

Mineralogy of a Chronosequence Formed in New River Alluvium¹

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ABSTRACT

The mineralogy of four soils (T₀, T₁, T₂, T₃), formed in progressively older New River alluvium, was studied with the assumption that trends could be attributed to age differences. Particle size was determined by pipette; free Fe by citrate-dithionite (CD) extraction; sand mineralogy by petrographic techniques, heavy liquid separation, and X-ray diffraction (XRD); and silt and clay mineralogy by XRD and thermal analysis. Clay illuviation and percent CD-Fe increase with age. An increase in CD-Fe with depth suggests Fe illuviation. Quartz, feldspar, mica, and amphibole are the major sand-sized minerals in T₀ and T₁, with slightly less amphibole in T₁. Pre-dominance of quartz and resistant heavy minerals in T₂ and T₃ suggests a wide time gap between T₁ and T₂. Trends in silt mineralogy parallel those of sand. Clay mineralogical data show an age progression from mica and vermiculite to hydroxy interlayered vermiculite to kaolinite. Kaolinite appears to form very slowly from other layer silicates, but whole-soil trends suggest that it may form more rapidly from sand and silt-sized feldspar. With time, hydroxy interlayered vermiculite becomes the dominant clay-sized mineral in Ap horizons while kaolinite and vermiculite increase with depth. The depth trend for kaolinite is attributed to eluviation and slow replenishment in the surface, and for the other minerals to decreasing weathering intensity with depth. Whole-soil data reflect illuviation and reveal a concentration of quartz in the surface with time. Comparison of mineralogical trends suggests that stability is reached sooner in coarser than in finer size fractions.

Additional Index Words: mineral weathering, whole-soil mineralogy, soil genesis.

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THE CHRONOSEQUENCE (Jenny, 1941) makes possible the isolation of time as the dominant influence in soil genesis. Mineralogical differences between soils can be attributed to time if the effects of other soil-forming factors have been relatively constant.

Some studies of time-related mineralogical changes have dealt with the >2- μ m fraction. Mokma et al. (1973b) noted that quartz increased with age in the sand and silt fractions of a New Zealand chronosequence, while mica, feldspar, and amphibole decreased or remained stable. They inferred that an initial increase of mica and decrease of quartz in the silt fraction from 0 to 250 years reflected a more rapid breakdown of mica particles from sand to silt. The presence of interstratified silt-sized mica and vermiculite in the 250-year-old soil was interpreted as evidence for a rapid transformation of mica to vermiculite. Carroll (1959) noted a higher concentration of sand-sized Fe minerals and resistant translucent heavy minerals

in older terrace soils of the Middle River drainage basin. Sidhu and Gilkes (1977) detected no trends in the sand and silt fractions in comparing alluvial soils of different ages on the Indo-Gangetic Plain.

Trends in time-related clay mineral transformations have been reported for a diversity of climates and parent materials. From their study of increasingly older glacial drift and residual soils derived from the same geological formation, Novak et al. (1971) formulated a sequence in which mica weathers to vermiculite and mixed-layer mica-vermiculite, which in turn weather to kaolinite. A similar trend was noted by Mokma et al. (1973b) except that they detected no increase of kaolinite with age up to 22,000 years. Others have confirmed the tendency for kaolinite to become dominant given adequate time (Brewer and Walker, 1969; Mokma et al., 1973a). However, in very old soils there is the possibility of paleoclimatic influence. Pedogenic kaolinite has been associated with intense leaching (Jackson, 1965), and is not likely to be a significant soil constituent in arid climates (Ahmad et al., 1977). Some studies have failed to reveal trends in clay minerals with age despite evident morphological differentiation among profiles (Sidhu and Gilkes, 1977; Torrent, 1976).

Genetic processes do not occur independently within selected size fractions, and a more thorough understanding of these processes may be attained by considering sand, silt, clay, and whole-soil trends. The purpose of this study was to investigate, on a whole-soil basis, the effects of time on the mineralogy of a New River terrace chronosequence.

