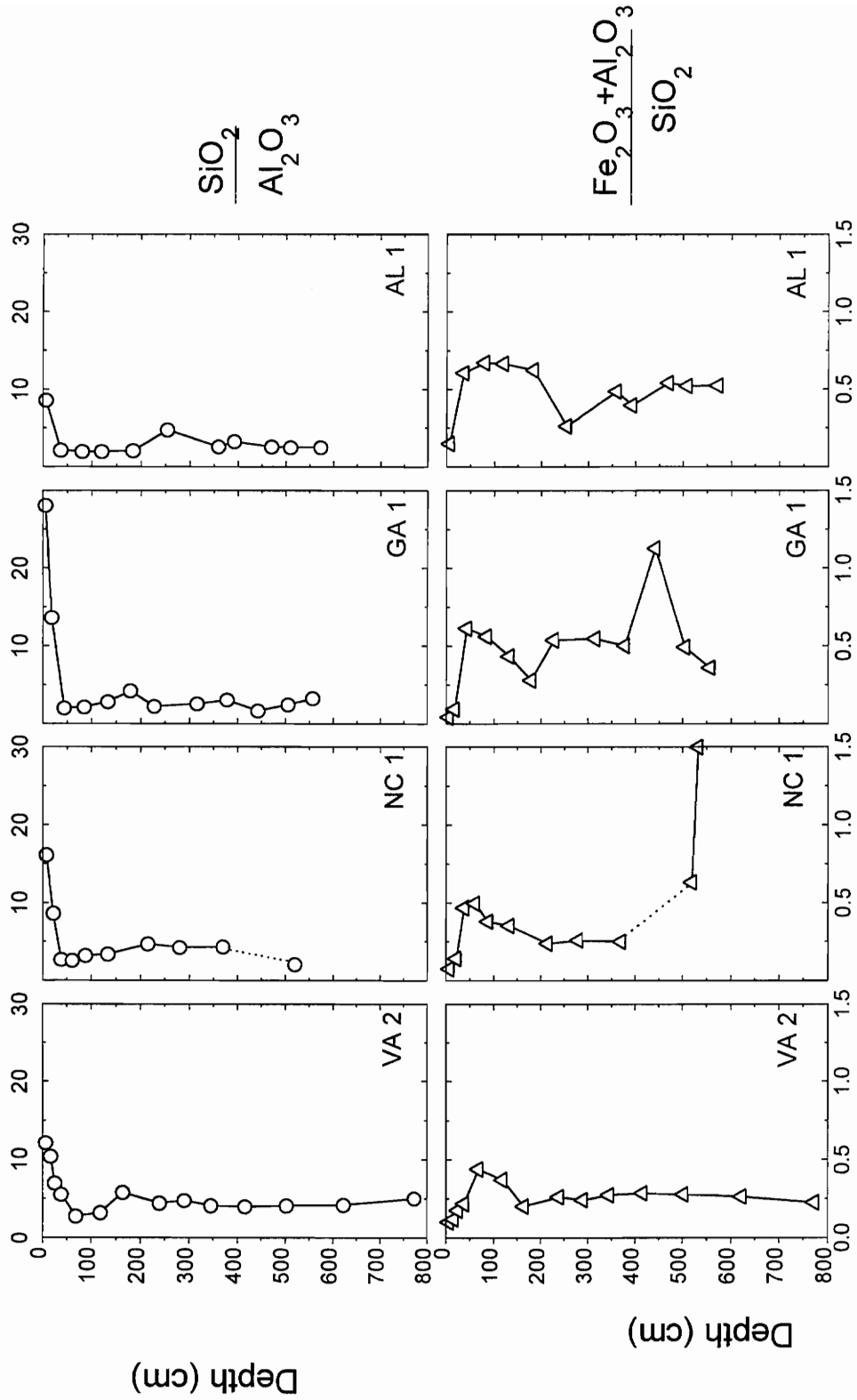


Fig. 5-6(b). Depth distributions of whole-soil (<2.0 mm) elemental properties for selected pedons.

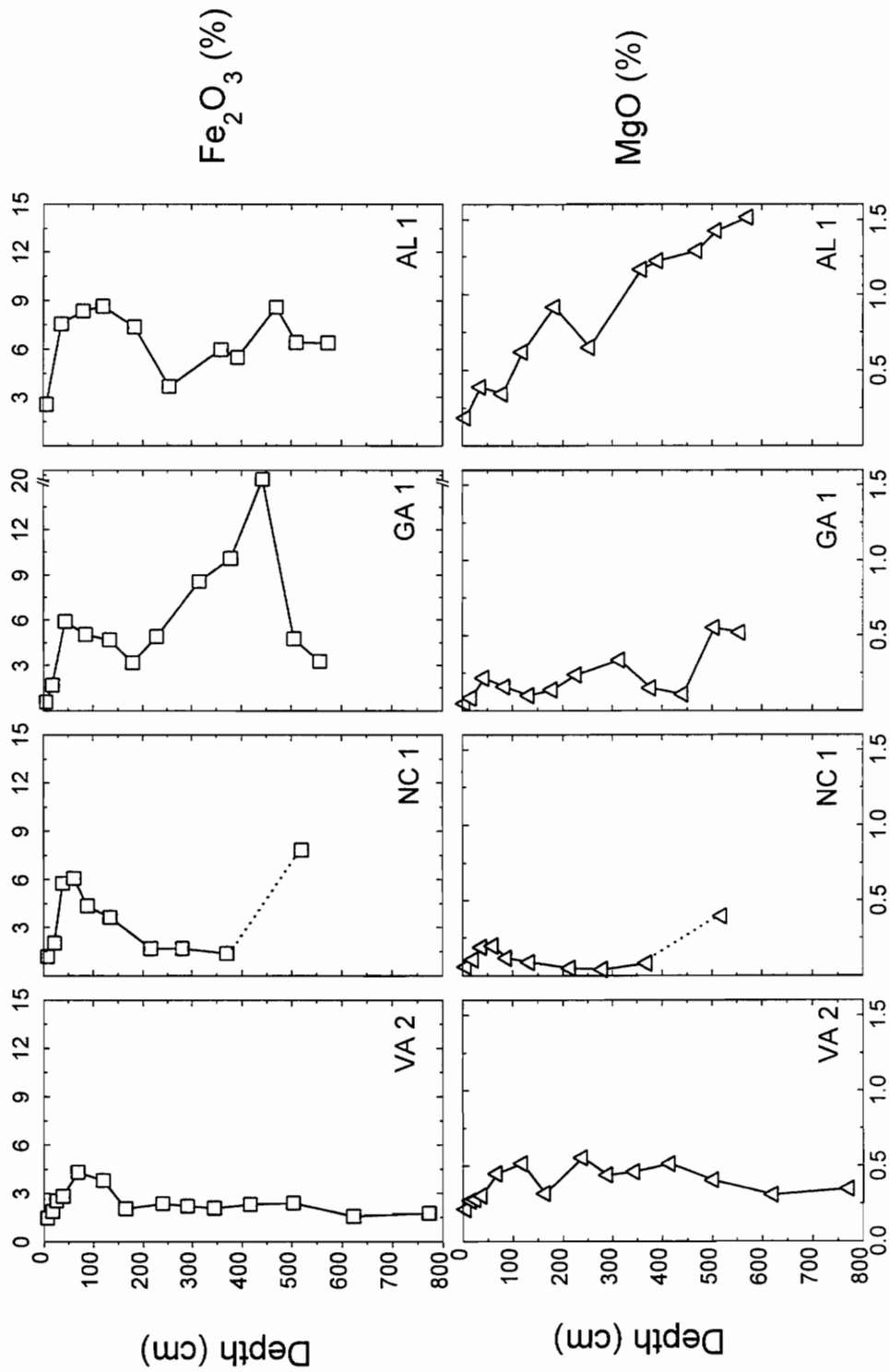


Fig. 5-6(c). Depth distributions of whole-soil (<2.0 mm) elemental properties for selected pedons.

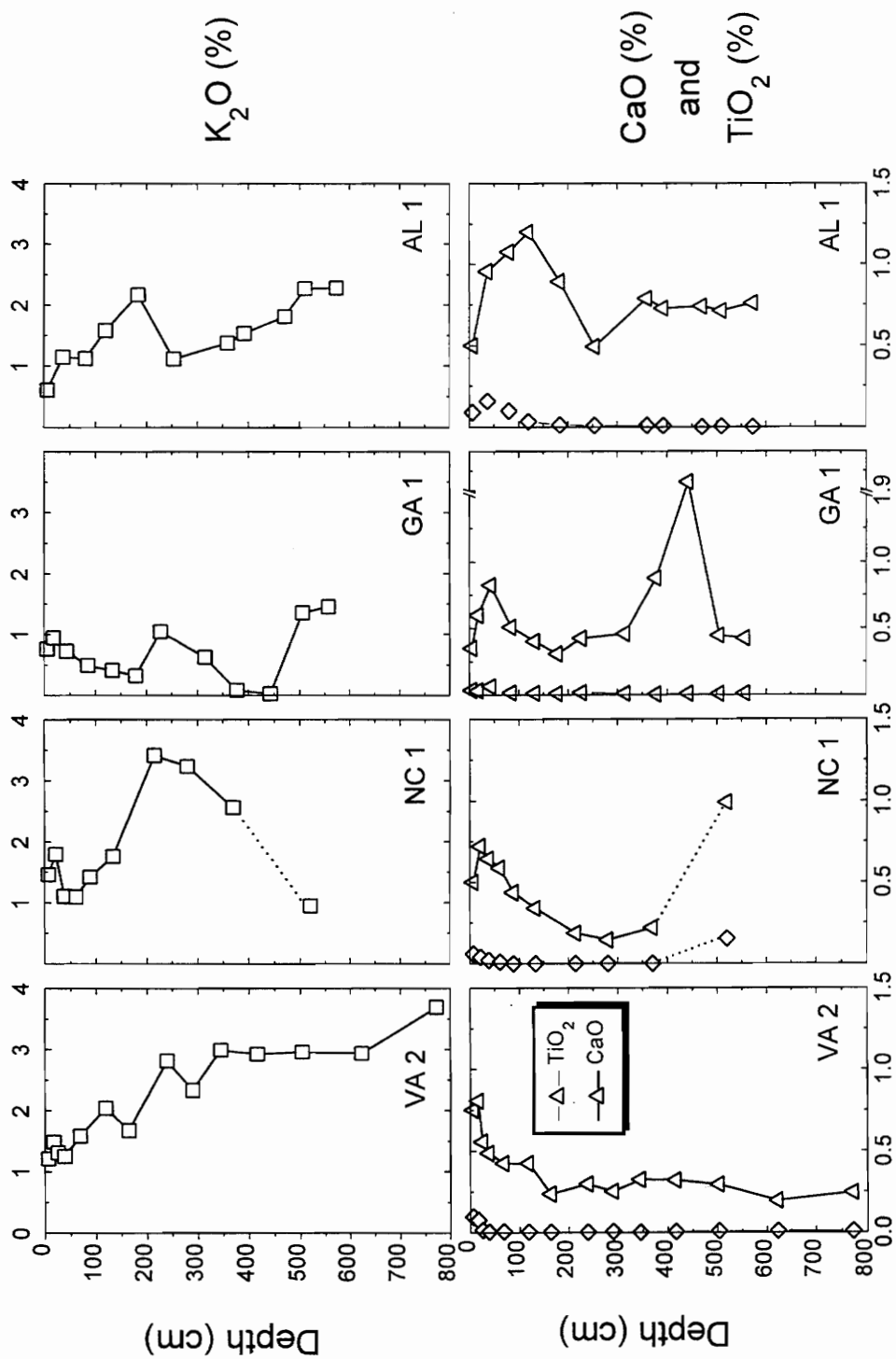


Fig. 5-6(d). Depth distributions of whole-soil (<2.0 mm) elemental properties for selected pedons.