MATERIALS AND METHODS

Study Area and Soils

The New River has been established as one of the oldest streams in North America (Fridley, 1950; Janssen, 1952; Rhodehamel and Carlston, 1963). It descends from the crystalline rocks of the Blue Ridge in southwest Virginia and flows roughly northward across the sedimentary rocks of the Appalachian Valley and Allegheny Plateau before discharging into the Ohio River (Fig. 1). The entrenched antecedent channel traverses resistant sandstone ridges of the Appalachians. Terraces are most extensive where the channel is least constricted. They become increasingly dissected with greater elevation, though a few gently sloping remnants of the older surfaces remain.

The chronosequential transect used in this study lies approximately 74 km (56 linear km) downstream from where the New River passes from crystalline to sedimentary rocks (Fig. 1). However, the site is only about 32 km (16 linear km) downstream from the confluence of the New River with a major tributary (Little River) which provides a fresh influx of crystalline alluvium. Little River alluvium is derived from a diversity of rocks including granites, granodiorites, and amphibole-rich foliates (Dietrich, 1959). Sedimentary rocks upstream from the study site include limestones, shales, siltstones, and sandstones from Precambrian to Mississippian age (Butts, 1938).

The chronosequence consists of a floodplain soil (T₀) and three successively older terrace soils (T₁, T₂, and T₃), at elevations of approximately 504, 510, 558, and 588 m above sea level, respectively. These soils have evolved under a udic moisture regime and a mesic temperature regime (Soil Survey Staff, 1975). All pedons sampled are well drained and lie on stable, nearly level to gently sloping, contiguous terrace surfaces forming a sequence

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roughly perpendicular to the river channel. The distance from T0 to T3 is approximately 1,400 m. Relative soil ages were inferred according to the principle of ascendancy (Ruhe, 1969). Based on regional stratigraphic and geomorphic relationships, Howard² estimates that high terraces on the James, Roanoke, and New Rivers may be as old as late Miocene. Relative soil ages are reflected in classification (Table 1), with taxonomic classes ranging from Mollisol to Ultisol (Soil Survey Staff, 1975). Horizonation and redness of hue (Table 2) increase with age.

Physical, Chemical, and Mineralogical Analyses

Samples from all genetic horizons were air-dried, ground, and sieved to remove coarse fragments (>2 mm). Particle size was determined by pipette analysis, exchangeable bases by sum of Ca, Mg, and K extracted with 1N NH₄OAc adjusted to pH 7, extractable acidity by back titration of unreacted extracts of BaCl₂-TEA adjusted to pH 8.2, and free Fe by citrate-dithionite extraction (Holmgren, 1967). Cation exchange capacity (CEC) was computed as the sum of exchangeable bases and extractable acidity. The pH of each horizon was determined from a 1:1 soil/water sample.

Samples for mineralogical analysis were pretreated with H₂O₂-NaOAc, adjusted to pH 5.0 and citrate-dithionite-bicarbonate to remove organic matter and oxide coatings, respectively. The sand-sized fraction was separated by wet-sieving. The silt and clay fractions were separated by centrifugation using dilute Na₂CO₃ adjusted to pH 9.5 as a dispersant. The <40-mesh sand fraction used for heavy mineral separations was wet-sieved from composite control section samples. Separations were made using 1,1,2,2,tetrabromoethane (sp gr = 2.95).

Oriented mounts of the clay fraction were prepared by depositing 250 mg of sample on a millipore filter, saturating with Mg or K, washing free of salts, and glycolating the Mg-saturated samples. The membrane, while still moist, was placed clay side down on a glass slide and transferred with a gentle rolling pressure of a glass rod (Drever, 1973). X-ray diffraction patterns of air-dry, 105, 300, and 550°C heated samples were obtained using a Diano XRD-8300AD X-ray diffractometer equipped with a graphite crystal monochromator, PDP-8 computer, and a print-out. The samples were scanned at 2° 2θ per min using CuKα radiation. Silt mineralogy was quantitatively estimated by integrating peak areas from X-ray diffractograms of randomly oriented powder mounts.

Thermal analysis of clay fractions was accomplished with a Dupont 990 Differential Scanning Calorimeter. Samples were heated from 50 to 700°C in a N₂ atmosphere at a rate of 50°C/min. Quantitative estimations of gibbsite and kaolinite were obtained by measuring the areas under the respective endothermic peaks at 280 and 520°C, and comparing with standard curves derived by regression analysis for Reynolds synthetic gibbsite RH-31F and poorly crystalline Georgia kaolinite obtained from the University of Missouri source clay minerals repository. Mica contents were calculated from total elemental analysis assuming 10% K₂O equaled 100% mica. Quantitative estimations of other clay-sized minerals were obtained by proportioning integrated peak areas of the appropriate X-ray diffractograms, using kaolinite as an internal standard and assuming the minerals detected equaled 100%.

Sand-sized minerals were identified using a Zeiss polarizing microscope. Grain mounts were prepared from the <40-mesh fraction, which is dominant and representative of total sand in these samples. Frequency per hundred grains was determined using the line count method (Galehouse, 1971). Random powder mounts were prepared from ground heavy mineral subsamples and XRD was used to further substantiate identity of heavy minerals.

RESULTS AND DISCUSSION

Particle-size data (Table 1) and subsoil morphology (clay films, Table 2) reflect an increase in clay illuviation with age. The uniform texture with depth and the absence of visible clay films suggest that T0 has undergone little weathering. T1 exhibits only a min-

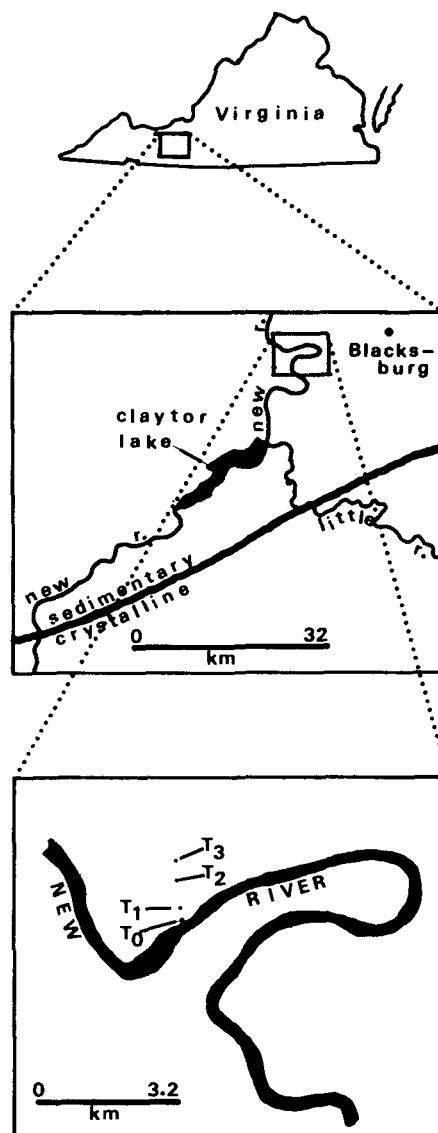


Fig. 1—Site location and boundary between crystalline and sedimentary rocks.

imal clay increase in the subsoil, but the presence of discontinuous clay films on ped faces confirms clay illuviation. T2 and T3, in contrast to the younger soils, show a substantial increase in clay with depth and have thick, continuous clay films in the B horizons. Particle-size discontinuities within T1, T2, and T3 profiles can be inferred from an abrupt increase in sand and coarse fragments in T1 and the presence of resistant cobble and gravel strata in the clay matrix of the T2 and T3 B horizons. These depositional variations within pedons preclude definite conclusions about age-related in situ physical weathering trends among pedons.

Citrate-dithionite extractable (CD) Fe (Table 1) is greater for T2 and T3 than for T0 and T1 in the subsoil. The Fe released from weathering of primary minerals such as amphibole and biotite would be expected to persist and accumulate over time in oxidized soil systems. The increase in CD-Fe with depth in T2 and T3 suggests that Fe has eluviated.

²J. Howard. 1978. Geologic and pedologic studies of the James, Roanoke and New River Basins, Virginia. Unpublished report, VPI & SU, Blacksburg, Va.

Table 1—Classification and selected physical and chemical properties of soils studied.