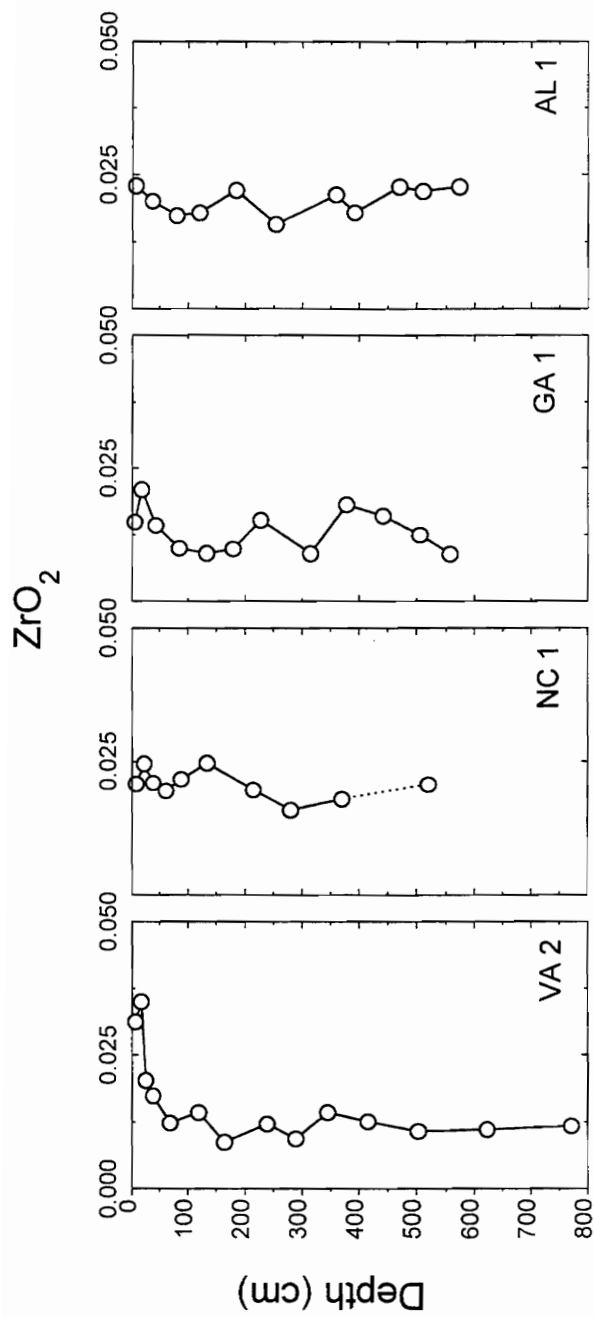


Fig. 5-6(e). Depth distributions of whole-soil (<2.0 mm) elemental properties for selected pedons.

relative depletion of CaO, K₂O, MgO, and ZrO₂ (Fig. 5-6c,d,e) at these same depths that is difficult to explain on the basis of geologic heterogeneity alone. Pedons VA 2, GA 1, and AL 1 have relatively large amounts of elemental MgO (Fig. 5-6c), but only VA 2 has any appreciable exchangeable Mg present (Fig. 5-2a,c,d), indicating that GA 1 and AL 1 have been highly leached. Potassium shows a decreasing upward trend in conjunction with mica depletion. Calcium is concentrated in clayey argillic horizons after the probable enrichment in surface horizons by liming practices (Fig. 5-6d). With the exception of some variability in pedons GA 1 and AL 1, ZrO₂ trends are relatively uniform with depth (Fig. 5-6e).

Selective Dissolution Analysis

Free Fe-oxides, including well-crystalline phases such as goethite and hematite, are very similar between replicate pedons at each site, and increase markedly from north to south. Maximum Fe_d values for whole-soil (<2 mm) samples ranged from ~1.6-2.0% in Virginia soils to ~4.5-10.4% in Alabama (Fig. 5-7) with depth distributions in all soils closely paralleling those of clay content (Fig. 5-2a-d, 5-3a). Values of Al_d have similar distributions as Fe_d trends, but they are about an order-of-magnitude lower, suggesting that Al_d comes from Al substituted in soil Fe-oxides.

Clay fraction Fe_d for four solum horizons of each pedon showed no trends in Virginia, North Carolina, and Georgia, but exponential growth in Alabama (Fig. 5-8). This may be more the result of the mafic character of the Elkahatchee Formation than any

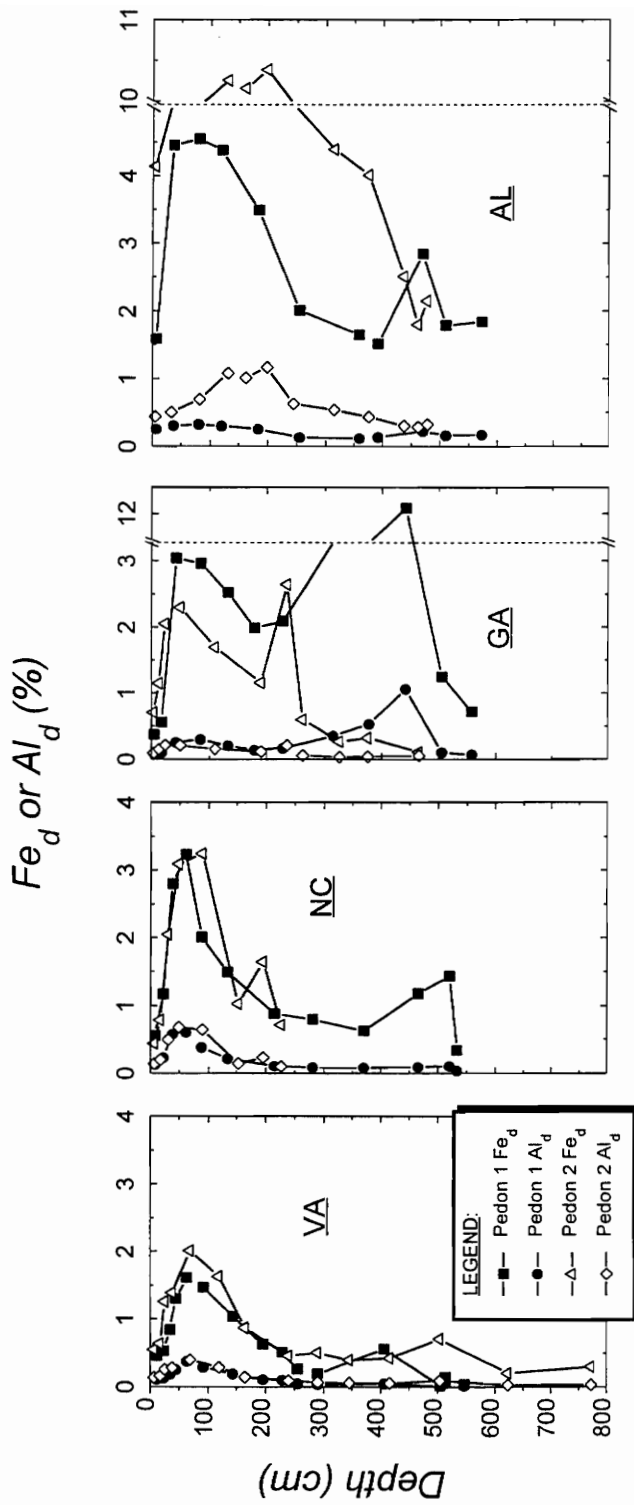


Fig. 5-7. Depth functions of whole-soil (<2 mm) Fe_d and Al_d for all pedons.

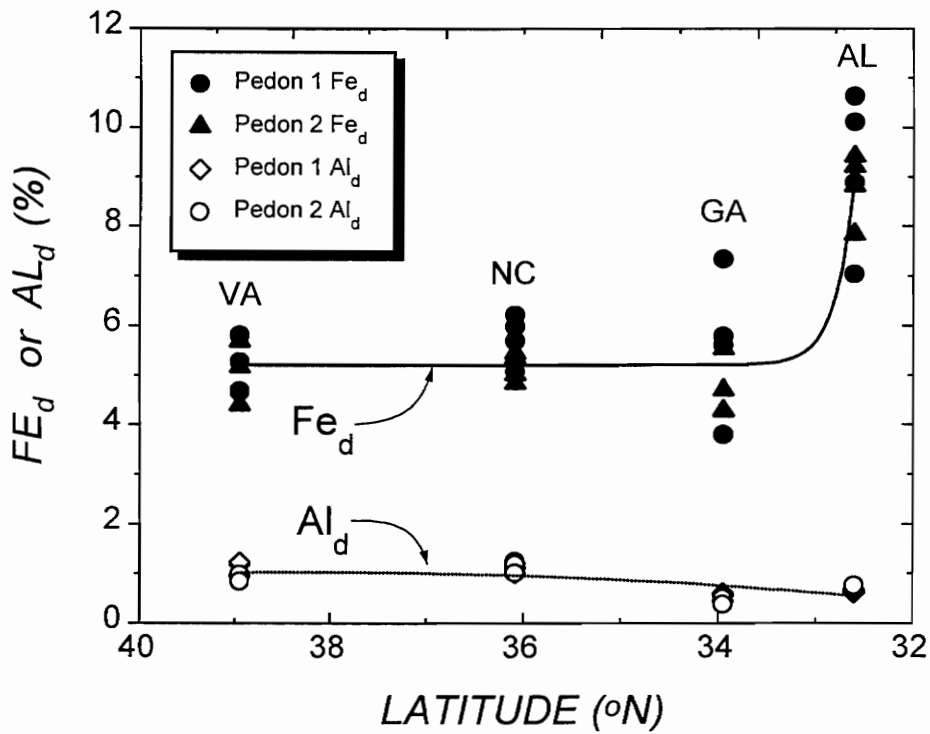


Fig. 5-8. Clay fraction Fe_d and Al_d for selected solum horizons as a function of latitude. Fe_d shows an exponential increase in Alabama soils. Loess horizons, as non-residual materials, have been omitted from the regression analysis.

climatic effect. There was no significant relationship between clay fraction Al_d and latitude, although Al_d appears to decrease slightly from north to south. Oxalate-extractable forms of Fe and Al, which consist primarily of poorly crystalline or amorphous Fe-oxides, ferrihydrite, and organically-bound Fe (Schwertmann, 1985, 1989) are present in very minor amounts that have the same trends with respect to latitude as Al_d (Fig 5-9).

The Fe_o/Fe_d ratio is used as an index of crystallinity for Fe-oxide minerals present (Schwertmann, 1988). Extremely low values of Fe_o/Fe_d , ranging from ~0.013 in Virginia and North Carolina to ~0.025 in Georgia and Alabama (Fig. 5-10) indicate high pedogenic Fe-oxide crystallinity. McFadden and Hendricks (1985) reported typical Fe_o/Fe_d values of 0.22-0.51 for mid-Holocene to late Pleistocene soils, and ratios of <0.1 in older soils. Alexander (1974) found that the Fe_o/Fe_d ratio decreased to 0.18 in the oldest (>0.4 ma) soil of a chronosequence along the Truckee River in Nevada. Torrent et al. (1980) similarly reported Fe_o/Fe_d values ranging from 0.15 to 0.04 for terraces in another chronosequence in Spain. Ratios of Fe_o/Fe_d were inversely related to hematite content and soil redness in these studies suggesting that i) lower ratios are an indication of increased soil age or weathering intensity and ii) ferrihydrite is an initial metastable precipitate formed prior to hematite. Although there is a significant increase in Fe_o/Fe_d from north to south on the Piedmont (Fig. 5-10), the absolute values are so small as to render any trends meaningless.