Horizon	Depth	Sand	Silt	Clay	pH	Σ Exch. bases	CEC	BaCl ₂ -TEA acidity	KCl acidity	CD-Fe†
	cm	%				meq/100 g soil				%
Cumulic Hapludoll (fine-loamy, mixed, mesic) (T0)										
Ap	0-23	36.8	42.3	20.9	5.7	8.0	16.9	8.9	-	2.2
C	23-30	45.3	32.2	22.5	6.1	6.9	13.7	6.8	-	2.7
A1	30-58	41.4	36.1	22.5	6.3	11.6	18.7	7.1	-	1.2
B21	58-103	42.9	36.2	20.9	6.4	15.1	21.1	6.0	-	1.1
B22	103-135	39.5	37.5	23.0	6.5	16.2	21.3	5.1	-	1.3
B23	135-200	44.3	34.4	21.3	6.6	14.9	19.7	4.9	-	1.2
C	200-250	47.5	32.8	19.7	6.7	14.6	18.9	4.3	-	1.0
Ultic Hapludalf (fine-loamy, mixed, mesic) (T1)										
Ap	0-35	40.2	39.4	20.4	6.1	7.8	16.6	8.8	-	1.4
B21t	35-50	37.5	39.7	22.8	5.5	4.2	12.6	8.4	0.5	1.6
B22t	50-65	38.1	38.7	23.2	5.4	3.8	11.8	8.0	0.5	1.3
B23t	65-80	41.3	34.2	24.4	5.5	4.5	11.3	6.8	0.4	1.6
B24t	80-95	30.4	46.0	23.6	5.5	4.5	11.4	6.9	0.3	1.5
IIB25t	95-115	51.8	25.4	22.9	5.5	4.5	11.3	6.8	0.3	2.1
IIB3t	115-138	56.1	21.5	22.5	5.3	5.0	11.4	6.4	0.2	2.2
IIC	138-185	48.8	28.7	22.5	5.5	5.6	12.0	6.4	0.2	2.3
Typic Hapludult (clayey, mixed, mesic) (T2)										
Ap	0-13	32.2	53.1	14.8	5.9	6.3	13.3	6.9	-	1.1
A2	13-38	30.2	46.2	23.6	5.0	1.8	8.4	6.6	1.7	1.6
B21t	38-63	24.3	35.1	40.6	4.8	3.5	15.1	11.7	3.9	2.2
IIB22t	63-88	24.3	29.4	46.3	5.0	3.1	17.3	14.2	5.4	3.0
IIB23t	88-120	21.5	22.3	56.2	5.1	2.2	19.3	17.0	7.1	3.9
IIB24t	120-145	15.8	27.2	56.9	4.9	1.2	18.7	17.4	8.2	4.1
IIB3t	145-180	23.5	33.1	43.5	4.9	1.1	15.2	14.2	6.2	4.1
IIIC	180-278	30.7	43.5	25.9	4.8	0.5	9.3	8.8	3.7	2.8
Typic Paleudult (clayey, kaolinitic, mesic) (T3)										
Ap	0-15	51.8	31.5	16.8	5.4	2.9	11.3	8.4	0.4	0.9
B21t	15-33	44.7	27.7	27.7	4.8	2.7	9.3	7.5	1.5	1.7
B22t	33-53	31.5	10.3	58.2	4.7	2.9	14.5	11.7	3.4	3.6
B23t	53-75	28.8	10.6	60.6	5.0	1.7	13.4	11.7	4.0	3.9
B24t	75-95	35.3	6.9	57.8	4.9	1.0	12.4	11.4	4.2	4.2
B25t	95-120	37.5	4.6	57.9	5.0	0.7	11.1	11.5	3.8	4.2
IIB26t	120-143	39.5	6.7	53.9	5.4	0.6	12.3	10.5	4.1	3.9
IIIB27t	143-158	39.8	6.2	54.0	5.5	0.5	11.0	10.6	3.9	3.6

† Citrate-dithionite extractable Fe.

Sand Mineralogy

Petrographic examination revealed the presence of quartz, feldspar, mica (biotite and muscovite), and heavy minerals (primarily amphibole) in the sand fraction of T0 (Table 3). T1 is mineralogically similar to T0 except for a slight increase in quartz and decrease in heavy minerals. Feldspar and mica fluctuations within pedons preclude a discernable trend for these minerals between T0 and T1. Feldspar, mica, and heavy minerals sharply decrease from T1 to T2 while quartz increases proportionately. Little change is evident from T2 to T3 except for a slight increase in heavy minerals.

The heavy mineral decrease from T0 to T1 suggests that amphibole, the major heavy mineral constituent of these two soils, is the most weatherable mineral present. Feldspar and mica prove more stable than amphibole, but eventually are removed by weathering leaving quartz to predominate in the two older soils. The increase in heavy mineral weight percentage from T2 to T3 (Table 3) is attributable to a concentration of resistant species which include ilmenite, zircon, tourmaline, and kyanite. An increase in ilmenite with soil age has also been detected by Carroll (1959). Sand-sized mineral trends in the present study are consistent with other works estimating relative resistance among minerals (Jackson and Sherman, 1953; Mitchell, 1975).

The near absence of weatherable minerals such as mica, feldspar, and amphibole in T2 and T3, in contrast to the abundance of these minerals in the younger alluvium, suggests a broad time gap between T1 and T2 geomorphic surfaces. Similar gaps between younger and older terraces have been noted on the James, Shenandoah, and Roanoke Rivers in Virginia.³ From landscape features it can also be inferred that T1 is much closer in age to T0 than T2. The T1 surface is only slightly above flood level and has not undergone significant dissection. In contrast, there is a conspicuous scarp between T1 and T2, and the T2 surface is moderately dissected. The time gap between T2 and T3 may be even greater than the gap between T1 and T2, considering the differences in elevation, the increasingly prominent dissection, and the formation of sinkholes on the T3 geomorphic surface. However, the sand fraction does not reflect age differences among these older soils because of the mineralogical stability attained once quartz and resistant heavy minerals predominate.

An alternative explanation for the contrast in sand mineralogy between T1 and T2 would be that parent material was radically different for the two older soils. However, considering the long-established course of the New and Little Rivers, both with head waters extending well into the Blue Ridge crystalline rocks, the source of alluvium for the high terraces probably

Table 2—Summary of selected morphological properties of soils studied.

Properties	T0	T1	T2	T3
Diagnostic horizon	Mollic	Weak argillic	Strong argillic	Strong argillic
Subsoil colors	7.5YR 3/2; 10YR 3/3	7.5YR 4/4	7.5YR 5/6; 10YR 4/4; 5YR 4/6; 2.5YR 3/4	2.5YR 3/6, 4/6; 10YR 6/6
Subsoil structure	Weak subangular blocky	Medium subangular blocky	Medium to strong subangular blocky	Strong angular blocky
Clay films	None	Common fine to medium discontinuous	Many medium to thick continuous	Many medium to thick continuous

Table 3—Mineralogy of the sand fraction of samples examined.†

Soil	Horizon‡	Depth, cm	Qz	Fd	Frequency/100 grains‡				Z/T	HM, wt.
					Mica	HM	Opaque	Other		
T0	Ap	0-23	48 ± 3.2	7 ± 1.3	23 ± 2.4	13 ± 1.8	4 ± 1.2	3 ± 1.0	2.5	12.2
	CS	25-100	52 ± 3.2	13 ± 1.8	19 ± 2.0	10 ± 2.0	1 ± 0.8	5 ± 1.4		
	B21	58-100	50 ± 3.2	9 ± 1.7	25 ± 1.8	11 ± 0.8	1 ± 0.8	3 ± 1.0		
	C	200-250	50 ± 3.0	17 ± 2.2	21 ± 2.4	12 ± 1.8	TR	TR		
T1	Ap	0-35	69 ± 2.6	12 ± 1.7	9 ± 1.7	7 ± 1.3	2 ± 0.7	1 ± 0.7	2.3	7.7
	CS	35-85	61 ± 2.7	12 ± 1.7	18 ± 2.2	4 ± 1.3	2 ± 1.7	2 ± 0.7		
	B23t	65-80	54 ± 3.4	12 ± 2.2	24 ± 2.4	6 ± 1.3	2 ± 0.8	3 ± 0.8		
	IIC	138-185	63 ± 2.7	19 ± 2.2	10 ± 1.7	6 ± 1.3	2 ± 0.6	TR		
T2	Ap	0-13	96 ± 1.6	TR	TR	TR	1 ± 0.8	2 ± 0.8	2.7	1.0
	CS	13-63	97 ± 0.9	TR	TR	TR	2 ± 0.9	TR		
	IIB22t	63-88	99 ± 0.8	TR	TR	TR	TR	TR		
	IIB23t	88-120	91 ± 2.2	1 ± 0.8	2 ± 0.9	TR	TR	5 ± 1.1		
T3	Ap	0-15	99 ± 0.9	TR	TR	TR	1 ± 0.9	TR	2.5	2.3
	CS	15-65	97 ± 0.8	-	TR	TR	2 ± 0.8	TR		
	B24t	75-95	96 ± 1.6	TR	TR	TR	TR	TR		
	IIB27t	143-158	93 ± 1.4	1 ± 0.8	3 ± 0.8	3 ± 0.8	TR	TR		