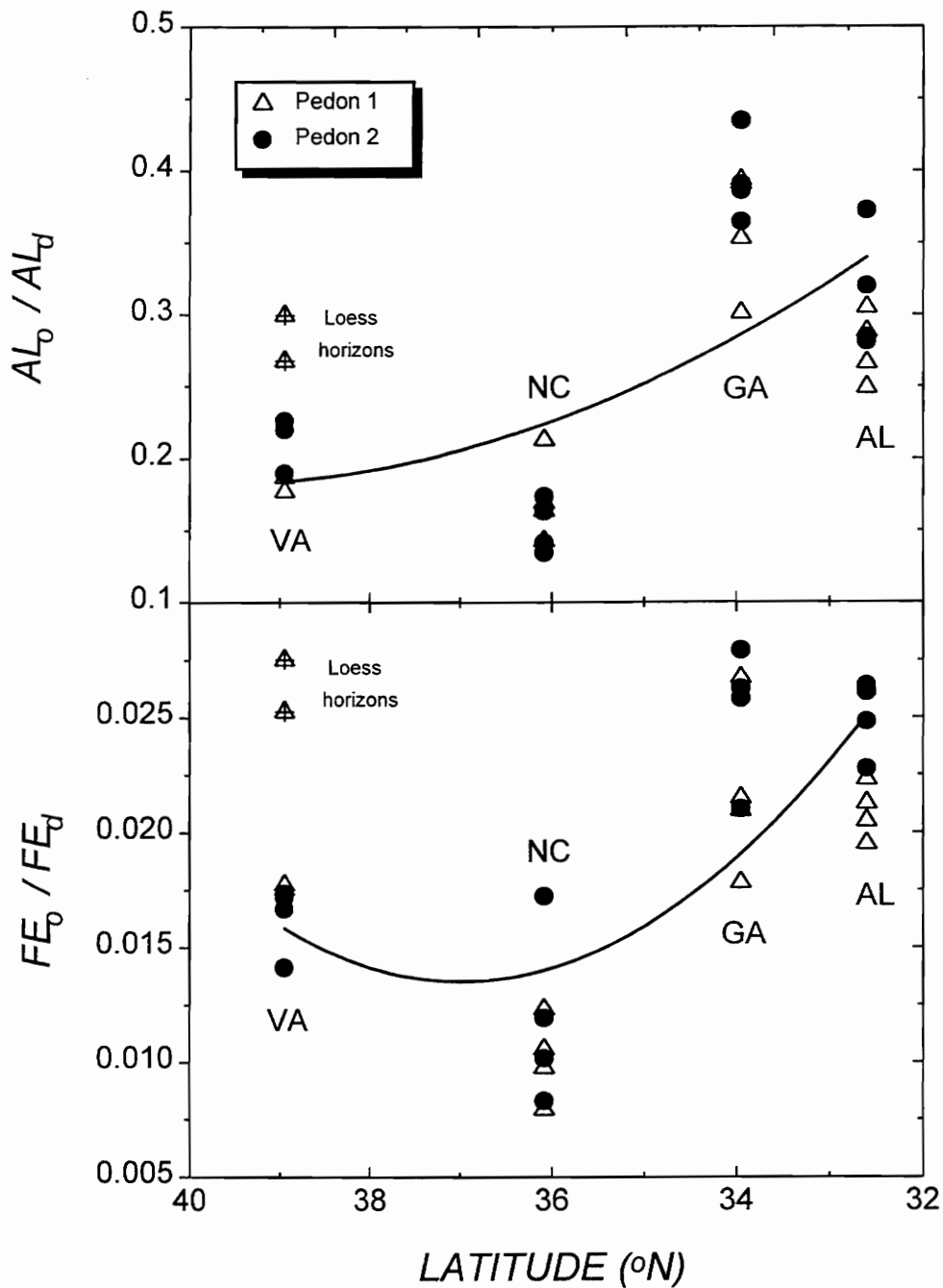


Fig. 5-10. Al_0 and Fe_0 show a curvilinear increase relative to Al_D and Fe_D from north to south on the Piedmont, suggesting that the relative degree of crystallinity of Fe-oxides decreases farther south.

Conclusions

With few exceptions, all morphological, physical, chemical, and mineralogical measures of weathering increase dramatically from north to south on the Piedmont over similar rock types. The effective depth of pedogenesis in the Georgia and Alabama soils studied, as measured by soil rubification, distribution of desilication features, accumulation and distribution of pedogenic clay and Fe_d , and the depth at which pseudomorphic gibbsite and halloysite forms in these profiles, is nearly twice that of soils in Virginia and North Carolina. These trends closely correspond to calculated effective leaching indices developed for each site based on modern climatic averages. This suggests that despite widespread absolute climatic fluctuations during the Quaternary, the relative differences in mean annual temperature, precipitation, and potential evapotranspiration that now exist from north to south on the Piedmont may have been analogous to the long-term climatic signal under which these soils formed.

Until we can obtain numerical dates for these Piedmont soils, one can only speculate as to whether the differences in observed weathering intensity are due to age effects, or solely to climatic effects. Alabama soils, in contrast to the soils studied in Virginia, North Carolina, and Georgia, appear to have reached a steady-state with regard to clay production. This may indicate a condition of greater stability and longer weathering farther south on the Piedmont.

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Chapter VI

Incipient Biotite Alteration in a Deeply-Weathered Granitic Saprolite

ABSTRACT

Despite widespread documentation of biotite kaolinization in many well-drained soils, the mechanism by which a 2:1 trioctahedral phyllosilicate is pseudomorphically replaced by a 1:1 dioctahedral mineral, while preserving the external outline of individual grains, remains obscure. Scanning electron micrographs of altered biotite grains obtained from deep granitic saprolite samples in the Virginia Piedmont show the topotactic formation of tubular halloysite within exfoliated biotite sheets. Halloysite formation proceeds from grain edges inward until little or no residual biotite remains in the interior. We interpret the phenomenon of “kaolinized” biotite previously reported in the literature as originating from replacement of halloysite by kaolinite through a dissolution/recrystallization mechanism higher in the weathering profile. In this scenario, kaolinite forms epitaxially with respect to the original mica host grain, thereby retaining the same orientation and similar crystallographic properties. This may explain why the close-packed anion layers of the tetrahedral and octahedral sheets appear to have been preserved.

Introduction

The kaolinization of biotite is a widely reported phenomenon occurring in many well-drained, highly leached soils throughout the world (Stoch and Sikora, 1976; Eswaran and Bin, 1978; Gilkes and Suddhiprakarn, 1979). Formation of kaolinite from biotite single grains involves the solid-state, pseudomorphic replacement of a 2:1 trioctahedral phyllosilicate by a 1:1 dioctahedral mineral with preservation of the external outline of individual grains. With the exception of lower refractive indices, paler colors, lower birefringence and pleochroism, and lower resolution of interference figures (Harris et al., 1985b), these pseudomorphs retain the dominant optical properties of biotite (e.g., $2V\alpha=5^\circ-15^\circ$, brown/green pleochroism, etc.) while displaying the 7.2\AA x-ray diffraction spacing characteristic of kaolinite.

Although much attention has been given to biotite kaolinization in the literature in recent years, questions regarding the pedoenvironmental factors regulating biotite alteration, and the specific mechanisms involved in the kaolinization of natural biotites have not been resolved. The effects of temperature on weathering pathways and products, for example, are unknown. Whereas kaolinite has been shown to be the dominant weathering product of biotite in mesic (MAST¹ 8° to 15°C) and warmer soils of the southeastern U.S. (Harris et al., 1985a,b) and the tropics (Ojanuga, 1973), Feldman et al. (1991) found that vermiculitization through an interstratified intermediary, and not kaolinization of biotite, was the dominant weathering mechanism in the intensively,

¹ Mean annual soil temperature (at 50 cm depth).

leached frigid soils at the higher elevations of the southern Blue Ridge. Similar results reported by Ghabru et al. (1989) for Canadian Spodosols suggest that the specific products of biotite weathering are temperature-dependent, but quantitative limits for this effect and the possible effects of other factors, such as high levels of soil organic acids, are not known.

The mechanism by which a 2:1 phyllosilicate isovolumetrically converts to 1:1 phyllosilicate without loss of grain integrity also remains obscure. Vermiculitization simply involves loss of K^+ and oxidation of structural Fe, while leaving the double siloxane structure intact. Kaolinization of biotite, on the other hand, requires i) loss of ~50% of the existing Si in the structure, ii) replacement of divalent octahedral cations (Fe, Mg, etc.), usually by trivalent Al, and iii) reduction of tetrahedral charge by substitution of Si^{4+} for Al^{3+} . Given the magnitude of structural and chemical alterations required to accomplish this transformation in both the anion and cation layers of the host, it is difficult to invoke anything other than a dissolution/recrystallization mechanism to explain the phenomenon. The dilemma of how the orientation and external integrity of isovolumetrically-altered biotite grains have been preserved after such a transformation, however, remains an enigma.

This paper presents previously unreported morphological and compositional evidence which suggests that following oxidation of structural Fe and loss of interlayer K, the initial stages of biotite kaolinization on the Virginia Piedmont actually involve formation of halloysite within frayed biotite planar (001) surfaces. Halloysite thus formed

is replaced by kaolinite in higher horizons, while retaining the external grain shape of the original host.

Methodology

Intact single grains of biotite were isolated from two deep soil/saprolite profiles located on convex interfluvies on the outer Piedmont Crystalline Province (Thornbury, 1965), near Woodbridge, VA. The area is underlain by the Occoquan Batholith, which consists of a medium-grained, light gray monzogranite with moderate to strong metamorphic foliation and a well-developed mineral lineation (Seiders et al 1975).

Soil samples were obtained with a bucket auger from a depth of 400-500 cm in profile VA 1, and from a depth of ~775 cm in profile VA 2. Samples were air-dried, sieved to remove coarse fragments (>2 mm), and thoroughly mixed. They were pretreated with dithionite-citrate-bicarbonate to remove Fe-oxide coatings (Holmgren, 1967) and then adjusted to pH 9.5 with *M* Na₂CO₃ to effect particle dispersion. Sand (>50 μ) was separated by wet sieving, and the remaining clay (<2 μ) and silt (2-50 μ) fractions were separated by repeated decantation after centrifugation. The silt fraction was further subdivided into coarse (20-50 μ), medium (5-20 μ), and fine (2-5 μ) separates by decantation after repeated centrifugation (Jackson et al., 1950). The coarse silt fraction was used for SEM analysis.

Samples were prepared for scanning electron microscopy (SEM) by inverting small vials of the coarse silt fractions onto an adhesive stub that was then coated with a thin

layer of Au-Pb to conduct excess charge away from the electron beam. Micrographs were obtained at a fixed working distance and operating voltage of 11 mm and 30 kV, respectively, using a Camscan™ Series 2 SEM. The system was also equipped with a HNU™ energy dispersive spectrometer, which was operated at a working distance of 27 mm, and a voltage of 15 kV.

Results and Discussion

The very early stages of biotite alteration in the soils studied involve neoformation of tubular halloysite on relatively unaltered biotite surfaces (Fig. 6-1a). The lack of exfoliation features on the biotite surface shown in Fig. 1a, and the presence of euhedral gibbsite crystals coexisting with halloysite suggest that the initial precipitation of these secondary phases sometimes occurs as Si and Al ions are released to solution by the weathering of other primary minerals.

More commonly, however, tubular halloysite crystals of $\sim 1 \mu$ in length grow topotactically with respect to biotite planar (001) surfaces where space permits within expanded sheets at frayed edges (Fig. 6-1b). Alteration of the biotite appears to proceed around the grain periphery, continuing from the edges inward as shown for grains in both pedons VA 1 and VA 2 (Fig. 6-2a-c). The kaolinized edges of these biotite grains are reduced to 1.5% and 1.3% K_2O and FeO , respectively (Table 6-1), indicating that oxidation of Fe and/or leaching of interlayer K is requisite to the exfoliation of biotite layers, which precedes formation of topotactic halloysite.