† Qz = quartz, Fd = feldspar, HM = translucent heavy minerals, Opaque = opaque grains or materials, Other = rock fragments and unidentified minerals, HM, wt. = total heavy minerals (weight percentage of <40-mesh sand) separated by 1,1,2,2-tetrabromoethane (sp gr = 2.95), and Z/T = zircon to tourmaline ratio.

‡ Counting error (± one standard deviation) obtained from nomograph (Folk, 1965).

§ CS = composite mineralogical control section (Soil Survey Staff, 1975).

was similar to the source for the present flood-plain. The presence of Blue Ridge derived quartz and quartzite clasts in the oldest terrace soils and a relatively constant zircon/tourmaline ratio (Table 3) corroborate this view. Reworked material from T3 could possibly have been a major alluvial source for T2, making these soils age-correlative with respect to sand. However, the crushable nature of quartzite clasts in T3 relative to T2 suggests a more prolonged or intense leaching of Si in the highest terrace.

Silt Mineralogy

The silt fractions of T0 and T1 comprise mainly quartz, feldspar, mica, hydroxy-interlayered vermiculite, and kaolinite (Table 4). Trace amounts of amphibole, vermiculite, and interstratified minerals are also present in these younger soils. Age trends are apparent when T0 and T1 are compared with T2 and T3. Feldspar, mica, hydroxy interlayered vermiculite and kaolinite are less abundant in the two older soils while quartz is more concentrated. Mica increases with depth in T3, reflecting diminished weathering intensity in the lower profile. The trends for quartz, feldspar, and mica parallel those noted for the sand fraction and are likewise in accord with other works establishing the relative resistance of these minerals (Jackson and Sherman, 1953; Mitchell, 1975).

Clay Mineralogy

Vermiculite and mica decrease with age from T0 to T3 (Table 5), and the former increases with depth in

Table 4—Mineralogy of the silt (50-2 μm) fraction of the samples examined.†

Soil	Horizon‡	Depth, cm	KK	HIV	M	QZ	GI	FD	%	
T0	Ap	0-23	10	-	8	73	2	7		
	CS	25-100	5	5	10	75	-	-		
	B21	58-103	3	2	12	75	1	7		
	C	200-250	3	10	5	74	1	7		
T1	Ap	0-35	-	-	5	85	-	10		
	CS	35-85	10	13	10	65	-	2		
	B23t	65-80	4	10	8	72	1	5		
	IIC	138-185	5	15	8	61	1	10		
T2	Ap	0-13	-	-	-	98	-	2		
	CS	13-63	-	-	2	97	-	1		
	IIB22t	63-88	-	-	-	97	-	3		
	IIB23t	88-120	3	-	-	95	-	2		
T3	Ap	0-15	3	-	-	97	-	-		
	CS	15-65	-	-	5	94	-	1		
	B24t	75-95	-	-	8	91	-	1		
	IIB27t	143-158	5	8	12	67	2	5		

† KK = kaolinite, HIV = hydroxy interlayered vermiculite, M = mica, QZ = quartz, GI = gibbsite, and FD = feldspar.