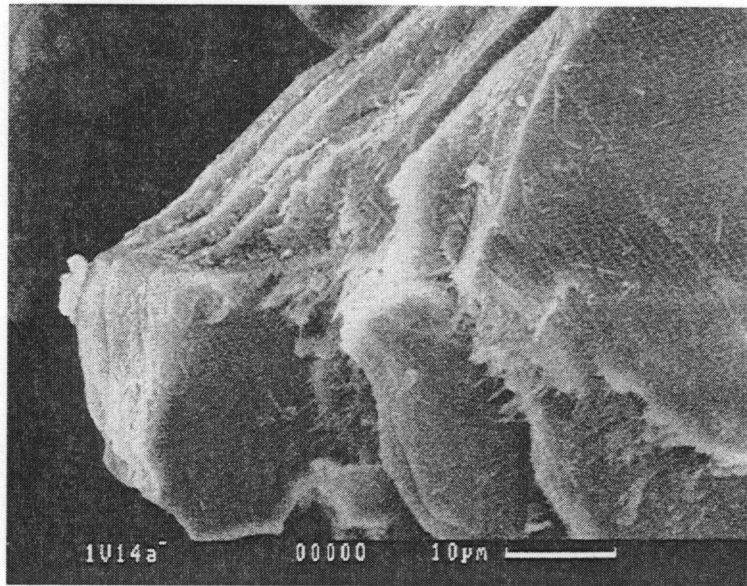
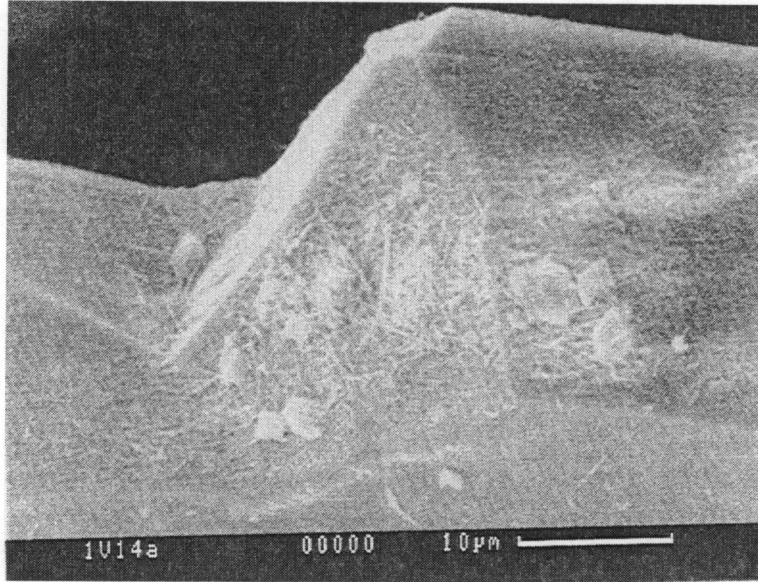


Fig. 6-1. Tubular halloysite and euhedral gibbsite crystals precipitated on an unaltered biotite surface (top). Incipient topotactic growth of halloysite tubes between expanded biotite sheets (bottom).

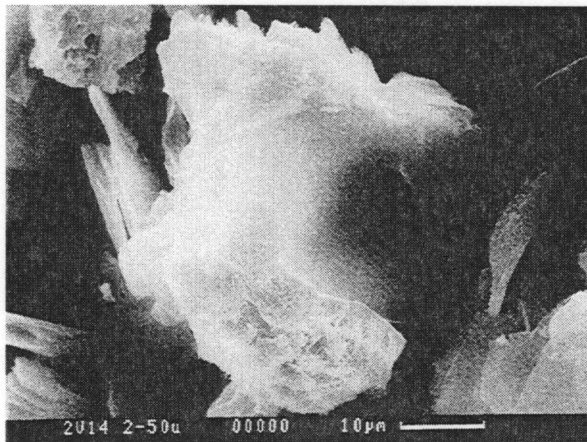
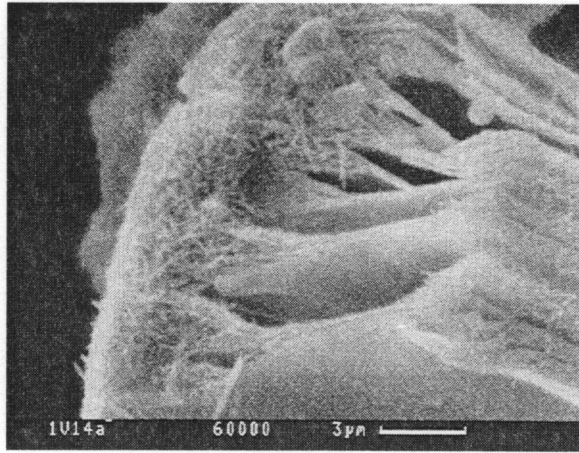
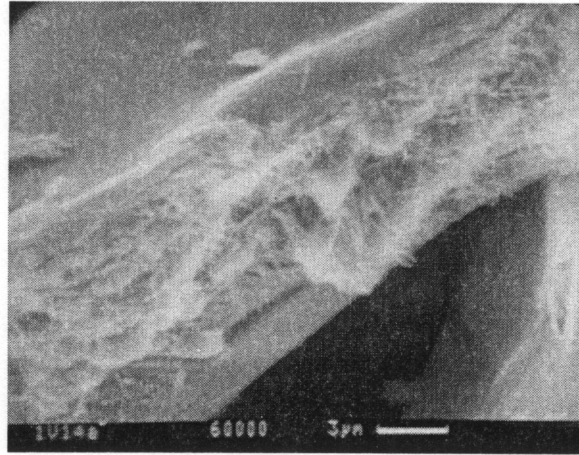


Fig. 6-2. (a,b) Continued replacement of biotite by halloysite at grain edges. (c) Alteration of biotite proceeds from the grain periphery toward an intact core.

Most unweathered biotite grains interiors appear to be morphologically unaltered, but they are in fact chemically weathered as indicated by low values of K₂O (~8-9%) and FeO (~4-5%) (Table 6-1).

Table 6-1. Chemical composition of selected weathered and unweathered biotite grains.

Component	Unweathered Biotite (n=4)	Weathered Biotite (n=2)
	%	
SiO ₂	52.2±2.2	52.6±1.4
Al ₂ O ₃	32.7±1.2	41.7±1.4
Fe ₂ O ₃	4.5±0.3	1.3±0.6
K ₂ O	8.2±0.8	1.5±2.0
MgO	2.7±0.8	2.7±0.5
SiO₂ / Al₂O₃	1.6±0.1	1.3±0.1

Compared with the corresponding values of 16-42% FeO and 9-10% K₂O for an ideal biotite (Fanning et al., 1989), it is apparent that even 'unaltered' biotite grain interiors undergo considerable oxidation and loss of Fe before removal of interlayer K and lattice expansion proceeds and kaolinization is initiated. The loss of Fe and K are therefore considered to be primary weathering mechanisms common to *both* the kaolinization and vermiculitization of biotite. These results contrast with the findings of Harris et al. (1985b) who found only minimal loss of FeO in unaltered biotite grain interiors.

Weathered (kaolinized) biotite grains are further depleted in FeO and K₂O and enriched in Al₂O₃ as this mineral recrystallizes in dioctahedral form. Although SiO₂ levels remain constant, possibly reflecting a volume increase associated with expanded

edges, the $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratio decreases from 1.6 in the unweathered biotite to 1.3 in the kaolinized grains.

Polished thin sections of grain mounts prepared for the sample VA 1 coarse silt fraction indicate that exfoliation of biotite edges results in a considerable volume expansion of ~300-400% (Fig. 6-2b, 6-3a-b) indicating that saprolite formation from weathered rock on the Piedmont may not be an entirely isovolumetric process, particularly in areas high in mafic rocks. The striking contrast between these expanded edges and grain interiors shown in polished section may partially explain why x-ray diffraction studies and petrographic investigations of grain mounts preferentially record only the oriented basal (00 l) properties of the residual mica.

Formation of halloysite appears to precede the precipitation of gibbsite, which is shown as high-contrast, euhedral crystals of ~1-3 μ across on biotite surfaces and interlayer areas (Fig 4-1b, 4-4a-d). A series of micrographs of increasing magnitude shows that halloysite tubes of ~1 μ in length commonly bisect gibbsite crystals (Fig. 4-4d), suggesting that gibbsite has precipitated around an existing halloysite framework. This trend is also corroborated by x-ray diffraction evidence which shows that halloysite, and not gibbsite, is present in the deepest, or youngest (2C9b), horizon of pedon VA 1. This horizon is the most recent saprolite layer formed as the weathering front descends toward bedrock. Gibbsite begins to appear in the profile only in the overlying and more highly weathered 2C8b horizon (Feldman et al., 1995a, in preparation).

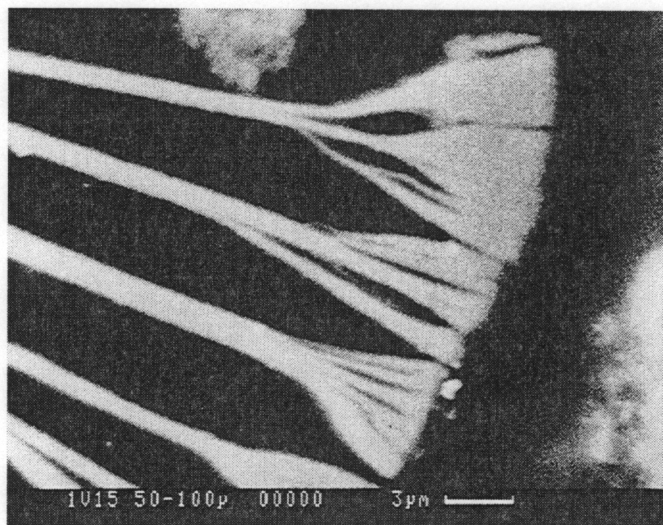
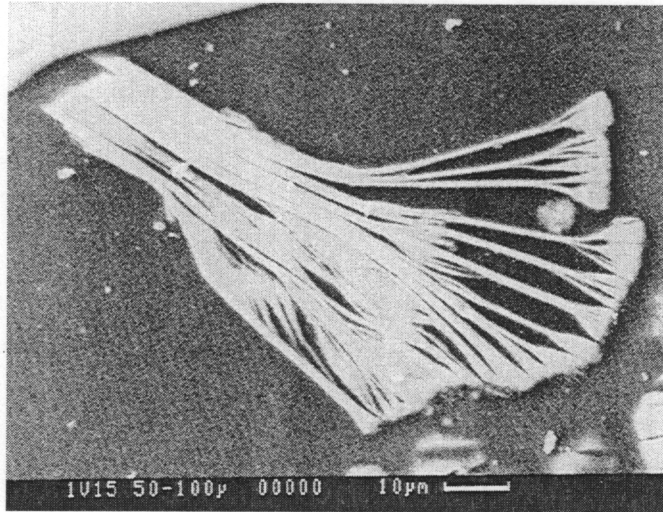


Fig. 6-3. (a,b) Polished thin sections of altered biotite grains showing considerable volume expansion associated with the exfoliation of edges.

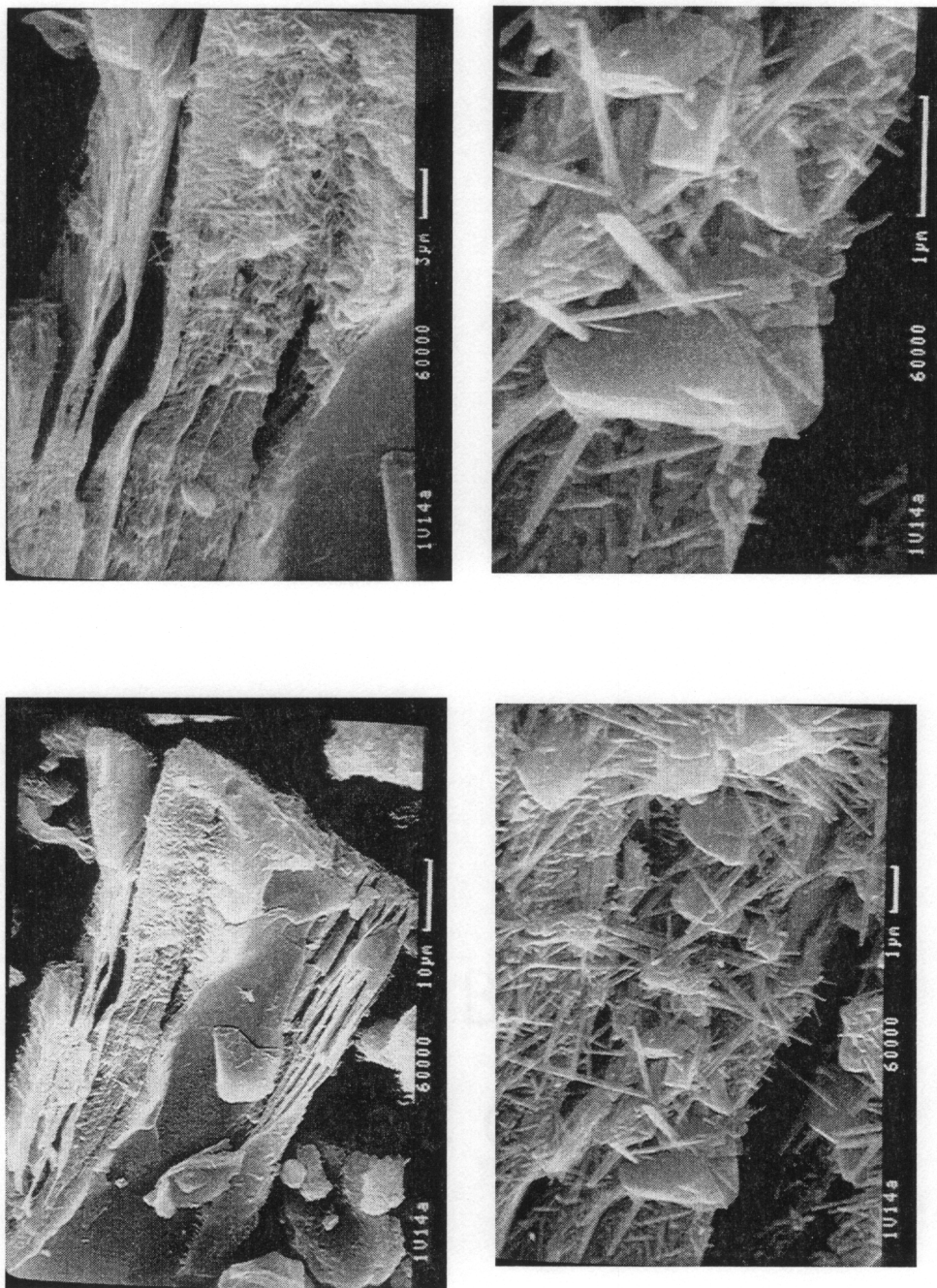


Fig. 6-4. (a-d) Increasing magnification of an altered biotite grain from a depth of 5 m, showing halloysite tubes of $\sim 1 \mu$ in length bisecting gibbsite crystals, and suggesting that gibbsite has precipitated around an existing halloysite framework.

Conclusions

Biotite weathering is clearly a complicated process that involves oxidation of Fe and loss of interlayer K, regardless of the secondary phase produced. The ubiquity of halloysite produced in saprolites throughout the Piedmont (Calvert et al., 1980a,b; Feldman, 1995b, in preparation), whether from a feldspar or a biotite precursor, indicates that the microenvironment of formation is an overriding factor in determining the nature of the secondary mineral produced, irrespective of the composition of the primary mineral undergoing transformation. Mixed-layer minerals and discrete vermiculite are not the normal weathering products in Piedmont soils formed from felsic materials, presumably because desilication and removal of ions in solutions proceeds too rapidly for the formation of 2:1 phyllosilicates.

The initial loss of Fe and K in biotite grains examined in this study is followed by topotaxial growth of halloysite, which recrystallizes to form kaolinite higher in the profile. In this scenario, kaolinite could form epitaxially with respect to the original mica host grain while retaining its same orientation. This may explain why the close-packed anion layers of the tetrahedral and octahedral sheets appear to have been preserved, and why single biotite grains having the diffraction characteristics of kaolinite have previously been observed (Harris et al., 1985a,b).

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Chapter VII

Summary and Conclusions

The Southern Piedmont physiographic region may be particularly sensitive to potential changes in climate because of high population density and agricultural drought hazard. Despite the acknowledged dominion of temperature, precipitation, and potential evapotranspiration over many soil properties such as depth of weathering and degree of profile development, there is currently a scarcity of quantitative data regarding the specific effects of both climate and time on weathering processes. This limitation, due largely to a lack of age control for deposits >40,000 yr, makes it impossible to differentiate between the combined effects of climate and geomorphic age on weathering intensity, soil profile development, mineral stability, and elemental flux within the pedoenvironment. Without this information, we can neither identify the degree to which soil properties and secondary phases persist, alter, or become obliterated in the weathering environment as conditions change, nor can we predict with any certainty which soil features are reliable as indicators of retrospective or prospective environmental change.

In addition to acknowledged uncertainties regarding climatic effects on weathering, other aspects concerning the evolution and age of the Appalachian landscape remain unresolved. The widespread occurrence of stone lines and colluvial caps on stable

uplands as far south as Georgia also suggests that climatic oscillations highly influenced geomorphic processes throughout much of the Piedmont during the Pleistocene. In this study, a previously unidentified, discontinuous loess deposit on the Piedmont in northern Virginia was dated by thermoluminescence at 13.8 ± 1.0 ka BP, documenting late glacial conditions of strong winds with little or no rain, and the existence of bare or sparsely vegetated ground. Although most of the loess was ultimately eroded by Holocene fluvial processes from all but the most stable, level ridgetops in the area, mineralogical evidence indicates that post-depositional periglacial activity incorporated much of it into the upper horizons of existing soils on the local landscape.

A weakly-developed argillic horizon formed in a ~1 m thick intact loess contrasts sharply with an underlying buried paleoargillic horizon formed in granitic residuum. Differences in clay-free silt content from between ~55-60% in the loess cap to <25% in the underlying paleosol correspond with similar abrupt increases in CEC, Fe_d , and exchangeable Mg below the stone line, indicating that these relict features could not have formed within the 13.8 ka since the loess was deposited. Moreover, a high-charge smectite phase was identified in the paleosol, which is inconsistent with long-term weathering under the modern conditions of intense, acid leaching. The formation of this mineral phase may be related to drier climatic conditions at or before the time of loess emplacement.

With few exceptions, all morphological, physical, chemical, and mineralogical measures of weathering increase dramatically from north to south on the Piedmont over

similar rock types. The effective depth of pedogenesis in the Georgia and Alabama soils studied, as measured by soil rubification, distribution of desilication features, and the depth at which pseudomorphic gibbsite and halloysite forms in these profiles, is nearly twice that of soils in Virginia and North Carolina. These trends closely correspond to calculated effective leaching indices developed for each site, suggesting that despite widespread absolute climatic fluctuations during the Quaternary, the relative differences in mean annual temperature, precipitation, and potential evapotranspiration that now on the Piedmont may be analagous to the long-term climatic signal under which these soils formed.

Kaolinite and gibbsite, which are commonly regarded as stable end products of weathering in humid-temperate soils, are poor indicators of soil age, weathering intensity, or paleoenvironment. Kaolinite genesis is shown to result from at least four complex mechanisms in Piedmont soils, including: i) desilication of 2:1 phyllosilicates induced by intense leaching in surface or near-surface horizons ii) hydrolysis of K-feldspars, iii) resilication of gibbsite deep in the saprolite, and iv) replacement of halloysite at a depth in saprolite horizons. Gibbsite similarly forms from both desilication of kaolinite in surface horizons, and also by reprecipitation after Na-feldspar dissolution at depths of >5 m. Halloysite is produced in all soils, illustrating the importance of microenvironment in determining the nature of the secondary minerals produced, irrespective of the composition of the primary mineral undergoing transformation.

Total clay and Fe_d contents directly reflect the relative age and degree of weathering of Piedmont soils. The relationship between clay content and Fe_d in Virginia, North Carolina, and Georgia soils is highly linear, with decreasing slopes from north to south. clay production levels off at higher Fe_d contents in Alabama soils. This indicates that i) steady-state conditions are approached in these soils as pedogenic clay formation and illuviation reach some intrinsic, threshold value that becomes self-limiting, and ii) the limiting factor in clay production over time is leaching intensity, and not the concentration of Fe-bearing primary minerals. These results may indicate a condition of greater stability and longer weathering farther south on the Piedmont.

There is a clear gradient in weathering intensity and both soil morphological and physicochemical properties that is related to latitude on the Piedmont. However, until we can obtain accurate numerical dates for these Piedmont soils, one can only speculate as to whether the differences in observed weathering intensity are due to age effects, climatic effects — or both.

Appendix A: Profile Descriptions of the Eight Primary Sampling Sites

PEDON: VA 1

Location: Prince William Co., Virginia; Occoquon quadrangle northwest of Woodbridge; from center of bridge over Occoquon River on VA 123 (Ox Rd.), 0.8 km (0.5 mi) south on VA 123; then 2.4 km (1.5 mi) west (right-hand turn) on VA 663 (Davis Ford Rd.); then straight for 3.2 km (2.0 mi) onto VA 641 (Old Bridge Rd.) where it intersects with VA 663 (663 goes to the left, 641 continues straight); roadcut is on right, just past Hedges Run Rd., and ~100 m SE of water tower.

Elevation: 97 m (320 ft)

Aspect: 82°SW

Slope: 3%

Geomorphic: Upland ridgetop summit; slightly convex

Relief: Overall relief in the area is about 70 m; landscape is highly dissected with narrow, highly incised ridgetops above backslopes that descend steeply into valleys occupied by perennial streams.

Parent Material: Fine-textured (silty) recent sediment (loess) overlying residuum and saprolite weathered from the Occoquon Granite.

Vegetation:

Canopy: None

Understory: None

Ground Cover: Pasture - fescue, broomsedge (second year fallow)

Classification: Typic Kanhapludult, fine-loamy, mixed, mesic

Description:

A --- 0 to 8 cm., Dark yellowish brown (10 YR 4/4) silt loam or loam; moderate fine and very fine granular structure; very friable; many fine and medium roots; abrupt smooth boundary.

E --- 8 to 15 cm., Grayish brown (10 YR 5/2) silt loam or loam; weak medium platy parting to fine and very fine granular structure; firm; many fine and medium roots; clear smooth boundary.

BE --- 15 to 30 cm., Yellowish brown (10 YR 5/4) silt loam with common fine prominent strong brown (7.5 YR 5/6) mottles in vesicular pores and along cracks, and few fine distinct grayish brown (10 YR 5/2) mottles throughout the matrix; weak medium subangular blocky structure; firm; common fine roots; clear smooth boundary.

Bt1 --- 30 to 38 cm., Yellowish brown (10 YR 5/6) silty clay loam with common medium distinct yellowish brown (10YR 5/4) mottles; weak medium subangular blocky structure; friable; common fine roots; few thin clay films on ped faces; gradual smooth boundary.

- Bt2 --- 38 to 51 cm., Strong brown (7.5 YR 5/6) heavy silty clay loam; weak medium subangular blocky structure; friable; few thin clay films; uniform ~8 cm thick stone line of ~50-75% unweathered angular vein quartz found in basal zone; clear smooth boundary.
- 2Bt3b --- 51 to 74 cm., Strong brown (7.5YR 5/8) clay with common medium distinct yellowish brown (10YR 5/4) mottles; strong medium subangular blocky structure; firm; thick continuous strong brown (7.5YR 5/6) clay films; few fine roots; gradual smooth boundary.
- 2Bt4b --- 74 to 109 cm., Strong brown (7.5YR 5/8) clay with common medium distinct yellowish brown (10YR 5/4) mottles; moderate very coarse and coarse prismatic parting to strong medium subangular blocky structure; firm; thick continuous strong brown (7.5YR 5/6) clay films on blocky peds, many continuous brown (7.5YR 4/4) clay flows on prism faces; few fine roots in cracks between prisms; gradual wavy boundary.
- 2CBb --- 109 to 178 cm., Mottled yellowish red (5YR 5/6), strong brown (7.5YR 5/6), and yellowish brown (10YR 5/4) sandy clay loam or sandy loam; massive saprolite structure in place breaking to weak medium subangular structure; friable; gradual smooth boundary.
- 2C1b --- 178 to 213 cm., Mottled strong brown (7.5YR 5/6), yellowish brown (10YR 5/4), and yellowish red (5YR 5/6); sandy clay loam or sandy loam; saprolite from granite; friable; gradual smooth boundary.
- 2C2b --- 218+ cm., Mottled yellowish red (5YR 5/6), strong brown (7.5YR 5/6), and yellowish brown (10YR 5/4) sandy clay loam or sandy loam; massive saprolite from granite; friable.
- 2C3b --- 244 to 264 cm., Mottled Yellowish brown (10YR 5/8), light gray (10YR 7/2), dark yellowish brown (10YR 4/4), and strong brown (7.5YR 4/6) loamy, massive, friable saprolite.
- 2C4b --- 284 to 295 cm., Mottled Yellowish brown (10YR 5/8), light gray (10YR 7/2), dark yellowish brown (10YR 4/4), and strong brown (7.5YR 4/6) loamy, massive, friable saprolite.
- 2C5b --- 401 to 411 cm., Dark and light gray weathered granite with dark yellowish brown (10YR 4/4) and yellowish brown (10YR 5/8) zones; highly micaceous.
- 2C6b --- 498 to 508 cm., Dark and light gray weathered granite.

2C7b --- 508 to 518 cm., Dark and light gray weathered granite.

2C8b --- 528 to 564 cm., Dark and light gray weathered granite.

2C9b --- 564+ cm., Dark and light gray weathered granite; pink feldspars evident.