‡ CS = composite mineralogical control section (Soil Survey Staff, 1975).

all profiles. In effect, these minerals apparently decrease with greater weathering intensity. The reverse is true for hydroxy interlayered vermiculite, which increases with age from T0 to T2 and decreases with depth in all profiles. These trends are consistent with the stability indices of Jackson (1968) and suggest a transformation from vermiculite or mica to hydroxy interlayered vermiculite. Such a transformation has

Table 5—Mineralogy of the clay (<2 μ m) fraction of samples examined.†

Soil	Horizon‡	Depth, cm	KK	HIV	VR	M	QZ	SM	GI	CL	FD		
												%	
T0	Ap	0-23	35	25	20	10	7	2	1	-	-		
	CS	25-100	24	26	20	25	1	3	1	-	-		
	B21	58-103	19	20	40	10	8	2	1	-	-		
	C	200-250	18	6	39	25	5	6	1	-	-		
T1	Ap	0-35	15	35	10	10	13	6	1	10	-		
	CS	35-85	25	38	15	20	Tr	3	1	-	-		
	B23t	65-80	22	25	15	15	27	2	1	5	-		
	IIC	138-185	26	14	25	28	1	5	1	-	-		
T2	Ap	0-13	10	53	0	4	27	3	1	-	2		
	CS	13-63	25	45	10	5	15	-	-	-	-		
	IIB22t	63-88	25	45	10	4	16	-	-	-	-		
	IIB23t	88-120	30	40	15	10	5	-	-	-	-		
T3	Ap	0-15	28	45	0	5	22	-	1	-	-		
	CS	15-65	40	45	0	10	5	-	-	-	-		
	B24t	75-95	56	25	5	5	7	2	-	-	-		
	IIIB27t	143-158	50	20	10	15	5	-	-	-	-		

† KK = kaolinite, HIV = hydroxy interlayered vermiculite, VR = vermiculite, M = hydrous mica, QZ = quartz, SM = smectite, GI = gibbsite, CL = chlorite, and FD = feldspar.

‡ CS = composite mineralogical control section (Soil Survey Staff, 1975).

been noted by others (Rich and Obenshain, 1955; Sawhney, 1960; Johnson et al., 1963).

An increase in kaolinite with depth for all profiles but T0 raises questions as to its stability relative to hydroxy-interlayered vermiculite, which concentrates in the surface. Similar depth trends for these minerals have been noted and discussed by others (Rich and Obenshain, 1955; Bryant and Dixon, 1963; Douglas, 1965; Dixon, 1966; Carlisle and Zelazny, 1973; Mokma et al., 1973a; Douglas, 1977; Dixon, 1977). At least three explanations have been proposed for these trends: (i) hydroxy-interlayered vermiculite is more stable in an acid environment (Carlisle and Zelazny, 1973; Rich and Obenshain, 1955); (ii) kaolinite forms more readily in the subsoil (Rich and Obenshain, 1955); and (iii) source material is less abundant for kaolinite than for hydroxy-interlayered vermiculite (Rich and Obenshain, 1955).

The third explanation partially accounts for the data of the present study if rate of transformation is considered as a factor. Hydroxy interlayered vermiculite could be forming relatively fast under current climatic conditions, given the abundance of mica and vermiculite in the younger alluvium. Kaolinite formation from 2:1 layer silicates could be very slow. If such were the case, differential weathering intensities would result in a faster interlayering of mica and vermiculite near the surface and a subsequent accumulation of hydroxy interlayered vermiculite in the Ap horizon. Kaolinite would not be significantly replenished through transformation or neof ormation and would be depleted from the surface through eluviation. This would explain why hydroxy-interlayered vermiculite shows substantial increases from T0 to T2 relative to kaolinite. The increase in kaolinite from T2 to T3 could be related to the extreme age of the latter soil or to past climatic conditions during which weathering was more intense than at present. Prolonged or intense desilication and alumination could have led to a transformation of 2:1 intergrades to 1:1 layer silicates (Jackson, 1965). This would explain the decline in hydroxy interlayered vermiculite from T2 to T3, but not the high amount of that mineral in the surface of T3. Similar abrupt increases in kaolinite

with age have been reported by Brewer and Walker (1969) and Mokma et al. (1973a).

Smectite and gibbsite, which are present in trace amounts, did not show a trend with age or depth. The near absence of gibbsite (even in the oldest soil) could be due to the presence of hydroxy-interlayered vermiculite resulting in the "anti-gibbsite effect" (Jackson, 1962).