Fieldnotes, Observations, Speculation, and Wild-Ass Guesses (VA1 Pedon):

Described from a vertical cut in a borrow pit. Horizons 2C3b - 2C8b were sampled and described by auger observation; therefore, reported thickness ranges for these horizons reflect only sampling depth, not true horizon thickness. Horizon designations are tentative, pending further laboratory analyses.

E and BE horizons have brittleness, massive to weak platy structure, and some mottling patterns that would suggest fragic character; however, strong morphological evidence is lacking for a fragipan (e.g., no prisms or E' horizon). Water movement and depth of rooting are restricted in this zone, however, as indicated by the wetting front observed during the rainy season which extends only to the contact with the underlying residuum.

Saprolite grades to weathered bedrock at a depth of ~5 m at this site. This depth is very shallow when compared both with deeper observations in the area, and with reports in the literature, suggesting that bedrock topography is irregular in localized areas.

This soil is polygenetic in that more than one interval of soil genesis has taken place. A stone line of unweathered, angular vein quartz truncates a buried 2Btb horizon continuously at a depth of about 50 to 75 cm throughout the entire exposure (>100 m distance). The loess overlying the stone line is of uniform thickness throughout the cut. The probable sequence of pedogenic events is as follows: Strongly developed horizons were formed initially in residuum from granite saprolite. Pedogenesis was interrupted at some point for a time with truncation of the profile and development of a veneer of lag gravel prior to deposition of the present silty mantle. After deposition of the loess, horizon differentiation resumed, and a weak argillic horizon developed within this material. The weak expression of this horizon, and the lack of morphological evidence to indicate pedogenic transfer of material from the silty mantle into the underlying buried Bt horizon, suggests that the most recent period of horizon differentiation has been short since the mantle was deposited. This hypothesis is currently being tested by thermoluminescence dating of the silty mantle.

This pedon is somewhat unique for several reasons: 1) it is one of the few remaining sites here the overlying silty cap remains more or less (?) intact, 2) it occupies the highest point in the local landscape, and 3) the stone line at the discontinuity of this soil is comprised of unweathered vein quartz which contrasts with the majority of other upland ridgetop soils throughout the area that have stone lines with 2-3 mm thick weathering rinds and are overlain by a much more highly developed (colluvial) surficial deposit. This would suggest that formation of the old erosional surface and emplacement

of the colluvium on these uplands occurred before deposition of the loess. Moreover, stone line gravels underlying this material is nearly identical to the stone lines underlying other upland colluvial deposits in Georgia and Alabama with respect to lithology and weathering rind thickness. Does this suggest synchronous deposition of upland colluvium across a large area of the Piedmont? If so, this would imply regional vs. local controls on sedimentation throughout much of the entire Piedmont, governed by adjustment to climatic and/or tectonic changes.

Strong brown (7.5YR 5/6) clay films are dominant on strongly developed subangular blocky peds in the upper horizon of the paleosol (2Bt3b horizon) at this site, and contrast sharply with the dark yellowish brown (7.5YR 4/6) clay flows on prism faces in the underlying (2Bt4b) horizon. This would indicate that they are either the result of pedogenesis following loess emplacement, or the well-preserved (uneroded) remnant of the paleo-argillic horizon. I would predict the latter.

PEDON: VA 2

Location: Occoquon Quad. - Ox Rd. area of Fairfax Co, VA, northwest of Lorton Correctional Facility; From center of Occoquon Bridge on VA 123, go 4.7 km (2.9 mi) N on VA 123, then left onto Hampton Station Rd for 0.2 km (0.1 mi) to end of cul-de-sac; site is in field on right.

Elevation: 97 m (320 ft)

Aspect: 90°E facing

Slope: 4% water flow direction; 2% along ridge.

Geomorphic: Upland ridgetop summit; slightly convex

Relief: Overall relief in the area is about 70 m; landscape is highly dissected with narrow, highly incised ridgetops above backslopes that descend steeply into valleys occupied by perennial streams.

Parent Material: Colluvium with an admixture of loess overlying residuum and saprolite from the Occoquon Granite.

Vegetation:

Canopy: None (mixed oak/hickory forest near site)

Understory: None

Ground Cover: Pasture - fescue, broomsedge (second year fallow)

Classification: Typic Kanhapludult, clayey, mixed, mesic

Description:

Ap --- 0 to 13 cm., Dark brown (10 YR 3/3 - 4/3) loam; weak fine and very fine granular structure; very friable; many fine and medium roots; about 10% quartz gravels (<1 cm); abrupt smooth boundary.

BA --- 13 to 20 cm., Dark yellowish brown (10 YR 5/4) clay loam; weak medium subangular blocky structure; friable; many fine and medium roots; about 10% quartz gravels (<1 cm); clear smooth boundary.

Bt1 --- 20 to 30 cm., Yellowish brown (10 YR 5/6) silty clay; weak medium subangular blocky structure; firm; common fine roots; about 10% quartz gravels (<1 cm); clear smooth boundary.

Bt2 --- 30 to 46 cm., Yellowish brown (10 YR 5/8) clay moderate medium subangular blocky structure; firm; common thin clay films strong brown (7.5YR 4/6) on ped faces; common fine roots; about 10% quartz gravels (<1 cm) throughout matrix; stone line of about 25% quartz gravel up to 7 cm across is found in the lower 10 cm of horizon; clear smooth boundary.

2Bt3b --- 46 to 91 cm., Strong brown (7.5 YR 5/6) clay with common fine yellowish red (5YR 4/6) mottles; strong subangular blocky structure; firm; common thick

continuous strong brown (7.5YR 4/6) clay films on ped faces; few fine roots; about 10% quartz gravels (<1 cm); clear wavy boundary.

2Bt4b --- 91 to 147 cm., Yellowish red (5YR 5/6) clay with common medium distinct strong brown (7.5YR 5/8) and yellowish brown (10YR 5/4) mottles; weak fine subangular blocky structure, massive in place; firm; clear wavy boundary.

2BC1b --- 147 to 182 cm., Mottled saprolite with banded zones of oxidized yellowish red (5YR 5/6) and strong brown (7.5YR 5/8) clay loam material, and unoxidized zones of light yellowish brown (10YR 6/4), brownish yellow (10YR 6/6), and yellowish brown (10YR 5/8) loam; massive; friable; high mica content in the unoxidized zones.

2BC2b --- 234 to 244+ cm., Mottled light gray (10YR 7/2), yellowish brown (10YR 5/6), yellowish brown (10YR 5/4), strong brown (7.5YR 5/6), and yellowish red (5YR 4/6) loam or sandy loam saprolite; dark mafic or iron rich nodules throughout.

2C1b --- 284 to 295 cm., Mottled yellowish brown (10YR 5/8), strong brown (7.5YR 5/6), dark yellowish brown (10YR 4/4), and light gray (10YR 7/2) loam or sandy loam saprolite.

2C2b --- 340 to 350 cm., Mottled yellowish brown (10YR 5/8), strong brown (7.5YR 5/6), dark yellowish brown (10YR 4/4), and light gray (10YR 7/2) loam or sandy loam saprolite with common dark mafic or Fe-rich nodules.

2C3b --- 411 to 421 cm., Mottled Yellowish brown (10YR 5/8), strong brown (7.5YR 5/6), dark yellowish brown (10YR 4/4), and light (10YR 7/2) loam or sandy loam saprolite with common dark mafic or Fe-rich nodules.

2C4b --- 498 to 508 cm., Mottled Yellowish brown (10YR 5/8), strong brown (7.5YR 5/6), dark yellowish brown (10YR 4/4), and light gray (10YR 7/2) loam or sandy loam saprolite with common dark mafic or Fe-rich nodules.

2C5b --- 615 to 625 cm., Mottled Yellowish brown (10YR 5/8), strong brown (7.5YR 5/6), dark yellowish brown (10YR 4/4), and light gray (10YR 7/2) loam or sandy loam saprolite with common dark mafic or Fe-rich nodules.

2C6b --- 767 to 777+ cm., Mottled Yellowish brown (10YR 5/8), strong brown (7.5YR 5/6), dark yellowish brown (10YR 4/4), and light gray (10YR 7/2) loam or sandy loam saprolite with common dark mafic or Fe-rich nodules.

Fieldnotes, Observations, Speculation, and Wild-Ass Guesses (VA2 Pedon):

Described from a vertical cut in a fresh excavation for a building foundation. Horizon designations are tentative.

Reported depths of 2C1 - 2C6 horizons represent auger sampling zones and do not reflect true horizon thickness.

Saprolite is extremely uniform to a depth of >7 m in the profile, with no evidence of weathered bedrock.

A stone line of vein quartz (&eqv.25% gravel) is uniformly found at a depth of about 40 cm throughout the pit, corresponding to the lower portion of the argillic/kandic horizon. The stone line in this soil, and other similar soils throughout the landscape,

is extremely similar to the material comprising the stone lines underlying upland colluvial caps in Georgia and Alabama soils (e.g., all areas have weathering rinds of about 2-3 mm thickness).

The argillic horizon in the colluvium above the stone line is much more highly developed than the corresponding material formed in the silty mantle of pedon VA1. It is distinguished from the underlying residual soil by:

- the higher content of disoriented coarse fragments (~5-15% by volume), sharply contrasting (low) mica content in the sand fraction,
- dominance of strong brown (7.5YR 5/6) or browner colors (vs. yellowish red (5YR) or redder colors in the underlying residuum),
- bimodal clay, exchangeable cation, and clay-free silt distributions with depth (data from a nearby satellite pedon, 200 m away on same ridgetop),
- the slight enrichment of untwinned feldspars (5%) in surface horizons,
- and the increase in whole-soil (<60 mesh) gibbsite content in the mixed zone near the paleo-surface.

No evidence of fragic properties was found in the colluvial cap of this pedon; however, a few areas of similar soils can be found that have weakly developed, brittle eluvial (E') horizons overlying Bt horizons with coarse prismatic structure.

PEDON: NC 1

Location: Person Co. North Carolina - Hurdle Mills Quad; Hurdle Mills area (Dan Whitfield, Jr. Farm); 200m SE on Charlie Monk Rd (off of Lewis Store Rd.) Site is in old beanfield about 100m NNE of the Pine Hill Church (across the street)

Elevation: 223 m (735 ft)

Aspect: 190° SW

Slope: 2%

Geomorphic: Upland ridgetop summit; convex contour, linear downslope

Relief: Maximum relief is about 66m with average relief of about 35-45m; Low-order streams occupy drainage basins of about 1 - 10km². Drainage divides are moderately incised with ridgetops extending to a maximum of about 300m in width before dropping off to shoulders/back slopes. Local perennial streams have gradients of about 0.004 m/m

Parent Material: Saprolite and residuum weathered from the Roxboro Formation granite

Vegetation: Soybean stubble

Classification: Typic Kanhapludult, clayey, kaolinitic, thermic

Description:

Ap --- 0 to 15 cm., Yellowish brown (10YR 5/4) sandy loam; weak fine granular structure; very friable; many fine roots; abrupt smooth boundary.

BA --- 15 to 28 cm., Brownish yellow (10YR 6/6) sandy loam; moderate medium granular structure; friable; many fine roots; clear smooth boundary.

Bt1 -- 28 to 48 cm., Strong brown (7.5YR 5/6) clay; few medium distinct yellowish red (5 YR 5/8) mottles in lower part; strong medium subangular blocky structure; firm; common thin discontinuous clay films on ped faces; common fine and very fine roots; clear smooth boundary.

Bt2 -- 48 to 74 cm., Strong brown (7.5YR 5/6) clay with many medium distinct yellowish brown (10YR 6/6) mottles; strong medium subangular blocky structure; firm; thick continuous strong brown clay films on ped faces; few very fine roots; clear wavy boundary.

BC1 -- 74 to 102 cm., Mottled yellowish brown, strong brown, yellowish red, and brownish yellow (7.5YR 5/8, 5/6; 5YR 5/8; and 10 YR 6/6) heavy clay loam; massive in place parting to weak medium subangular blocky structure; friable; few large strong brown clay flows on ped faces and relict rock joints; few very fine roots; clear wavy boundary.

- BC2 -- 102 to 165 cm., mottled yellowish red, strong brown, and brownish yellow (5YR 5/8; 7.5YR 5/6; and 10YR 6/6) loam; mostly massive in place with weak fine subangular blocky structure breaking out in hand samples; very friable; few clay flows; common quartz veins of 2- to 10-mm in width, oriented (dipping) about 75°, and spaced about 20cm apart; clear irregular boundary. (This is mostly highly oxidized saprolite)
- C1 -- 165 to 265 cm., mottled brownish yellow, yellow, white, reddish yellow, and yellowish red (10YR 6/8, 7/8, 8/1 7.5YR 6/8, and 5YR 5/8) loam (unoxidized saprolite with many quartz veins); very friable; relict rock structure apparent; patchy discontinuous clay films along joints; gradual boundary.
- C2 -- 265 to 295 cm, mottled brownish yellow, yellow, white, and reddish yellow (10YR 6/8, 7/8, 8/1 [quartz]; and 7.5YR 6/8) massive, unoxidized saprolite
- C3 -- 356 to 381 cm, mottled brownish yellow, yellow, white, and reddish yellow (10YR 6/8, 7/8, 8/1 [quartz]; and 7.5YR 6/8) massive, unoxidized saprolite
- C4 -- 457 to 483 cm, mottled brownish yellow, yellow, white, and reddish yellow (10YR 6/8, 7/8, 8/1 [quartz]; and 7.5YR 6/8) massive, unoxidized saprolite
- C5 -- 508 to 533 cm, mottled brownish yellow, yellow, white, and reddish yellow (10YR 6/8, 7/8, 8/1 [quartz]; and 7.5YR 6/8) massive, unoxidized saprolite with few pink feldspars

Fieldnotes:

Described from backhoe pit (deep samples >265 cm augered by hand).
No evidence of lithologic discontinuity.
Sampled and described by SBF.

PEDON: NC 2

Location: Person Co. North Carolina - Roxboro Quad; SE of Roxboro (Brooksdale) and SW of Oakley; From Hwy 158 South of Roxboro, 0.7mi NE along Oxford Rd; then left at the fork onto Allensville Rd for 1.1 mi. (right fork is *Old Allensville Rd.*); site is on right, in harvested beanfield, 20m off the road.

Elevation: 231m (760 ft)

Aspect: 20° NE

Slope: 2%

Geomorphic: Upland ridgetop summit; convex contour, linear downslope

Relief: Maximum relief in the area is about 30m

Parent Material: Saprolite and residuum weathered from the Roxboro Granite.

Vegetation: Recently-harvested soybeans

Classification: Typic Kanhapludult, clayey, kaolinitic, thermic

Description:

Ap --- 0 to 13 cm., Dark yellowish brown (10YR 4/4) loam.

BA -- 13 to 20 cm., Yellowish brown (10YR 5/4) clay loam.

Bt1 -- 20 to 41 cm., Yellowish brown (10YR 5/8) clay with moderate medium subangular blocky structure observable in auger specimen; thin continuous clay films.

Bt2 -- 41 to 56 cm., Yellowish brown (10YR 5/8) clay with common medium distinct yellowish red (5YR 5/8) mottles; moderate medium distinct subangular blocky structure; common continuous clay films.

Bt3 -- 56 to 122 cm., Red (2.5YR 4/8) clay; strong medium subangular blocky structure; thick continuous clay films.

BC1 -- 122 to 183 cm., Reddish yellow (5YR 5/6) sandy loam; common medium distinct yellowish brown and pale brown (10YR 5/8 and 6/3) mottles; oxidized saprolite with weak medium subangular blocky structure.

BC2 -- 183 to 208 cm., Yellowish brown (10YR 5/6) sandy loam saprolite.

BC3 -- 208 to 244 cm., Yellowish red (5YR 5/6) sandy loam saprolite.

Fieldnotes:

Described from adjoining auger holes with core samples taken in hole #1, and bulk samples taken from hole #2.

Ridgetop is about 400m wide at the site.

Sampled and described by SBF.

PEDON: GA 1

Location: Oglethorpe Co. Georgia; Lexington area (Alan Huff Farm); From Lexington, GA: N on Hwy 78 2.6 mi; Right at Ga. Dept. of Transportation Maintenance Shop; NE 0.7 mi and park; Site is 48 m at 5° NE from clump of trees at edge of field near road.

Elevation: 229 m (750 ft)

Aspect: 10° NE

Slope: 2-3% (water flow direction); 1% (along ridge)

Geomorphic: Upland ridgetop summit; convex contour, linear downslope

Relief: Ridgetop is about 90 m wide before dropping off to 6-8% shoulder slopes and 20% sideslopes below. Maximum relief is about 30 m with average relief of about 18 m.

Low-order streams occupy drainage basins of about 1 to 10km². Drainage divides are broad, flat, elongated uplands, and hillslopes grade gently into the main valleys at about 20%. Local perennial streams have gradients of about 0.01 m/m, ranging to about 0.015 m/m headward (near the site).

Parent Material: Saprolite and residuum weathered from Elberton Granite.

Vegetation: Fallow after no-till wheat.

Classification: Typic Rhodudult, clayey, oxidic, thermic

Description:

Ap --- 0 to 10 cm, brown (10YR 4/3) fine sandy loam; weak fine granular structure; very friable; many fine roots; abrupt smooth boundary.

BA --- 10 to 25 cm, strong brown (7.5YR 5/4 - 5/6) sandy clay loam; weak medium granular structure; very friable; common fine roots; abrupt smooth boundary.

Bt -- 25 to 61 cm, red (10R 4/6 - 4/8) clay to sandy clay; moderate medium subangular blocky structure; firm; common thin continuous clay films on ped faces; common fine and very fine roots; common fine mica flakes; gradual smooth boundary.

BC -- 61 to 109 cm, red (10R 4/6 - 4/8) clay or sandy clay with common medium distinct strong brown (7.5YR 5/6) and few medium distinct yellowish brown (10YR 5/4) mottles; weak medium subangular blocky structure breaking from platy relict rock structure; friable; common thin discontinuous clay films on ped faces and along joints; few fine roots; common fine mica flakes; gradual wavy boundary.

CB1 -- 109 to 157 cm, red (2.5YR 4/6) clay loam; common medium distinct strong brown (7.5YR 5/6) and few medium distinct yellowish brown (10YR 5/8) mottles; massive in place parting to weak medium subangular blocky structure; friable; few thin discontinuous clay films on ped faces and relict rock joints; many fine flakes of mica; gradual wavy boundary.

CB2 -- 157 to 201 cm, red (2.5YR 4/6 - 4/8) clay loam; common medium distinct strong brown (7.5YR 5/6) and few medium distinct yellowish brown (10YR 5/4 and 5/8) mottles; massive in place parting to weak medium subangular blocky structure; friable; common discontinuous clay films oriented along relict rock joints; many fine flakes of mica; clear smooth boundary.

CB3 --- 201 to 254+ cm, mottled red (2.5YR 4/6 - 4/8), strong brown (7.5YR 5/6), and yellowish brown (10YR 5/4 and 5/8) clay loam; massive in place parting to weak medium subangular blocky structure; friable; few discontinuous clay films oriented along relict rock joints; many fine mica flakes.

C1 -- 307 to 323 cm, red (2.5YR 4/6) massive oxidized saprolite.

C2 -- 370 to 386 cm, red (2.5YR 4/6) massive oxidized saprolite.

C3 -- 434 to 450 cm, yellowish red (5YR 5/6) massive oxidized saprolite.

C4 -- 497 to 513 cm, strong brown (7.5YR 5/6) massive unoxidized saprolite.

C5 -- 536 to 549 cm, dark yellowish brown (10YR 4/6) massive unoxidized saprolite.

Fieldnotes:

Described from spaded pit (deep samples >254 cm augered by hand).

Quartz vein extended from bottom of pit upward to surface.

Weathering rinds on vein quartz are 2-4 mm.

Sampled and described by SBF, LWZ, Larry West, Louie Frost, Tom Macfie, Jody Palmer, and Larry LaPlante.

PEDON: GA 2

Location: Oglethorpe Co, GA, Crawford area (Huey Sanders Farm); From the intersection of Hwy 78 and Hwy 22 in Lexington, go W on Hwy 78 for 1.75 mi; Turn right at Crawford city limit sign; Go 1.0 mi to stop sign and turn right; Go 0.6 mi and park; Pasture on right-hand side of road upslope from about a 1 AC pond.

Elevation: 227 m (745 ft)

Aspect: 88° NE

Slope: 1-2% (water flow direction); 1% (along ridge)

Geomorphic: Upland ridgetop summit; linear contour, convex downslope

Relief: Relief is similar to the area around pedon GA 1. Maximum relief is about 30 m with most areas averaging about 18m. Low-order streams occupy drainage basins of about 1 to 8 km². Drainage divides are relatively broad, flat uplands, and hillslopes grade gently into the main valleys at about 20%. Local perennial streams have gradients of about 0.008 m/m, ranging to about 0.15 m/m headward (near the site).

Parent Material: Saprolite and residuum weathered from Elberton Granite.

Vegetation: Pasture - fescue, broomsedge - second year fallow

Classification: Typic Rhodudult, clayey, oxidic, thermic

Description:

Ap --- 0 to 8 cm, dark brown (7.5YR 4/4) fine sandy loam; weak fine granular structure; very friable; many fine and medium roots; clear smooth boundary.

BA -- 8 to 18 cm, red (2.5YR 4/6) sandy clay loam moderate fine subangular blocky structure; friable; common fine and medium roots; gradual smooth boundary.

Bt1 -- 18 to 30 cm, red (2.5YR 4/6) clay; strong medium subangular blocky structure; firm; common thick continuous clay films on ped faces; few fine mica flakes; few fine roots; clear wavy boundary.

Bt2 -- 30 to 69 cm, red (2.5YR 4/8) clay; weak coarse prismatic parting to moderate medium subangular blocky structure; friable; common thin discontinuous clay films on ped faces; common fine flakes of mica; gradual smooth boundary.

BC1 -- 69 to 152 cm, red (2.5YR 4/6 - 4/8) clay; common medium distinct strong brown (7.5YR 5/6) mottles; massive in place parting to weak medium subangular blocky structure; friable; common thin discontinuous clay films on ped faces and relict rock joints; many fine flakes of mica; gradual smooth boundary.

BC2 -- 152 to 229 cm, red (2.5YR 4/6 - 4/8) clay loam with common medium distinct strong brown (7.5YR 5/6) and yellowish brown (10YR 5/4) mottles; massive in place with platy relict rock structure parting to weak medium subangular blocky structure;

friable; few thin discontinuous clay films along rock joints; many fine flakes of mica; gradual wavy boundary.

CB -- 229 to 240+ cm, red (2.5YR 4/6 - 4/8) clay loam with common medium distinct strong brown (7.5YR 5/6) and yellowish brown (10YR 5/4) mottles; mostly massive with platy clods parting along relict rock joints, weak medium subangular in places; friable; few thin discontinuous clay films along vertical joints and ped faces; many fine mica flakes.

C1 -- 254 to 270 cm, red (2.5YR 4/8) massive, oxidized saprolite.

C2 -- 318 to 336 cm, yellowish red (5YR 5/8) massive, oxidized saprolite.

C3 -- 368 to 384 cm, strong brown (7.5YR 5/6) massive, oxidized saprolite.

C4 -- 457 to 473+ cm, light yellowish brown (10YR 6/4) massive, unoxidized saprolite.

Fieldnotes:

Described from spaded pit (deep samples >270 cm augered by hand).

Auger stopped by rock floater at 4.7 m.

Sampled and described by SBF.

PEDON: AL 1

Location: Coosa Co. Alabama; Kelleyton area (Bubba Holman Farm); 506m N and 122m E of SW corner Sec 25, T. 23 N., R. 20 E. (Kellyton Quad).

Elevation: 222 m (730 ft)

Aspect: 60° SE

Slope: 2% (water flow direction); 1% (along ridge)

Geomorphic: Upland ridgetop summit; convex contour, linear downslope

Relief: Maximum relief is about 24 m with average relief of about 12 m; Low-order streams occupy drainage basins of about 1 to 10 km². Drainage divides are relatively broad, flat uplands, and hillslopes grade gently into the main valleys at about 20%. Local perennial streams have gradients of about 0.005 m/m, ranging to about 0.1 m/m headward (near the site).

Parent Material: Saprolite and residuum weathered from Elkahatchee Quartz Diorite

Vegetation: Hayfield - fescue, clover

Classification: Typic Rhodudult, clayey, oxidic, thermic

Description:

Ap1 --- 0 to 5 cm., dark reddish brown (5YR 3/4) loam; weak fine granular structure; very friable; many fine roots; abrupt smooth boundary.