Depositional variation in particle size could have affected the proportion of inherited secondary clay minerals both within and among profiles. However, the presence of gravel and cobble strata in T2 and T3 suggests that these soils did not develop from fine-textured slack water deposits, and that an appreciable amount of the clay present formed from the weathered products of primary minerals. Furthermore, illuviation in these older soils should have minimized the effects of within-profile depositional variation in the clay fraction.

Clay mineralogical trends do not show as abrupt a break between the younger and older soils as was evident for the sand fractions and to a lesser extent for the silt fractions. Whereas the sand fractions of T2 and T3 have virtually identical mineralogy (primarily quartz), the clay fraction continuously changes from youngest to oldest soil. In effect, mineralogical stability is more rapidly attained in the coarser fractions, and the finer fractions continue to undergo chemical alterations, transformations, and neof ormations over a prolonged time span.

Whole Soil Mineralogy

The mineralogical composition of the whole soil, derived from particle size data (Table 1) and estimated mineral percentages (Tables 3-5), shows a concentration of quartz in the surface with age (Table 6). This trend can be attributed to the stability of quartz, to its predominance in the coarser fractions which do not eluviate, and to the illuviation of layer silicates, some of which could be comparably stable.

The near depletion of feldspar in T2 and T3 suggests that chemical weathering has destroyed or transformed this mineral over time before it could under-

Table 6—Whole soil mineralogy of samples examined.†

Soil	Horizon‡	Depth, cm	QZ	FD	M	HM	KK	HIV	VR	%						Total
										SM	CL	GI	Fe ₂ O ₃	Other		
T0	Ap	0-23	50	6	13	5	9	4	3	-	-	1	6	3	100	
	CS	25-100	50	8	17	4	6	7	4	1	-	-	3	3	103	
	B21	58-103	54	7	16	5	4	5	7	-	-	-	3	2	103	
	C	200-250	49	9	15	5	4	4	7	1	-	-	3	-	97	
T1	Ap	0-35	63	9	8	3	3	6	2	1	2	-	4	1	102	
	CS	35-85	48	6	15	2	9	12	3	1	-	-	4	2	102	
	B23t	65-80	53	7	16	-	6	8	3	-	1	-	4	2	100	
	IIC	138-185	49	12	12	-	5	6	4	1	-	-	6	1	96	
T2	Ap	0-13	86	1	1	-	1	6	-	-	-	-	3	1	99	
	CS	13-63	61	-	3	-	9	16	4	-	-	-	7	1	101	
	IIB22t	63-88	59	1	2	-	10	17	4	-	-	-	8	-	101	
	IIB23t	88-120	44	-	5	-	16	20	7	-	-	-	7	1	100	
T3	Ap	0-15	85	-	1	-	4	7	-	-	-	-	2	1	100	
	CS	15-65	53	-	5	-	16	18	-	-	-	-	7	1	100	
	B24t	75-95	43	-	3	-	26	12	2	1	-	-	11	-	98	
	IIB27t	143-158	43	-	9	1	23	9	5	-	-	-	9	-	99	

† QZ = quartz; FD = feldspar; M = mica; HM = translucent heavy minerals; KK = kaolinite; HIV = hydroxy interlayered vermiculite; VR = vermiculite; SM = smectite; CL = chlorite; GI = gibbsite; and other = rock fragments, opaque minerals, and unidentified minerals.
‡ CS = composite mineralogical control section (Soil Survey Staff, 1975).

go physical weathering to clay-size particles. Mica, though also diminished in the Ap horizon of T2 and T3, still persists in moderate quantities in lower horizons of these soils. The persistence of mica relative to feldspar suggests greater stability of the former, with its depletion in the sand fraction being perhaps a result of comminution. The depletion of mica and vermiculite in Ap horizons of T1, T2, and T3 is attributable to illuviation and a higher rate of transformation to more stable clay minerals near the surface. Illuviation is also evident for kaolinite and hydroxy-interlayered vermiculite. The increase in these minerals from T1 to T2 suggests that they are forming via transformation of feldspar and mica, respectively. However, the increase in kaolinite from T2 to T3 cannot totally be attributed to feldspar weathering since feldspar is nearly depleted in T2. A decrease in hydroxy-interlayered vermiculite concomitant with the increase in kaolinite could be reflecting a hydroxy-interlayered vermiculite to kaolinite transformation, as was suggested above.

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