Ap2 --- 5 to 13 cm., yellowish red (5YR 4/6) loam; weak fine granular structure; very friable; many fine roots; clear smooth boundary.

Bt1 -- 13 to 60 cm., red (2.5YR 4/6 - 4/8) clay; moderate medium subangular blocky structure; firm; common thin discontinuous clay films on ped faces; common fine and very fine roots; few fine mica flakes; gradual smooth boundary.

Bt2 -- 60 to 100 cm., red (2.5YR 4/8) clay with common medium distinct yellowish red (5YR 5/8) mottles; weak coarse parting to moderate medium subangular blocky structure; friable; common discontinuous clay films on ped faces; few very fine roots; common fine mica flakes; gradual wavy boundary.

BC1 -- 100 to 140 cm., red (2.5YR 4/8) clay; common medium distinct strong brown (7.5YR 5/6) mottles; massive in place parting to weak medium subangular blocky structure; friable; few thin discontinuous clay films on ped faces and relict rock joints; few very fine roots; common fine flakes of mica; clear irregular boundary.

BC2 -- 140 to 228 cm., red (2.5YR 4/6 - 4/8) clay loam; common medium distinct strong brown (7.5YR 5/6) mottles; mostly massive in place with platy clods breaking out along old rock joints, weak fine subangular blocky structure in places; friable; few

thin discontinuous clay films along rock joints; many fine flakes of mica; gradual smooth boundary.

CB -- 228 to 270 cm., red (2.5YR 4/6 - 4/8) clay; common medium distinct strong brown (7.5YR 5/6) and light yellowish brown (10YR 6/4) mottles; blocky structure, massive in pockets with platy relict rock structure; friable; patchy discontinuous clay films along joints; many fine flakes of mica; gradual boundary.

C1 -- 351 to 367 cm, yellowish red (5YR 4/6) massive, unoxidized saprolite.

C2 -- 384 to 400 cm, strong brown (7.5YR 4/6) massive, unoxidized saprolite.

C3 -- 462 to 478 cm, strong brown (7.5YR 4/6) massive, unoxidized saprolite.

C4 -- 502 to 518 cm, strong brown (7.5YR 4/6) massive, unoxidized saprolite.

C5 -- 549 to 565+ cm, dark yellowish brown (10YR 4/4) massive, unoxidized saprolite.

Fieldnotes:

Described from spaded pit (deep samples >350 cm augered by hand).

Bottom of Bt2 horizon appears to truncate quartz vein which descends at a steep angle from a depth of about 100 cm to the bottom of the pit. Vein could not be traced to the top of the pit to confirm uniformity of parent material.

BC2 horizon has two zones of saprolite/rotten rock. One is dominantly very pale brown (10YR 7/4) fine sandy loam, and the other is brown (7.5YR 5/4) fine sandy loam.

Ap1 and Ap2 horizons sampled together.

Sampled and described by SBF and Lawrence McGhee.

PEDON: AL 2

Location: Coosa Co. Alabama; Keyno area (J.T. Hallman Farm); 671m N and 46m E of SW corner Sec 25, T. 23 N., R. 20 E. (Kellyton Quad).

Elevation: 235 m (770 ft)

Aspect: 88° SE

Slope: 3% (water flow direction); 2% (along ridge)

Geomorphic: Upland ridgetop summit; convex contour, linear downslope

Relief: Landscape is somewhat more highly dissected than the area surrounding pedon AL

1. Maximum relief is about 30 m with most areas averaging about 18m Low-order streams occupy drainage basins of about 1 to 8km². Drainage divides are relatively broad, flat uplands, and hillslopes grade gently into the main valleys at about 20%. Local perennial streams have gradients of about 0.008 m/m, ranging to about 0.15 m/m headward (near the site).

Parent Material: Saprolite and residuum weathered from Elkahatchee Quartz Diorite; parent material reflects a local, gravelly compositional band of highly resistant and spessartine (Mn) and almandine (Al) garnet.

Vegetation: Pasture - fescue, broomsedge - second year fallow

Classification: Typic Rhodudult, clayey, oxidic, thermic

Description:

Ap --- 0 to 10 cm., dark reddish brown (5YR 3/4) loam to clay loam; weak medium granular structure; very friable; many fine and medium roots; clear smooth boundary.

Bt1 -- 10 to 55 cm., dark red (2.5YR 3/6) clay to gravelly clay (about 15% black concretions or euhedral garnets of <4 mm across); moderate medium subangular blocky structure; firm; few thin discontinuous clay films on ped faces; common fine and medium roots; gradual smooth boundary.

Bt2 -- 55 to 108 cm., red (2.5YR 4/6 - 4/8) clay to gravelly clay (about 10 to 15% coarse fragments <4 mm across); weak coarse prismatic parting to strong medium subangular blocky structure; firm; common thick continuous clay films on ped faces; few fine mica flakes; few fine roots; clear wavy boundary.

Bt3 -- 108 to 155 cm., red (2.5YR 4/6 - 4/8) clay; common medium distinct strong brown (7.5YR 5/6) mottles; weak coarse prismatic parting to moderate medium subangular blocky structure; friable; common thin discontinuous clay films and dark (Mn?) coatings along prism faces; common fine flakes of mica; clear wavy boundary.

Bt4 -- 155 to 170 cm., red (2.5YR 4/6 - 4/8) clay; common medium distinct strong brown (7.5YR 5/6) mottles; weak medium subangular blocky structure; friable; common thin discontinuous clay films and many dark (Mn?) coatings along prism faces; many fine flakes of mica; gradual smooth boundary.

Bt5 -- 170 to 228 cm., red (2.5YR 4/6 - 4/8) clay; common medium distinct strong brown (7.5YR 5/6) mottles; weak medium subangular blocky structure, massive in pockets with platy relict rock structure; friable; few thin discontinuous clay films and many dark (Mn?) coatings along peds and joint faces; many fine flakes of mica; gradual wavy boundary.

BC -- 228 to 260+ cm., red (2.5YR 4/6 - 4/8) clay loam with common medium distinct strong brown (7.5YR 5/6) and yellowish brown (10YR 5/4) mottles; mostly massive with platy clods parting along relict rock joints, weak medium subangular in places; friable; few thin discontinuous clay films along vertical joints and ped faces; many fine mica flakes; gradual boundary.

CB1 -- 307 to 323 cm, reddish brown (2.5YR 4/4) massive, oxidized saprolite.

CB2 -- 368 to 384 cm, reddish brown (5YR 4/4) massive, oxidized saprolite.

C1 -- 429 to 445 cm, yellowish red (5YR 4/6) massive, oxidized saprolite.

C2 -- 454 to 470 cm, dark yellowish brown (10YR 4/4) massive, unoxidized saprolite.

C3 -- 470 to 486 cm, dark yellowish brown (10YR 4/4) massive, unoxidized saprolite.

Fieldnotes:

Described from spaded pit (deep samples >260 cm augered by hand).

Auger stopped by rock floater at 4.9 m.

Bottom of Bt2 horizon appears to truncate quartz vein which descends at a steep angle from a depth of about 100 cm to the bottom of the pit.

Dark (Mn?) stains are oriented along the vein. Vein could not be traced to the top of the pit to confirm uniformity of parent material.

Rhodic

Sampled and described by SBF.

Appendix B. In Hindsight....

After having completed this work, there are several things that I would do differently, if I had this project to do over again. In general, I would tackle a smaller project with a narrower focus, trying to steer clear of the sweeping objectives.

I would try to sample more carefully and in greater detail. Sampling deeper, and in narrower increments would have permitted more detailed analysis of the initial stages of weathering as well as determination of the nature of transitional phases. Better sampling for bulk density analysis would have enhanced mass balance calculations, which can be quite sensitive to variations in this parameter.

Obtaining data on the composition of soil solution at each sampled depth would also provide considerable insight into the thermodynamic stability of minerals present in the system. Given the fact that different primary minerals often weather to a common secondary product in Piedmont soils, it is crucial to unravel the geochemical factors governing mineral stability and weathering mechanisms.

Age control is a critical element that was unfortunately omitted from this study, although it was central to the original study objective. Despite the fact that I obtained some interesting results, it remains unclear as to whether climate or time is the primary factor in silicate mineral weathering on the Piedmont. Successful attempts to constrain the ages of Piedmont soils and geomorphic surfaces in the future will probably involve measurement of cosmogenic isotopes produced *in situ* on rock surfaces.

There are clearly interesting questions that have arisen from this work. The assessment of whether biotite weathering is a temperature-dependent process is one

fruitful area for future consideration. A link between specific mechanisms of biotite weathering and temperature may provide a direct means to assess continental paleoclimates. Finally, further study of the processes leading to saprolite formation from weathered rock needs to be undertaken to determine, by measurement and not by speculation, whether saprolite is truly an isovolumetric weathering product as the literature infers. If any volume expansion occurs during this process as suggested in Chapter V, age estimates for the Piedmont surface based on regolith thickness may be highly inflated.

Vita

Part I

Contrary to what some people may think, I was not born in a log cabin in 1874, nor have I had my beard since birth. I appeared in this world in a suburban environment on December 23, 1954, in the days before 'biotechnology' and am thus reasonably assured of not being the product of artificial ingredients. Requiring another suitable outlet for my energies after having quit the trombone in the fourth grade (my arm was too short to reach the 'sixth position'), I became somewhat rebellious and soon took up permanent residence at the desk directly in front of the teacher's watchful eye, where masking tape was frequently and liberally applied to my mouth. I occasionally attended high school in West Orange, NJ, and matriculated in 1972 in spite of my best efforts. After finding city/suburban life beginning to fail me, I relocated to Tennessee in the Fall of 1974, where I gained a renewed appreciation for green grass, the trees, the rocks, and the hills—recurring theme in my life. It was also during this time that I developed a deep and abiding love for southern Appalachian mountain music (some say that there is a fine line between playing this 'old time' music and not being able to play at all), possibly because the necks of a fiddle and banjo are considerably shorter than that of a trombone.

I acquired a genuine interest in the study of soils and landscapes at the University of Tennessee at Knoxville, where, under the guidance of Dr. Frank F. Bell (a.k.a. F2), I was elevated to the status of 'serious student' and where I graduated with a B.S. degree in

Plant and Soil Science in 1980. I subsequently began a career as a field soil scientist, working on progressive soil surveys throughout Tennessee for the USDA-Soil Conservation Service (now Natural Resources Conservation Service). I enrolled in VPI&SU during the summer of 1984 in order to pursue my academic interests, enhance my professional skills, and after learning that the leading cause of retirement among field soil scientists is disability resulting from back injury! In the Fall of 1984, I accepted a position as soil scientist with the Crop and Soil Environmental Sciences Department (formerly the Dept. of Agronomy), VPI&SU, with the responsibility of running the soil mineralogy lab where I remain, committed to pursuing the ‘terminal degree’, otherwise known, in the words of G. Norman White, as ‘the elusive Ph.D.’

Part II

Well, here I am — I finally finished this sucker and, yes, it has been elusive. It’s been 11 years since I came here to Virginia Tech. I now look for some meaningful employment, hopefully in a position that would permit me to take advantage of what I’ve always thought were somewhat unique, if not marketable, skills. At the present time, things look rather bleak on the job front. After being laid off in June of 1994 from the job that I had for 10 years, and after the subsequent graduate student wages ran out this past month, I begin work this week painting and remodeling houses — a job and a salary that I might have had 20 years ago.

It’s hard not to be disillusioned and even demoralized when I think that this whole ordeal might turn out to be no more than an exercise in intellectual stimulation. The job market is so poor right now, due largely, in my opinion, to conservative-minded

legislators at both the state and national level who seem to view higher education and scientific inquiry as poor investments, that it appears unlikely that I would be able to find even a Post-Doctoral position a position that would permit me to work in my areas of expertise and interests.

That's unfortunate because I feel that I am just now figuring out how to do this type of work. I'd like to take advantage of these skills and make some contribution to what we already know about soils, weathering, land use, and the environmental implications of our post-technological society. On a simpler level, at this point in my life, I'd just like to have a job that I enjoy with a modest income, opportunity for occasional travel, and the beginnings of a retirement account.

On the other hand, maybe I'll just ride my bike.